ARTICLE IN PRESS
MARGO-05052; No of Pages 34
Marine Geology xxx (2014) xxx–xxx
Contents lists available at ScienceDirect
Marine Geology
journal homepage: www.elsevier.com/locate/margeo

The Messinian Salinity Crisis: Past and future of a great challenge for marine sciences

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ARTICLE INFO

Article history:
Received 6 June 2013
Received in revised form 14 February 2014
Accepted 14 February 2014
Available online xxxx

Keywords:
Messinian Salinity Crisis
Marine geology
Mediterranean
Evaporites
Palaeoceanography

ABSTRACT

Forty years after the image of the Mediterranean transformed into a giant salty lake was first conceived, the fascinating history of the Messinian Salinity Crisis (MSC) still arouses great interest across a large and diverse scientific community. Early outcrop studies which identified severe palaeoenvironmental changes affecting the circum-Mediterranean at the end of the Miocene, were followed by investigations of the marine geology during the 1950s to 1970s. These were fundamental to understanding the true scale and importance of the Messinian event. Now, after a long period of debate over several entrenched but largely untested hypotheses, a unifying stratigraphic framework of MSC events has been constructed. This scenario is derived mainly from onshore data and observations, but incorporates different perspectives for the offshore and provides hypotheses that can be tested by drilling the deep Mediterranean basins.

The MSC was an ecological crisis, induced by a powerful combination of geodynamic and climatic drivers, which had a great impact on the subsequent geological history of the Mediterranean area, and on the salinity of the global oceans. These changed the Mediterranean's connections with both the Atlantic Ocean and the freshwater Paratethyan basins, causing high-amplitude fluctuations in the hydrology of the Mediterranean. The MSC developed in three main stages, each of them characterized by different palaeoenvironmental conditions. During the first stage, evaporites precipitated in shallow sub-basins; the MSC peaked in the second stage, when evaporite precipitation shifted to the deepest depocentres; and the third stage was characterized by large-scale environmental fluctuations in a Mediterranean transformed into a brackish water lake.

The very high-resolution timescale available for some Late Miocene intervals in the Mediterranean makes it possible to consider environmental variability on extremely short time scales including, in some places, annual changes. Despite this, fundamental questions remain, some of which could be answered through new cores from the deepest Mediterranean basins. Improvements in seismic imaging and drilling techniques over the last few decades make it possible to plan to core the entire basinal Messinian succession for the first time. The resulting data would allow us to decipher the causes of this extreme environmental change and its global-scale consequences.

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http://dx.doi.org/10.1016/j.margeo.2014.02.002

Please cite this article as: Roveri, M., et al., The Messinian Salinity Crisis: Past and future of a great challenge for marine sciences, Marine Geology (2014), http://dx.doi.org/10.1016/j.margeo.2014.02.002

1. Introduction. The Messinian Salinity Crisis: an enduring challenge for land-based and offshore researchers

Unraveling the history of the environmental modifications experienced by the Mediterranean region at the end of the Miocene is an intellectual challenge that has fascinated a large community of earth and life scientists for almost 50 years. As with the other saline giants that developed episodically during the Earth's history (Hsü, 1972; Warren, 2010), the causes of, and mechanisms by which more than a million cubic kilometers of salt accumulated on the Mediterranean sea-floor over a brief period of less than 700 ka are difficult to understand. In part this is because of the absence of modern evaporitic systems comparable in terms of both size and mineralogy. However, what is commonly known as the Messinian Salinity Crisis (MSC), i.e. the transformation of a small ocean into a giant evaporitic pool and then into a brackish-water lake, is unique amongst the saline giants, because it occurred relatively recently, in a basin that has not subsequently undergone significant modifications. Consequently, the majority of the MSC's sedimentary record is still preserved below the present-day Mediterranean seafloor (Fig. 1). Marine scientists, and amongst them marine geologists, therefore have a key role in exploring and elucidating this extraordinary event and extracting its general implications for the formation of saline giants.

The concept of a Messinian salinity crisis (Selli, 1954) was first formulated from onshore studies, which demonstrated the widespread, coeval development of hyper- and hyposaline environments all around the Mediterranean at the end of the Miocene (Ogniben, 1957; Selli, 1960; Ruggieri, 1967). However, the true scale and importance of the Messinian environmental changes in the Mediterranean was fully realized only after pioneering marine geology studies carried out during the late 50s to early 70s (Bourcart et al., 1958; Bourcart, 1959a,b; Alinat and Cousteau, 1962; Bourcart, 1962; Hersey, 1965; Alinat et al., 1966; Glangeaud et al., 1966; Cornet, 1968; Mauffret, 1969, 1970; Montadert et al., 1970; Auzende et al., 1971; Ryan et al., 1971; Biscaye et al., 1972; Mauffret et al., 1972; Bellaiche and Recq, 1973; Bellaiche et al., 1974). These studies identified in the western Mediterranean diapiric structures rooted in an offshore salt layer up to 2 km thick, an erosional surface on basin margins, and a deep basin trilogy of seismic units (Fig. 1) topped by a series of strong reflectors (the "M" reflectors) that were soon correlated with the onshore evaporitic units. The 1970 Deep Sea Drilling Project (DSDP) Leg 13 (Hsü, 1972, 1973; Hsü et al., 1973a,b), drilled in the Mediterranean, recovered cores from the top of the evaporite unit for the first time in the deep basins (Fig. 1). This confirmed the early hypothesis about the age and nature of the MSC and generated an explosion of interest and publicity. Partly as a result of this, the MSC burst on to the pages of *Marine Geology*, becoming one of the most common topics of papers published by the journal, with 153 articles citing the phrase *Messinian Salinity Crisis* from 1972 to present.

The main scientific challenge following that initial drilling has been combining the onshore and offshore records of the MSC into a single, unified scenario. The different approaches, instruments, scale of observations and degree of resolution that characterize the two domains and their rather separate scientific communities, account, in part, for why this has proved so difficult to achieve. Perhaps more important however, is the absence of what could be a common basis for a synthesis, i.e. a complete cored succession from the deep basinal Mediterranean. As a result, the MSC represents an enduring collaborative opportunity for land and marine-based scientists to contribute to, not only a full understanding of what happened in the Mediterranean 6 million years ago, but also to unraveling the complex mechanisms involved in Earth's responses to environmental changes at different temporal and spatial scales. In this respect, the Messinian evaporite record offers a great opportunity for studying the limits of life in the extreme environments of our planet, with important implications for planetary sciences (see Section 5.3). Moreover, Messinian events and deposits are undergoing a renewed interest for the assessment of the petroleum system's potential of the pre-Messinian subsalt successions in several Mediterranean provinces (Pawlewicz, 2004; Belopolsky et al., 2012).

2. Discoveries and controversies

Following the first Messinian colloquium, C.W. Drooger compiled a collection of papers for a book, entitled *Messinian Events in the*

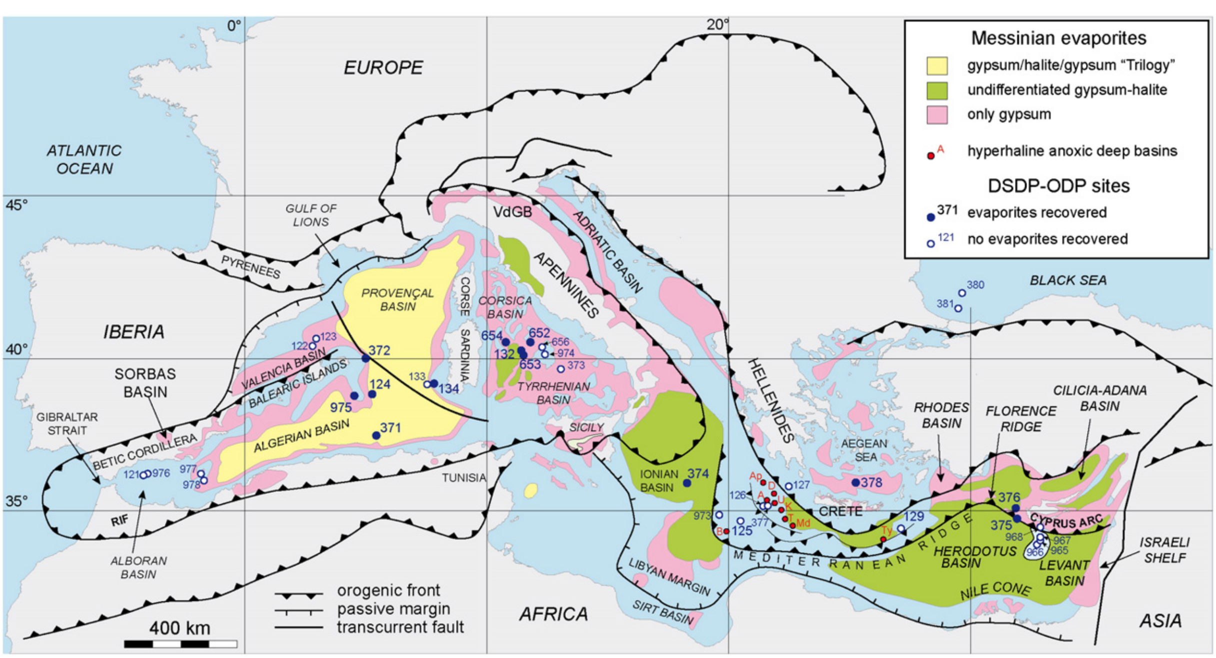

Fig. 1. Distribution of Messinian evaporites and location of the DSDP-ODP boreholes which recovered Messinian deposits. The location of the main hyperhaline anoxic deep basins on top of the Mediterranean Ridge is also shown: Ap, Aphrodite; A, Atlante; B, Bannock; D, Discovery; K, Kryos; M, Medee; T, Thetis; Ty, Tyro; U, Urania.
Modified from Rouchy and Caruso (2006) and Manzi et al. (2012).

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Mediterranean (1973). In his review of this volume — published by *Marine Geology*, D. J. Stanley wrote:

> "*It is a safe prediction that we are going to hear more about these phenomenal Messinian events and consequences, and oceanologists will find the few hours spent reading this book a profitable experience*".
> [Stanley, 1975]

Stanley provided an exhaustive technical review of the scientific debate surrounding the Mediterranean Messinian, some features of which continue to be debated today. He was also successful in describing (with a hint of irony) the somewhat overheated climate of such a debate and the potential implications of Messinian events at a global scale. Since then, the history of Messinian research has been characterized by endless controversies among different research groups concerning both fundamental and minor MSC issues. Different approaches, analytical tools and study areas, mainly onshore versus offshore, but also onshore articulation at regional scale, created multiple, often strongly opposing perspectives of the MSC.

2.1. Deep versus shallow

The first controversy concerned the deep versus shallow nature of the late Miocene Mediterranean basin. The hypothesis of a shallow Mediterranean throughout the Messinian (200–500 m water depth; Nesteroff, 1973) was disproved by the recovery of deep-water sediments below, within and above the basinal evaporites (Hsü et al., 1973a,b; Ryan, 1976). Debate then focussed on the origin of evaporites themselves; were they of deep or shallow water origin (Debenedetti, 1976, 1982; Dietz and Woodhouse, 1988)? The deep-water hypothesis for evaporite precipitation based on the theoretical model of Schmalz (1969) was ruled out using several lines of evidence: the cored evaporites had characteristics resembling those forming in modern sabkhas and contained features interpreted as desiccation cracks, indicative of subaerial exposure; bull's eye patterns of evaporite distribution were observed in some locations; and the apparent absence of turbidites in the deepest areas (Hsü et al., 1973a,b). However, the clinching argument supporting the deep-basin shallow-water hypothesis was the widespread evidence of Late Miocene rejuvenation of the rivers draining into the Mediterranean (Chumakov, 1967, 1973; Ryan, 1978; Barber, 1981; Clauzon, 1982), causing large erosion with entrenched subaerial canyon incision onshore (Denizot, 1952) and on present-day shelf and extending as submarine canyons on the Mediterranean shelf and slopes. It was these Messinian erosional surfaces, rather than the evaporitic deposits themselves, that turned out to be the "*Rosetta Stone*" of the Messinian Salinity Crisis (Ryan, 2009).

2.2. The deep desiccated basin model (1973): birth of a paradigm

Despite its shocking and improbable nature, the deep-basin shallow-water scenario implying the desiccation of the Mediterranean (Hsü et al., 1973a,b; Hsü, 1984) and its transformation into a giant salt desert more than 1000 m below global sea-level, turned out to be the only model able to provide a *possible* explanation for all the available data and became the paradigm of the Messinian Salinity Crisis (Roveri and Manzi, 2006; Roveri et al., in press). Over the following 40 years, although many important aspects of the MSC were vigorously debated (e.g. timing of and trigger for the MSC, onshore–offshore equivalence of evaporite units, Mediterranean connections with the Atlantic, Red Sea and Paratethys, processes and rates of evaporite accumulation), the paradigm itself was rarely questioned, and importantly, no alternative models were proposed (see for example Busson, 1990). The main criticism of the deep desiccated basin model focussed on doubts about the interpretation of the basinal evaporites as shallow water precipitates (Hardie and Lowenstein, 2004), and on the implications of the widely distributed clastic evaporites emplaced by deep-water gravity flows, which had been overlooked (Martinez del Olmo, 1996; Roveri et al., 2001; Manzi et al., 2005; Roveri and Manzi, 2006). These studies suggest that deep-water conditions may have persisted throughout the MSC.

2.3. Timing of evaporite deposition: synchronous vs. diachronous

The hypothesis that Messinian evaporite precipitation was a diachronous process was first introduced by Rouchy (1982) and further elaborated by Rouchy and Saint Martin (1992). In this scenario, the onset of evaporite deposition started in the shallowest basins and migrated progressively into the deeper basins with increasing brine concentration and sea level fall (Rouchy and Caruso, 2006; Fig. 2e). A diachronous scenario was also proposed by Butler et al. (1995); Fig. 2a) who showed that the Sicilian sub-basins (Fig. 1) formed syntectonically in the moving synclines of a fold-and-thrust belt during an overall phase of gradual uplift, and proposed that evaporite deposition in each sub-basin was triggered diachronously. Based on magnetostratigraphic data they claimed that the earliest evaporite deposits formed at ~6.88 Ma while the youngest deposits precipitated more than 800 ka later. Riding et al. (1998; Fig. 2c), based on the sedimentary record of the Betic Cordillera (Spain; Fig. 1), suggested that evaporite precipitation started in early Messinian in onshore basins, then moved to the Mediterranean deep basin in the mid-Messinian when it was desiccated, and ended in onshore basins in the late Messinian during its reflooding.

At the time of these publications, these diachronous scenarios were difficult to test because of the absence of reliable age control. Based on magnetostratigraphic data, Gautier et al. (1994) suggested a synchronous development of the MSC in Sicily and Andalusia and its limitation to chron C3r (Gilbert epoch), but accurate timing of main events was not possible at that time (Fig. 2b). In the late 1990s, however, the development of an astronomical time scale for the lower Messinian deposits (Hilgen et al., 1995) provided a mechanism for evaluating the diachronous hypothesis. On Sicily, the Messinian evaporites are directly underlain by the cyclically-bedded diatomites of the Tripoli Formation that are ideal for astronomical dating. Consequently, the first detailed cyclostratigraphic studies showed that the oldest Sicilian evaporites in these subbasins, all formed synchronously at an age of 5.98 Ma (Hilgen and Krijgsman, 1999); then astronomical dating also showed that evaporite deposition in Spain, Greece and Cyprus took place at approximately the same age of 5.96 ± 0.02 Ma (Krijgsman et al., 1999a, 2002) demonstrating that, at the resolution of a precessional cycle, the onset of evaporite deposition was synchronous throughout both east and west Mediterranean basins (Fig. 2d).

In broad terms then, the coeval nature of the beginning of the MSC has been settled. However, in detail, questions are still debated because all the astronomically calibrated gypsum deposits are derived from marginal settings with relatively shallow water depths (~200 m). The deepest water section (~1000 m; Falconara on Sicily) exposed on land shows a transition to evaporitic dolostones at the age of 5.98 Ma (Hilgen and Krijgsman, 1999) rather than primary gypsum. This leaves the nature and timing of the onset of the MSC in deep basinal settings still open to debate.

3. State-of-the art, 40 years after the desiccation model

3.1. Marginal vs. deep, onshore vs offshore: fundamental definitions

The *onshore* Messinian successions outcropping or buried in the peri-Mediterranean areas have provided the majority of the sedimentological, stratigraphical, paleontological and geochemical data for the reconstruction of a Messinian stratigraphic model (Fig. 3), providing useful elements for attempting correlations with offshore successions. Onshore Messinian successions are usually described as *marginal* or *peripheral* basins. The term *marginal* refers to their position with respect to the deep, central Messinian basins that largely correspond to the present-day Mediterranean slopes and bathyal plains (*offshore* basins).

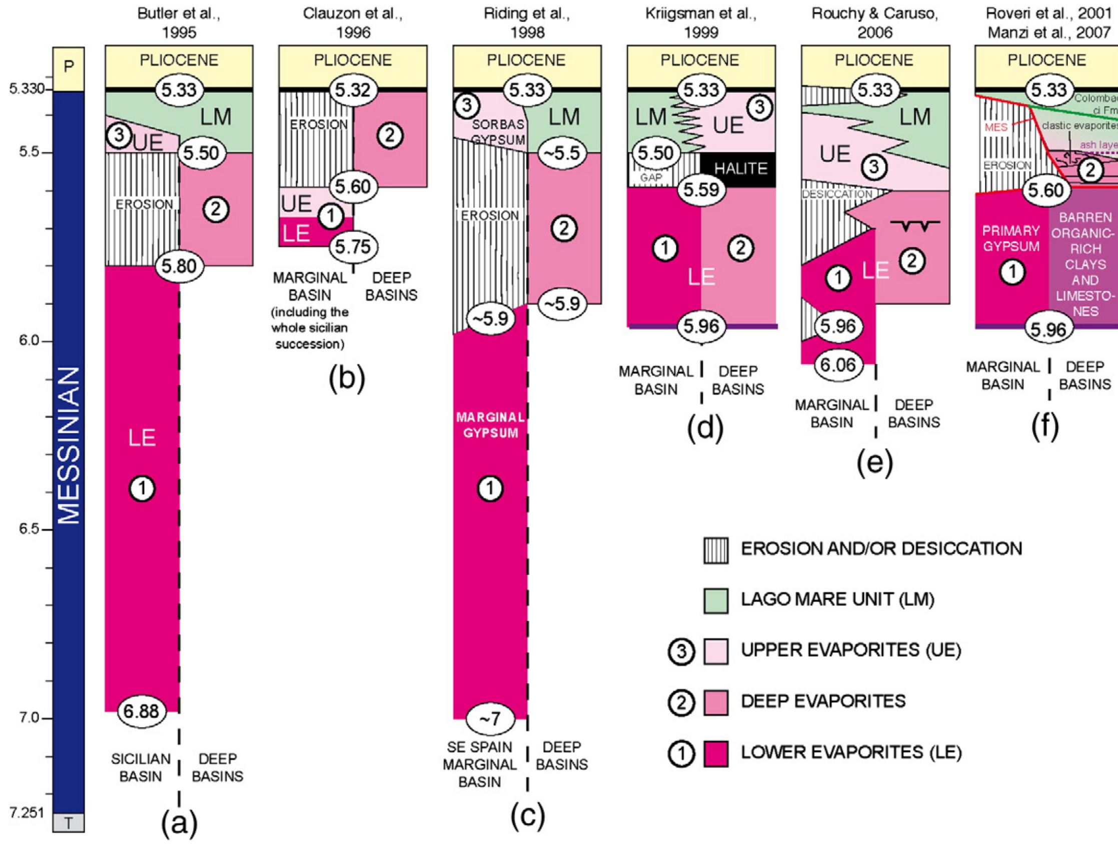

Fig. 2. Diachronous (a, b, c) and slightly diachronous (e) vs. synchronous (d, f) stratigraphic models of the Messinian Salinity Crisis. Modified from Rouchy and Caruso (2006) and Manzi et al. (2007).

This resulted in the equation: marginal/peripheral basin = onshore = shallow-water; deep basin = offshore = deep-water; which is not fully correct and caused some confusion in the literature. This is not a trivial problem of nomenclature since one of the main problems hampering the full comprehension of Messinian events is the uncertainty inherent in correlating onshore exposure with deep offshore records (see Section 4.1). For this reason we want to clarify this important point here.

For a long time Sicily was considered a perfect equivalent of the offshore succession because the clear tripartite character of its Messinian sequence recalls the seismic "trilogy" of offshore basins (Decima and Wezel, 1971). However, some authors suggested that Sicily, though deeper than many of the marginal basins exposed around the Mediterranean margins, should actually be considered as a "shallow" setting (Clauzon et al., 1996). Recent studies have partially confirmed this latter view, reconstructing, however, a more complex framework of the Sicilian stratigraphy (Roveri et al., 2008a). Like those of the Apennines (Roveri et al., 2001), the Sicilian basins formed on a growing orogenic wedge; this structural setting resulted in an articulated palaeotopography, with both shallow- and deep-water depocenters, the latter attaining palaeodepths of up to 1000 m. As a consequence, onshore *marginal/peripheral* basins actually also include *intermediate* basins with a palaeobathymetric range between the shallow-water (<200 m) and deep-water settings (>1000 m; Fig. 4). These intermediate basins may be the key to successful correlations to deep offshore basins.

3.2. MSC stratigraphic framework: the onshore perspective

3.2.1. Messinian chronostratigraphy

Recent advances in the development of the astronomical time scale have also resulted in establishing the age of the Messinian Stage (Fig. 5). The Global Stratotype Section and Point (GSSP) for the base of the Messinian has been defined at a level closely associated with the first occurrence of the planktic foraminifer *Globorotalia miotumida* in the Oued Akrech section (Atlantic Morocco) at 7.25 Ma (Hilgen et al., 2000). The top of the Messinian is defined by the Zanclean GSSP at Eraclea Minoa (Sicily) at the level that corresponds to the base of the Trubi marls, closely coinciding with the Pliocene reflooding of the Mediterranean, with an age of 5.33 Ma (Van Couvering et al., 2000). The Messinian Stage thus has a total duration of 1.92 Ma. Cyclostratigraphic correlations between Mediterranean pre-evaporite successions are rather straightforward and astronomical tuning to successive insolation peaks generally shows a good to excellent fit between the characteristic sedimentary cycle patterns and the astronomical target curve, including precession/obliquity interference patterns in insolation (Hilgen and Krijgsman, 1999; Sierro et al., 2001; Hilgen et al., 2007). Integrated high-resolution cyclo-, bio- and magnetostratigraphic studies of the pre-evaporite successions showed that the transition to evaporitic conditions occurred at 5.96 ± 0.02 Ma, about 4 cycles above the C3An.2n(y) magnetic reversal and was synchronous across the western and eastern Mediterranean (Krijgsman et al., 1999a, 2002). Recently, a re-evaluation of the cyclostratigraphic pattern of the evaporites of the Sorbas basin refined the tuning of the gypsum cycles and defined the onset of the MSC to 5.97 Ma (Manzi et al., 2013).

The pre-evaporitic marl–sapropel cycles are followed by the marl–gypsum cycles of the Primary Lower Gypsum (see description in Section 3.2.3.2). This suggests that the evaporite cycles are related to precession-controlled oscillations in (circum-) Mediterranean climate as well, where the gypsum beds correspond to precession maxima (insolation minima) and relatively dry climatic episodes (Krijgsman et al., 2001). Cyclostratigraphic correlations provide similar numbers of sedimentary cycles: ~16 cycles in the Sorbas basin of SE Spain and in the Vena del Gesso basin of NE Italy (Vai, 1988, 1997; Krijgsman et al., 2001; Lugli et al., 2010). The most logical tuning of the cyclostratigraphic

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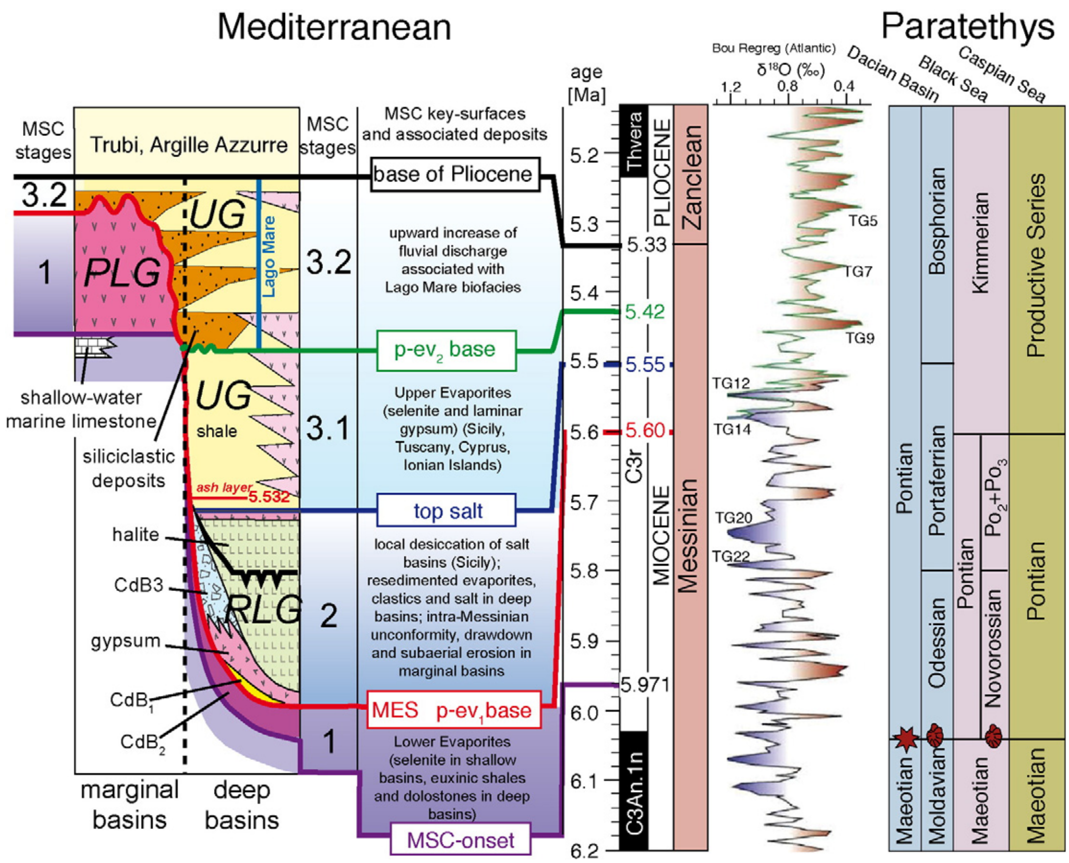

Fig. 3. Chronostratigraphy of Late Miocene to Early Pliocene with MSC events in the Mediterranean (modified from CIESM, 2008; Manzi et al., 2013) and Paratethyan basins (from Krijgsman et al., 2010) and correlations to the oxygen isotope curves of the Atlantic margin of Morocco (Hilgen et al., 2007). Stage 2 of the Mediterranean scheme corresponds to stage 2.1 of CIESM (2008); stages 3.1 and 3.2 were grouped into stage 2.2 in the CIESM (2008) model. PLG, Primary Lower Gypsum; RLG, Resedimented Lower Gypsum; UG, Upper Gypsum; CdB, Calcare di Base. Red star/foraminifer indicates influx of marine nannofossils (Marunteanu and Papaianopol, 1998) and/or foraminifera.

patterns to the astronomical curve results in a total duration of ~380 ka and an age of ~5.59 Ma for the top of the Primary Lower Gypsum, although this calibration is not as straightforward as for the pre-evaporites due to lack of bio- and magnetostratigraphic constraints (Vai, 1997; Krijgsman et al., 1999b; Hilgen et al., 2007; Manzi et al., 2013; Fig. 6).

The final stage of the MSC is marked by deposition of the Upper Gypsum and Lago-Mare units, overlying erosional surfaces and covered by marine Pliocene. These units also display a marked cyclicity, comprising seven to ten sedimentary cycles in the Upper Gypsum of Sicily (Decima and Wezel, 1971; Van der Laan et al., 2006; Hilgen et al., 2007; Manzi et al., 2009), the post-evaporitic deposits of Northern Italy (Vai, 1997;

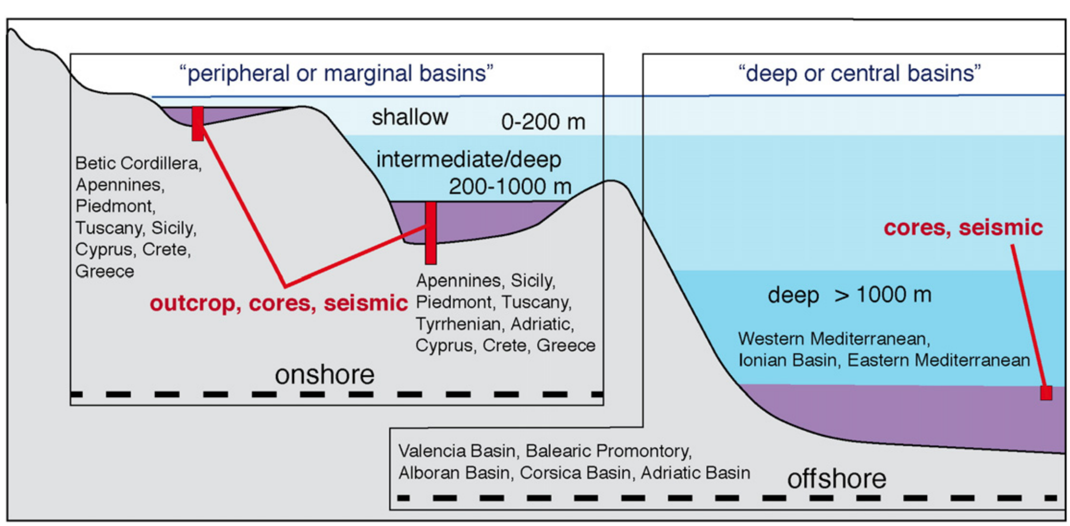

Fig. 4. Schematic classification of the Messinian sub-basins in the Mediterranean. Marginal/peripheral basins include both shallow-water (0–200 m) and relatively deep-water settings (300–1000 m) and are mainly represented onshore. Deep sub-basins (>1000 m) are only in the offshore domain, where also intermediate depths sub-basins (300–1000 m) occur (compare with Fig. 12). Note that all the sub-basins are physically disconnected.

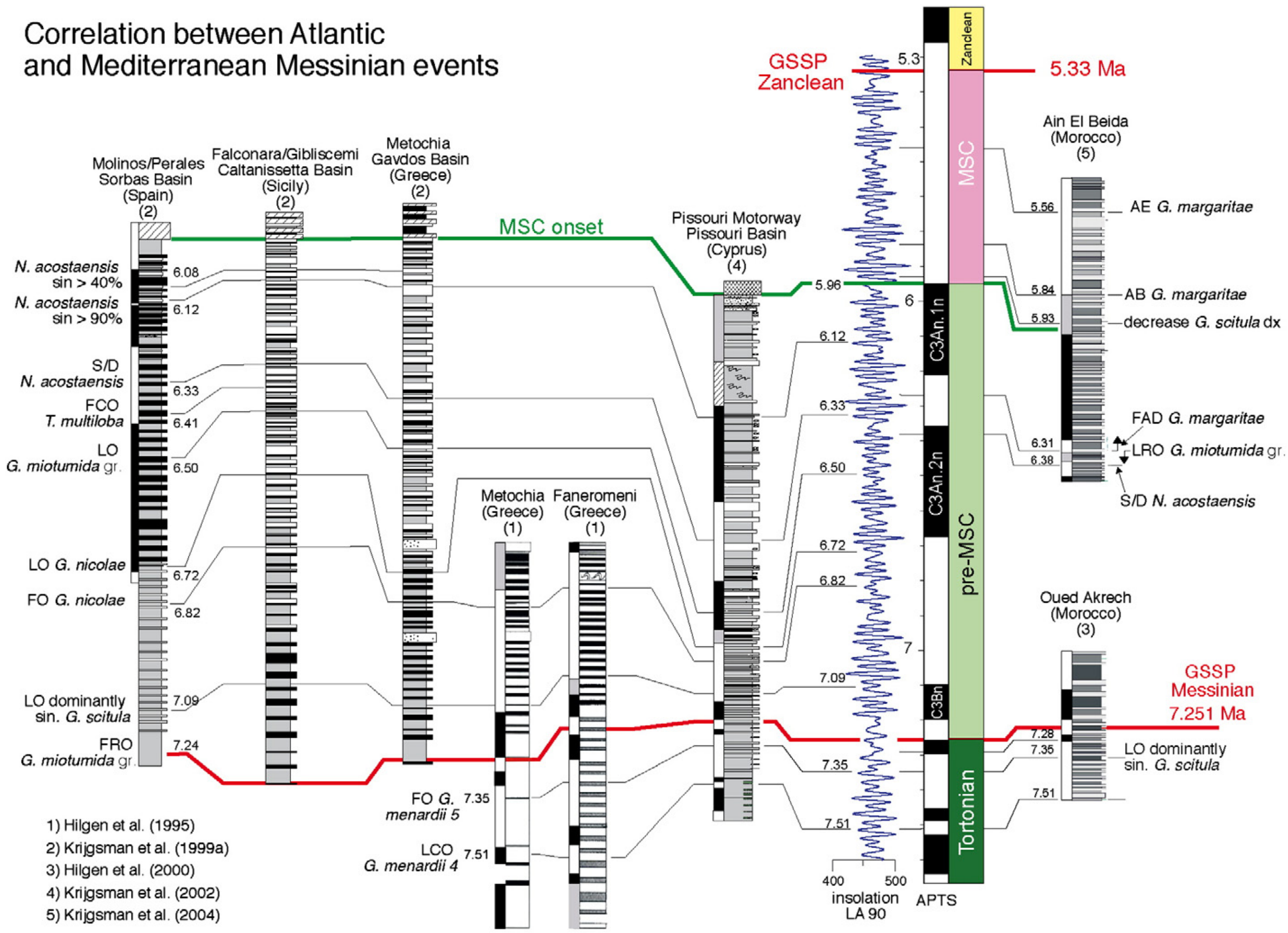

Fig. 5. Messinian pre-MSC chronostratigraphy. Astronomical tuning of Messinian key sections located in the Mediterranean or in the adjacent Atlantic Ocean. Modified from Krijgsman et al. (2004).

Roveri et al., 2009) and the Zorreras/Feos units of southeast Spain (Fortuin and Krijgsman, 2003; Bassetti et al., 2006; Omodeo-Sale' et al., 2012). If these units are coeval and the cycles are also insolation driven, this suggests that the interval comprises at least ~180 ka. Downward tuning from the base of the Pliocene results in an age for the post-evaporitic base of 5.53 Ma and a "Messinian Gap" of approximately 60 ka, during which Messinian erosion and/or deposition of resedimented gypsum and halite took place (Roveri et al., 2008b; Manzi et al., 2009). Recent U–Pb zircon dates of ~5.5320 ± 0.0046 (or 0.0074) Ma for an ash layer in the basal part of the post-evaporitic unit in the Apennines are in good agreement with the cyclostratigraphic estimates, but point to a slightly shorter duration of the hiatus (Cosentino et al., 2013).

3.2.2. Messinian oxygen isotope framework: the role of glacio-eustacy in the MSC evolution

Oxygen isotope records are critical for deciphering the role played by glacio-eustacy in the evolution of the MSC. Benthic oxygen isotope records for the Messinian come from outside the Mediterranean, because the paleoenvironmental conditions during the MSC were too extreme to support normal faunal communities. Messinian obliquity-controlled glacials have been estimated to impact sea level on the order of ~50–60 m for peak glacials (Kastens, 1992; Shackleton et al., 1995). Distinctly positive isotope peaks are numbered from youngest to oldest following the nomenclature of Shackleton et al. (1995) and their tuning to the astronomical curves is well established (Hodell et al., 2001; Vidal et al., 2002; Van der Laan et al., 2005; Fig. 5). The oxygen isotope records clearly show obliquity controlled glacial cycles in the interval ranging from ~6.3 to 5.5 Ma, and the onset of the MSC closely coincides with glacial stage TG32 at 5.97 Ma (Manzi et al., 2013).

The most prominent glacial peaks are TG20–22 (at 5.75 and 5.79 Ma) and TG12–14 (at 5.548 and 5.582 Ma; Fig. 6). These last two glacials may correspond to the hiatus between the Primary Lower Gypsum and Upper Gypsum units and might have been the trigger for halite and potash salts deposition at the peak of the MSC (Hilgen et al., 2007; CIESM, 2008; Roveri et al., 2008a; Speranza et al., 2013). The deposits of the final stage of the MSC (Upper Gypsum, post-evaporitic, Lago-Mare) are coeval with stepwise deglaciation from TG12 to TG9, which is associated with a distinct glacio-eustatic sealevel rise. The end of the MSC appears not to coincide with any major deglaciation, which gives credibility to alternative causes for the Pliocene flooding (Van der Laan et al., 2006).

3.2.3. A 3-stage stratigraphic model for the MSC

The long-lived controversy on the diachronous versus synchronous onset of Messinian evaporite deposition has recently been addressed with a consensus stratigraphic model arose from a community workshop organized in 2007 by CIESM (Commission Internationale pour l'Exploration Scientifique de la mer Méditerranée). This stratigraphic model (CIESM, 2008, Fig. 3), which was inspired by the two-step model of Clauzon et al. (1996; Fig. 2b), was constructed using field and subsurface data from onshore basins (mainly Sicily and Northern Apennines) and by integrating bio- and magnetostratigraphic data with facies analysis and physical stratigraphy (Roveri et al., 2001;

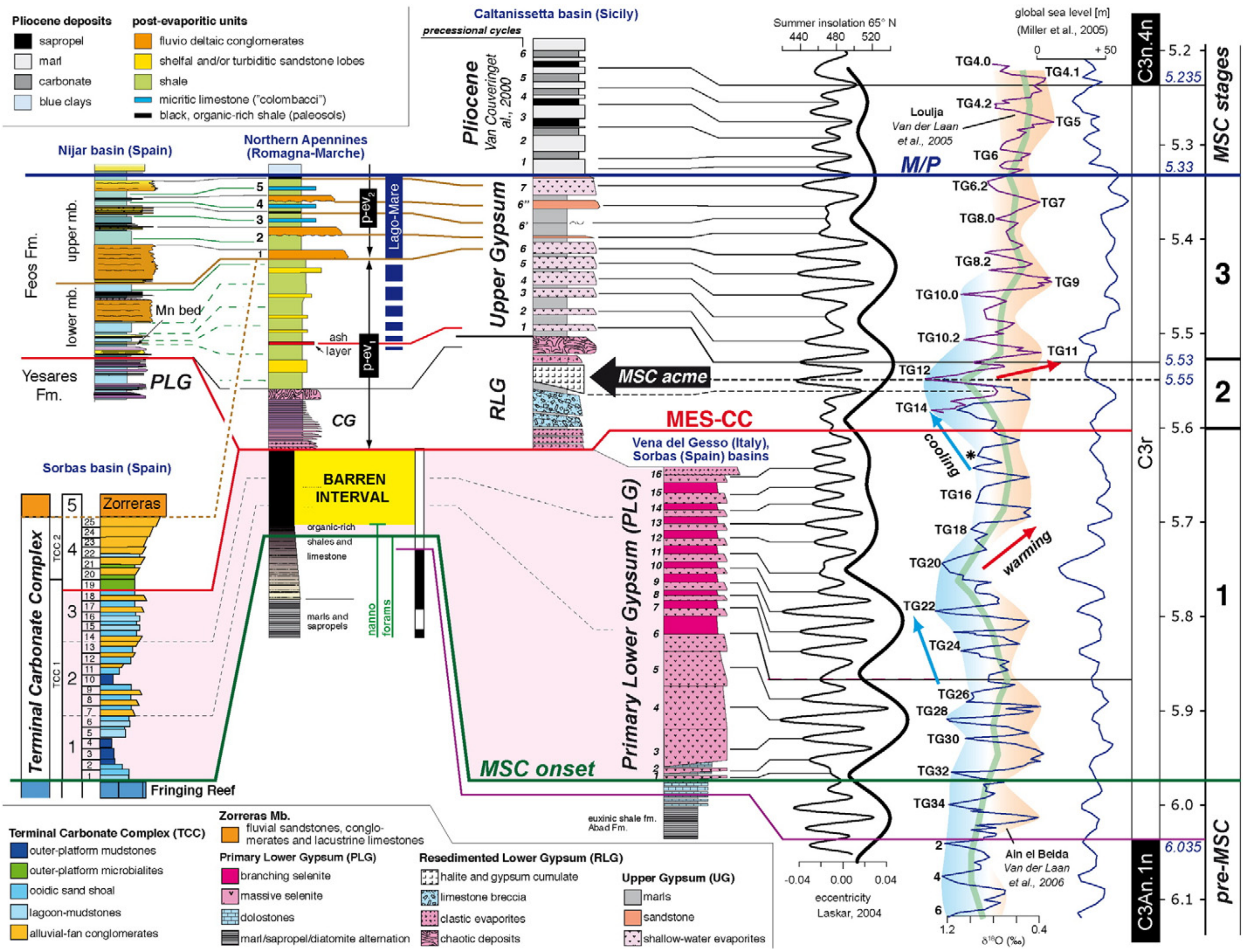

Fig. 6. Messinian MSC chronostratigraphy and astronomical tuning of the Mediterranean Messinian key sections. The Primary Lower Gypsum unit (PLG, Lugli et al., 2010) was deposited during stage 1 of MSC. Resedimented Lower Gypsum unit (RLG) is a complex unit including clastic gypsum, halite, gypsum primary cumulate and limestone breccia (Calcare di Base type 3, sensu Manzi et al., 2011). During stage 3 the Upper Gypsum (UG, see Manzi et al., 2009) was deposited in Sicily and Eastern Mediterranean, while the Northern Apennines and the Betics are characterized by deposition of terrigenous units (modified from Roveri et al., 1998; Omodeo-Sale' et al., 2012). The carbonate unit of the Terminal Carbonate Complex (TCC; Bourillot et al., 2010; Roveri et al., 2009) are also included. MSC key-surfaces: MSC onset (start of evaporite deposition in the Mediterranean); MES-CC (correlative conformity of the Messinian erosional surface; see Roveri et al., 2008a,b,c,d); M/P (Miocene–Pliocene boundary).
Modified from Manzi et al. (2012) and Omodeo-Sale' et al. (2012).

Manzi et al., 2007; Roveri et al., 2008a,b; Fig. 2f). A critical element in the model has been the recognition, distribution and interpretation of the various evaporite facies, the clastic evaporites in particular (Ricci Lucchi, 1973; Manzi et al., 2005; Lugli et al., 2010).

The model has three evolutionary stages characterized by specific evaporite associations (Fig. 3). Stages 1, 2.1 and 2.2 of the original version (CIESM, 2008), have been subsequently changed into, respectively, 1, 2 and 3 (Roveri et al., 2009; Manzi et al., 2012, 2013), with stage 3 subdivided into 3.1 and 3.2 sub-stages. During stage 1, similar to what envisaged by Clauzon et al. (1996, 2005), evaporite precipitated only in shallow-water marginal basins. In stages 2 and 3 evaporites, showing different facies and recording different hydrological conditions compared to stage 1 (see Section 3.6.2; Flecker et al., 2002; Flecker and Ellam, 2006; Topper et al., 2011; Roveri et al., 2014), reached the deeper basins now preserved both onshore (thus significantly deviating from Clauzon et al.'s model) and offshore.

3.2.3.1. Early Messinian stage (7.251–5.97 Ma): the pre-conditioning phase of the MSC. The onset of the Messinian Salinity Crisis is traditionally marked by the deposition of the Primary Lower Gypsum in the marginal basins. However, evaporite precipitation in the Mediterranean was the culmination of a series of events that changed the water chemistry, increasing salinity, as the Atlantic–Mediterranean connection was progressively restricted. In fact, Selli (1960) argued that the base of the Messinian stage should be placed at the level that coincides with the first conspicuous environmental change marked by dystrophic faunal elements, which he interpreted as the actual beginning of the salinity crisis. For this reason, the first occurrence of *Globorotalia conomiozea* (=*G. miotumida*) in the Mediterranean, now dated at 7.25 Ma, was selected as boundary criterion.

The gradual restriction of the marine connections to the Atlantic is commonly related to tectonic uplift processes in the gateway area (Duggen et al., 2003; Garcia-Castellanos and Villaseñor, 2011). Lithospheric slab detachment and roll back processes underneath the Gibraltar Arc (Villaseñor et al., 2003), possibly in combination with slab tear propagation, are recently proposed mechanisms that explain the gradual uplift during the MSC (Govers, 2009).

The first evidence of significant restriction in Mediterranean–Atlantic exchange was recorded by a reduction of deep-water ventilation all over the Mediterranean at 7.15 Ma immediately after the Tortonian–Messinian boundary (Kouwenhoven et al., 1999, 2003; Seidenkrantz et al., 2000). This event is also associated with the increased deposition of diatom-rich sediments (Tripoli Formation, Sicily) or opal-rich deposits (Upper Abad, southern Spain) observed between 7.15 Ma and 6.7 Ma throughout the Mediterranean.

A second response to ongoing restriction is the sudden drop in diversity of calcareous plankton that occurred at 6.7 Ma. This was probably caused by enhanced surface salinities, especially during summer insolation minima (Sierro et al., 1999, 2003; Blanc-Valleron et al., 2002). An intensification of bottom-water stagnation and stratification of Mediterranean waters, especially during northern hemisphere summer insolation maxima, is also thought to have occurred (Kouwenhoven et al., 1999; Sierro et al., 1999; Blanc-Valleron et al., 2002; Sierro et al., 2003; Van Assen et al., 2006).

The widespread precipitation of authigenic calcite, dolomite and/or aragonite between 6.3 and 5.97 Ma, immediately prior to the MSC onset, is another response to ongoing restriction. In addition, the complete disappearance of planktic foraminifera during summer insolation minima at this time indicates that surface waters reached salinities above the maximum tolerance of these organisms (Sierro et al., 1999; Bellanca et al., 2001; Blanc-Valleron et al., 2002; Sierro et al., 2003; Manzi et al., 2007).

The deterioration of palaeoenvironmental conditions leading up to the MSC did not progress gradually but rather stepwise, with a periodicity of 400-kyr between major steps (Hilgen et al., 2007; Hüsing et al., 2010; Fig. 5). This clearly suggests that long-term orbital cycle forcing, superimposed on a more gradual tectonic trend of gateway closure, played a critical role in the exact timing of the events (Krijgsman et al., 1999a; Hilgen et al., 2007), although the additional influence of non-gradual tectonics cannot be completely excluded.

In the case of the onset of the MSC, the most significant palaeoclimatologic changes coincide with the amplitude increase in insolation following the ~400-kyr eccentricity minimum dated around 6.0–6.1 Ma. A similar orbital configuration coincides with the MSC peak event which triggered halite deposition and erosional surfaces ~5.6 Ma, but obliquity-controlled glacial stages TG12 and TG14 are also likely to have played a role.

3.2.3.2. Stage 1 — 5.97–5.6 Ma: MSC onset and the first evaporitic stage. In marginal sub-basins, the MSC onset is marked by the deposition of the lowermost gypsum bed at ~5.97 Ma (Manzi et al., 2013). The first MSC stage is characterized by the rhythmical alternation of up to 16 beds of massive, bottom-grown selenite, 1–35 meter thick (Primary Lower Gypsum, PLG; Roveri et al., 2008a,b,d; Lugli et al., 2010) alternating with thinner shale. In situ PLG deposits are well preserved in the western and central Mediterranean areas (Betic Cordillera, Apennines, Sicily). Elsewhere, the development of PLG evaporites can be deduced from their occurrence as isolated blocks (e.g. Zakynthos, Aegean Sea; Karakitsios et al., 2013) or within resedimented deposits (e.g. Cyprus, Levant margin; Lugli et al., 2013).

The facies sequence in a gypsum-shale couplet records an evaporitic cycle characterized by a progressive upward increase of salt saturation up to a maximum and the subsequent transition to less saline conditions. The initial description and interpretation of an idealized cycle (Vai and Ricci Lucchi, 1977) included sabkha facies and subaerial exposure at its top. Reexamination of the facies, however, led Lugli et al. (2010) to conclude that the gypsum was precipitated entirely subaqueously, in shallow-water settings (<200 m) with moderate oxygenation. They also found no evidence for subaerial exposure within or at the top of the couplets.

Sections of PLG across the Mediterranean are sufficiently similar in terms of number, vertical facies distribution and overall stacking pattern of cycles that they have been used for basinwide correlation of individual cycles (Lugli et al., 2010). The well expressed gypsum–shale cyclicity characterizing both PLG and Upper Gypsum units reflects the alternation of drier versus more humid conditions which was likely controlled by orbital variation, particularly in precession (see Section 3.2.1; Vai, 1997; Krijgsman et al., 1999a,b; Hilgen et al., 2007). This assumption permits tuning to the astronomic curves (Fig. 6) and thus their accurate chronostratigraphic calibration (Krijgsman et al., 1999a; van der Laan et al., 2006; CIESM, 2008; Roveri et al., 2008b; Manzi et al., 2009; Lugli et al., 2010). Although this chronology is not proven by independent data, it fits with several lines of evidence (see Section 3.2.1) and provides a robust time framework and a key for defining the role of climate forcing on the MSC.

The common appearance of branching selenite facies from the 6th cycle upwards is a particularly good marker bed, providing a reliable correlative constraint (Fig. 7). These basin-wide stratigraphic similarities combined with the characteristic Sr isotope values which are within error of, or close to open ocean values, suggest that the PLG precipitated from a relatively homogeneous Atlantic-fed water body with a restricted outflow and a significant contribution of continental waters (Lugli et al., 2007, 2010).

Gypsum formation and/or preservation was apparently limited to water depths shallower than 200 m (De Lange and Krijgsman, 2010; Lugli et al., 2010), where PLG beds show downslope lateral transitions to dolostones and/or organic-rich shales (Manzi et al., 2007; Lugli et al., 2010; Dela Pierre et al., 2011; Manzi et al., 2011; Dela Pierre et al., 2012) at palaeo-depths likely exceeding 200 m, based on geological and paleontological features. This observation can be explained by considering the dual role of sulfate as both gypsum component and oxidant for organic matter mineralisation (De Lange and Krijgsman, 2010). In hypersaline settings, the water column is commonly stratified due to large salinity-driven density gradients. The surface layer (upper few 100 m) is regularly mixed and ventilated either on a seasonal or decadal scale, whereas the deeper water mass remains more stagnant, resulting in oxygen depletion (Fig. 8). In the absence of oxygen, sulfate is the predominant oxidant for the remineralization of organic matter (e.g. De Lange, 1986). Sustained sulfate consumption by this process leads to a diminished sulfate concentration in the deep water and undersaturation with respect to gypsum, causing it to dissolve. The balance between organic matter remineralization-related sulfate consumption and the supply of sulfate from shallow-water gypsum formation determines if the deep-water sulfate concentration becomes depleted and consequently whether gypsum is preserved there or not (De Lange and Krijgsman, 2010). A gypsum compensation depth interface separating environmental conditions favorable (above) or unfavorable (below) to gypsum preservation can be defined. The observation that anoxic environments contain no gypsum (e.g. Nurmi and Friedman, 1977) is therefore related to diminished dissolved sulfate via organic matter degradation and not to oxygen-free conditions directly.

The formation of dolomite is also linked to this process since it is known to form in low-sulfate conditions (e.g. Baker and Kastner, 1981). The process of sulfate oxidation of organic matter not only produces this low sulfate environment but also increases the dissolved carbonate concentration while the dissolution of gypsum provides additional Ca. It is therefore consistent to find dolomite forming in such deep-water, gypsum-free settings, while coeval gypsum precipitation occurs in shallow, marginal settings. These processes have important implications for the correct definition of the onset of the MSC because locally the lowermost gypsum bed may be significantly younger than 5.97 Ma (e.g. Apennines, Tertiary Piedmont Basin; Lugli et al., 2010; Dela Pierre et al., 2012; Gennari et al., 2013). Consequently the start of the MSC is best defined by the complete and sustained disappearance of microplanktic assemblages that in evaporite-free successions can be cyclostratigraphically dated at around 5.97 Ma (Manzi et al., 2007; Dela Pierre et al., 2011, 2012; Gennari et al., 2013; Manzi et al., 2013).

The top of the PLG deposits is usually incised by a high-relief erosional surface (the Messinian erosional surface — MES) developed during the subsequent MSC stages and showing evidence of subaerial exposure (e.g. the karstic surface in the Vena del Gesso basin; Marabini and Vai, 1988). In the shallow-water portions of marginal basins the MES may

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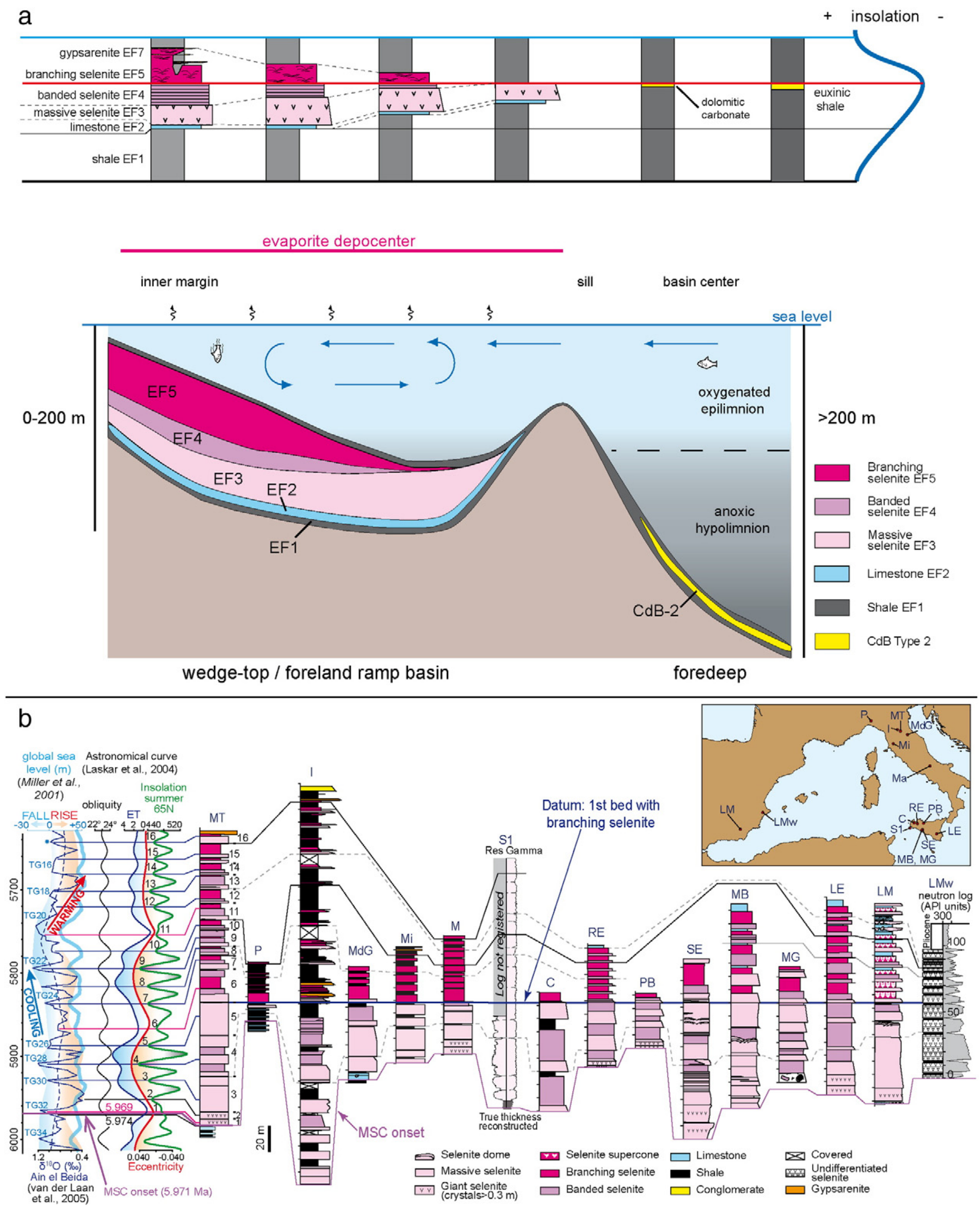

Fig. 7. a) — The different facies sequences observed in Primary Lower Gypsum (PLG) cycles and their envisaged tuning with insolation curve (above) and distribution along a idealized depositional profile; facies interpreted as recording the peak of brine saturation correlate with the insolation minima; in this model PLG cycles are suggested to form in silled sub-basins. b) — High-resolution basinwide correlation of PLG outcrop successions and tuning with insolation and oxygen curve; note the impressive similarity in terms of number of cycles (up to 16), absolute and relative thickness of individual cycles (with the 3rd, 4th and 5th attaining the maximum thickness) and stacking pattern of facies (with branching selenite only appearing from the 6th cycle upwards).
Modified from Lugli et al. (2010) and Manzi et al. (2013).

cut PLG and pre-MSC units and is sealed by stage 3 deposits (Fig. 3). This is clearly observed in the Northern Apennines, Calabria, Sicily and Betic Cordillera (Nijar) basins (Roveri et al., 2001; Fortuin and Krijgsman, 2003; Bassetti et al., 2004; Roveri et al., 2003; CIESM, 2008; Roveri et al., 2008a; Omodeo-Sale' et al., 2012).

3.2.3.3. Stage 2 (5.6–5.55 Ma) — the acme of the MSC. The acme of the salinity crisis was reached during the second stage (5.6–5.55 Ma; Fig. 3; Hilgen et al., 2007; stage 2.1 of CIESM, 2008), which was dominated by thick primary halite (Sicily, Calabria, Cyprus) and clastic gypsum deposits (Fig. 9; Sicily, Calabria, Apennines, Spain, Cyprus, Crete), grouped

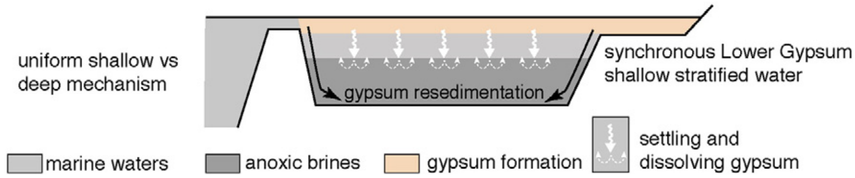

Fig. 8. Schematic illustration of a scenario explaining the deposition of the Primary Lower Gypsum (PLG) evaporites. Gypsum precipitates within shallow-water levels and is preserved in shallow settings while the deep basins are characterized by dolomite formation and gypsum resedimenation.
Modified from De Lange and Krijgsman (2010).

into a unit labeled Resedimented Lower Gypsum (RLG, Roveri et al., 2008a,b). This stage records a phase of widespread subaerial erosion with the development of the Messinian erosional surface (MES) usually related to a high-amplitude Mediterranean base-level drop (see Section 4.2). The TG14 and TG12 glacials occurred during this stage and probably caused further restriction in the sub-basins' hydrology and a significant reduction of the Mediterranean–Atlantic exchange which likely resulted in the blockage of the Mediterranean outflow (Meijer and Krijgsman, 2005; Hilgen et al., 2007).

The stage is also characterized by a phase of Mediterranean-wide tectonic activity (Pedley and Grasso, 1993; Butler et al., 1995; Robertson et al., 1995; Butler et al., 1999; Roveri et al., 2001; Duggen et al., 2003;

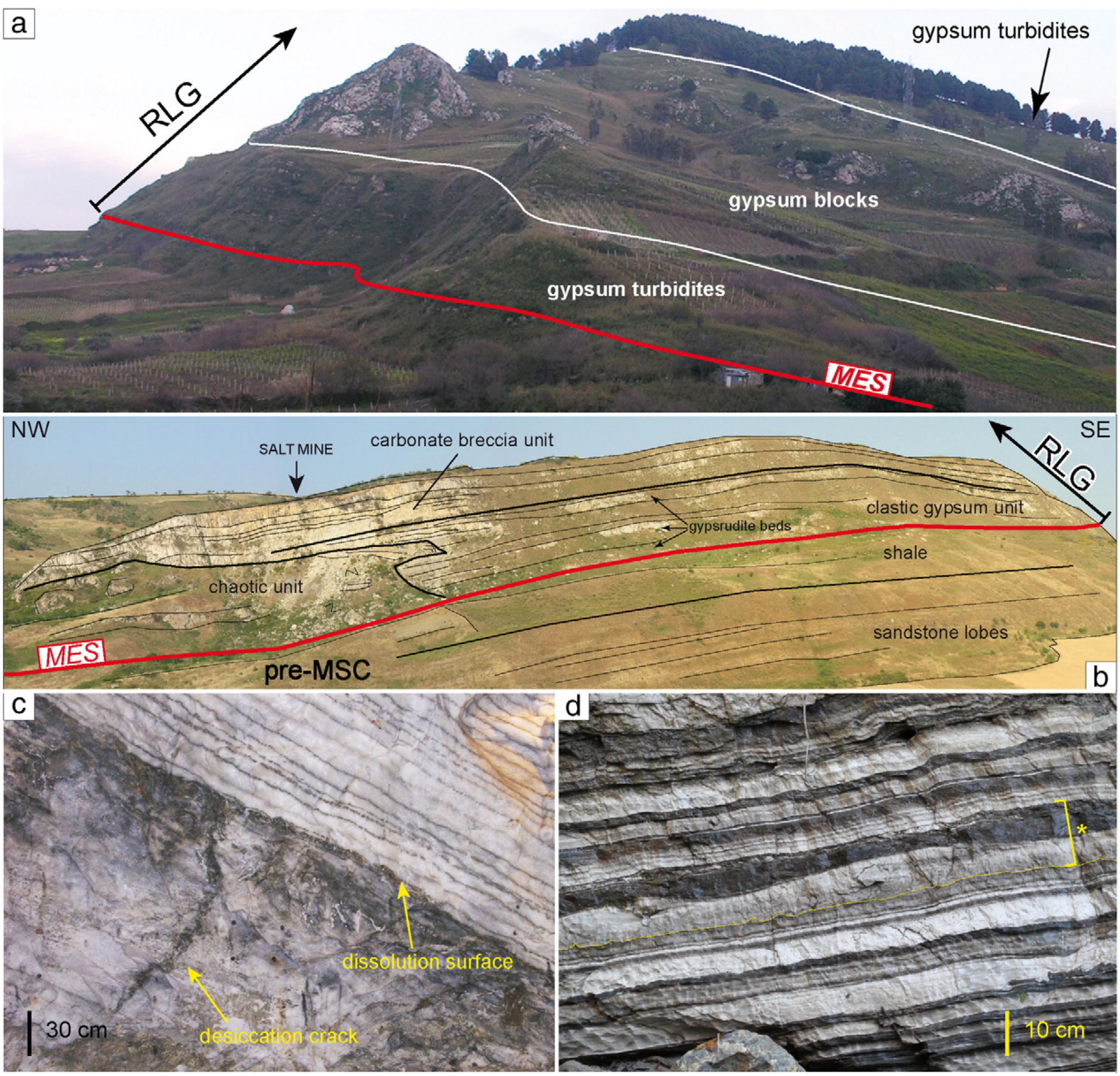

Fig. 9. Evaporitic deposits included in the Resedimented Lower Gypsum unit (RLG). (a) Primary Lower Gypsum disarticulated blocks sandwiched between gypsum turbidites. Belice basin, Sicily (from Roveri et al., 2006). (b) different clastic facies included in the Resedimented Lower Gypsum unit resting unconformably on the late Tortonian early Messinian deposits of the Terravecchia Formation. The chaotic body, to the left, include bocks of the lower gypsum unit. Balza Bovolito, Petralie, Sicily. MES, Messinian erosional surface. The carbonate breccia unit is formed by the Calcare di Base type 3 (Manzi et al., 2011). (c) dissolution surface and desiccation crack in the upper part of the salt unit (boundary of mine horizons B and C). Church section, Realmonte mine (−28 m). Annual scale cycles are visible above the surfaces. (d) clastic gypsum-bearing graded beds deposited by low-density gravity flows. Arrows indicate a tripartite turbiditic bed with lower crude laminated division, overlay by a climbing megaripple interval abruptly capped by dark euxinic shales representing the finer-grained tail of the flow. The base is slightly erosive and shows small fluid-escape features. Fanantello riverbed, Sapigno syncline, northern Apennines.

Fortuin and Krijgsman, 2003; Roveri et al., 2003; Roveri and Manzi, 2006; CIESM, 2008; Fortuin and Dabrio, 2008; Roveri et al., 2008a; Omodeo-Sale' et al., 2012), which is also suggested by the angular discordance often associated with the MES, testifying a local or regional-scale tectonic driver (Roveri et al., 2003). Initial interpretation of RLG deposits suggested they were coeval with the Primary Lower Gypsum (e.g. Rouchy and Caruso, 2006). However, in marginal basins such as the Apennines and Sicily, the MES, which cuts through PLG, can be traced to its correlative conformity at the base of RLG deposits where they overlie gypsum-free sediments coeval of the PLG unit (Fig. 12; Manzi et al., 2007; Roveri et al., 2008a,b). This stratigraphic relationship clearly indicates that RLG deposits post-date precipitation of the PLG.

A sea-level fall of still uncertain amplitude (see Section 4.2), the resulting pressure release and rapid fluid migration (Lazar et al., 2012; Bertoni et al., 2013; Iadanza et al., 2013), and steepening of the basin margins due to crustal loading by halite (Govers et al., 2009), are all factors linked by complex feedbacks which are likely to have promoted slope instability and gravity failure. As a consequence, the PLG units were deeply eroded and resedimented producing the gypsum clastic deposits ranging from mountain sized mass-wasting deposits to gravity flows that are exposed spectacularly in various depocentres in Italy (Fig. 9a, d) (Ricci Lucchi, 1973; Roveri et al., 2003; Manzi et al., 2005; Roveri and Manzi, 2006; Manzi et al., 2011), Greece (Kontopoulos et al., 1997), Cyprus (Robertson et al., 1995), Spain (Fortuin and Krijgsman, 2003; Omodeo-Sale' et al., 2012) and also recognized in the canyon fills of the Israel margin (Lugli et al., 2013). This mechanism of rapidly transporting large volumes of gypsum to deep-water settings is likely to have reduced both the time available for interaction with gypsum-undersaturated deep water and the degree of undersaturation itself, allowing gypsum preservation in deep environments.

The primary evaporitic deposits during this phase were mostly gypsum cumulates precipitating along the basin margins which were not eroded (e.g. Sicily; see also Section 3.4.1; Manzi et al., 2012) and thick and extensive primary halite and K–Mg salts (kainite, carnallite and minor bishofite) that rapidly filled some of the sub-basins, as testified by shallowing upward sequences that culminate in exposure (Fig. 9c; e.g. Sicily; Lugli et al., 1999). Both primary facies are characterized by an impressive small-scale lithological cyclicity (Ogniben, 1957) that has been related to annual or pluri-annual cyclicity (Bertini et al., 1998). Statistical studies carried out on these primary evaporites in the Northern Apennines (based on laser ablation on gypsum; Galeotti et al., 2010) and, in a longer record, in Sicily (based on petrographic and sedimentological characterization of gypsum and halite; Manzi et al., 2012; Fig. 10b) confirmed the annual periodicity of the cycles. These authors argue that the acme of the MSC was not permanent evaporitic conditions, but a dynamic situation with seasonal high-amplitude climatic oscillations. Spectral analysis reveals significant climatic periodicity peaks at around 3–5, 9, 11–13, 20–27 and 50–100 years that could be related to quasi-periodic oscillations like the Quasi-Biennial Oscillation, El Niño Southern Oscillation, Atlantic Multidecadal Oscillation, Pacific Decadal Oscillation and decadal to secular lunar- and solar-induced cycles (Galeotti et al., 2010; Manzi et al., 2012). One interesting implication is that even during insolation minima, evaporitic conditions were achieved but persisted only at the seasonal scale. Moreover the number of primary elementary cycles suggests, at least for the Sicilian succession, that the precipitation of halite and lateral equivalent deposits could have taken place within a couple of thousand years (Bertini et al., 1998; Roveri et al., 2008a; Manzi et al., 2012).

On Sicily, the halite deposits are encased within RLG deposits; the latter in turn interfinger with a limestone/dolostone evaporitic–microbialitic unit known as the "Calcare di Base" (CdB; Decima et al., 1988; Guido et al., 2007), because it was thought to have formed during the initial phase of the MSC (Decima and Wezel, 1971; Rouchy and Caruso, 2006). More recently however, it has been suggested that the Calcare di Base actually comprises at least three carbonate units with different origins and stratigraphic positions (CdB 1, 2, 3; Roveri et al., 2008a; Manzi et al., 2011). At a regional-scale, the most common unit, consisting of brecciated limestones (CdB 3), is floored by an unconformity (Roveri et al., 2008a; Manzi et al., 2011). This, combined with its exclusive association with the RLG deposits, led Roveri et al. (2008a) to suggest that the Calcare di Base formed

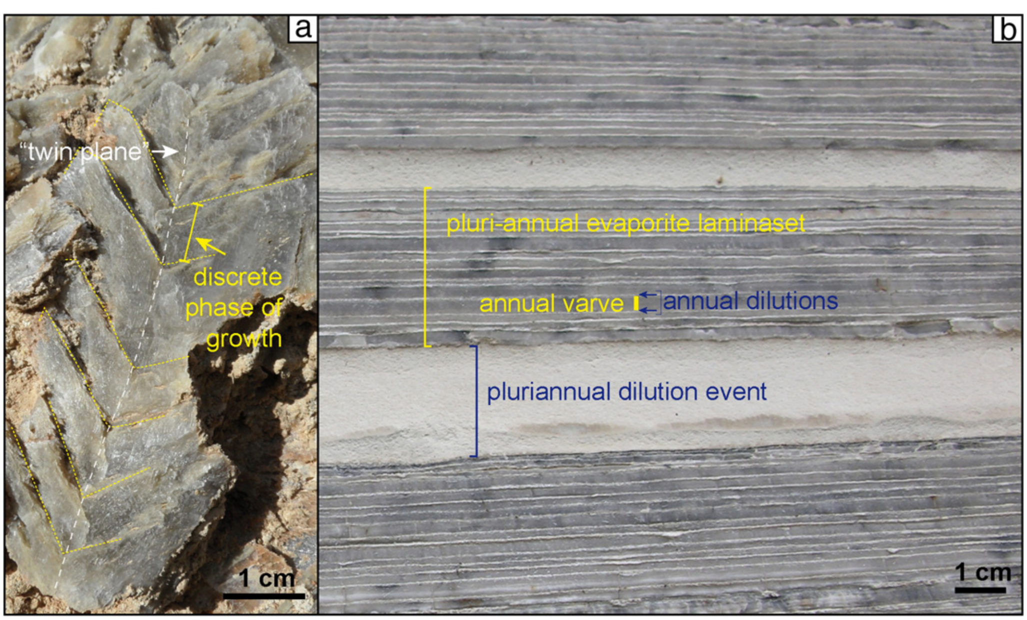

Fig. 10. Examples of small-scale lithological cyclcity in the MSC evaporites. (a) Perales (Sorbas basin, Spain), cycle 2 of the Primary Lower Gypsum unit (stage 1). Notice the discrete phase of growth of the selenite crystals associated with a "telescopic" reduction of the crystal size and marked by a thin carbonate/shale veneers. (b) Pasquasia section, Caltanissetta basin, Sicily. Primary deposits included in the RLG (Resedimented Lower Gypsum) unit (stage 2). Notice the mm-thick laminae of cumulate gypsum representing annual varves each one showing a characteristic inverse gradation (Ogniben, 1957). They are arranged in laminaset or cluster separted by cm-thick marl interval representing pluri-annual events of dilution of the water mass (Manzi et al., 2012).

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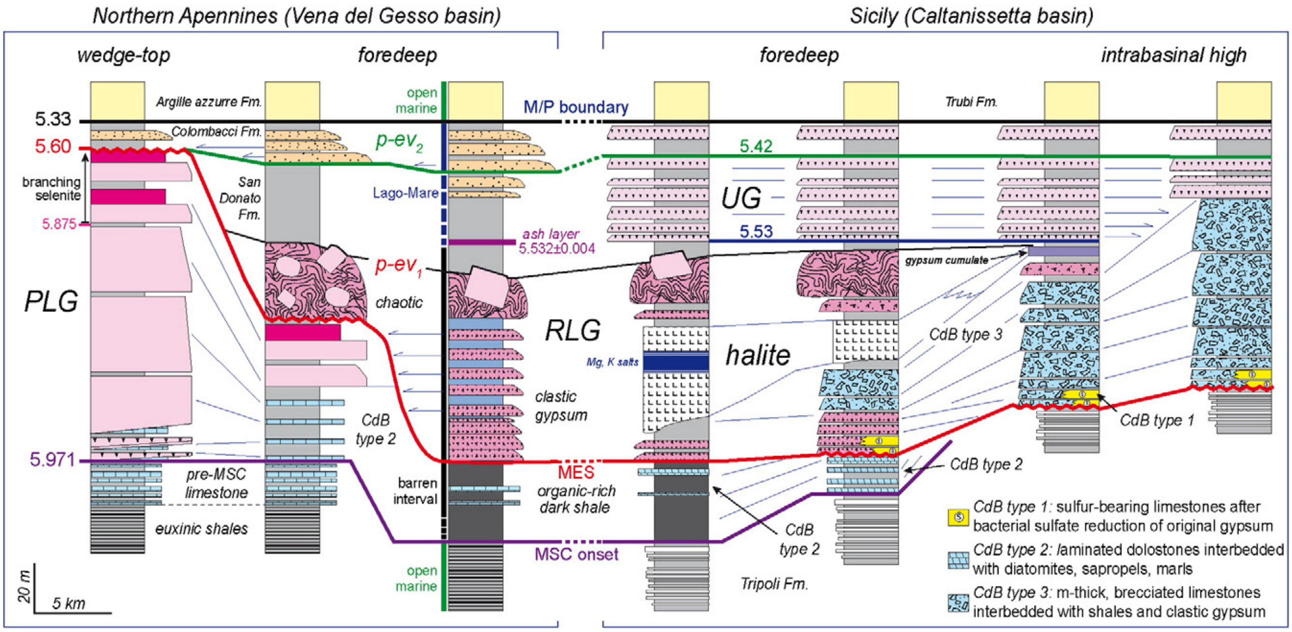

Fig. 11. Stratigraphic distribution of Calcare di Base (CdB) types and their relationships with the other Messinian units. Note that the largest volume of CdB deposits (i.e., types 1 and 3) of Sicily lies above the Messinian erosional surface (MES), thus largely postdating the Messinian salinity crisis onset, which is marked by the base of type 2 CdB. Fm., Formation; PLG, Primary Lower Gypsum; RLG, Resedimented Lower Gypsum; UG, Upper Gypsum; MSC, Messinian salinity crisis; MES, Messinian erosional surface; M/P — Miocene–Pliocene.
Modified after Manzi et al. (2011).

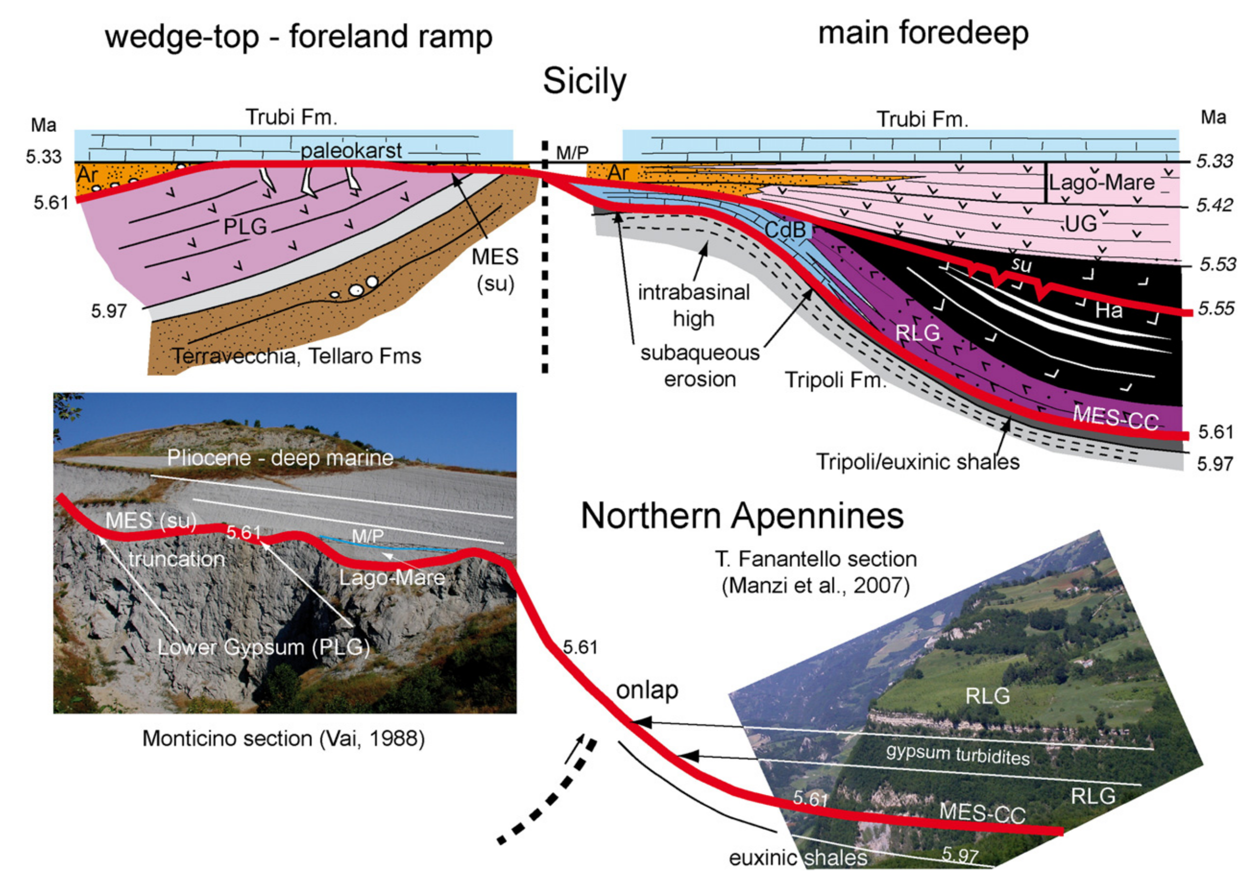

Fig. 12. Shallow to deep-water setting physical–stratigraphic correlation in marginal basins of Sicily and Northern Apennines. The subaerial unconformity cutting the Primary Lower Gypsum (PLG) splits downslope in two surfaces: i) the MES-CC represents the correlative conformity of the MES, it marks the base of the Resedimented Lower Gypsum (RLG) and ii) the subaerial unconformity (su) at the top of the shallowing-upward lower salt unit in Sicily (Ha). These surfaces enclose a forced regression wedge deposited during a relative sea level falling stage associated to the acme of the MSC. Ar, Arenazzolo; CdB, Calcare di Base; UG, Upper Gypsum; M/P, Messinian/Pliocene boundary.
Modified after Roveri et al. (2008d).

during stage 2 above tectonically active intrabasinal highs (Butler et al., 1995) and was penecontemporaneously resedimented together with the PLG (Fig. 9b) in adjacent depocenters (Manzi et al., 2011; Fig. 11). The origin of brecciated limestones has been also related to the migration of hydrocarbon-charged fluids (Iadanza et al., 2013).

3.2.3.4. Stage 3 (5.55–5.33 Ma) — Upper Evaporites and Lago Mare event(s). After the peak of the crisis, the third MSC stage is characterized by a completely different scenario. Both selenite and cumulate gypsum facies (Upper Gypsum — UG) precipitated mainly in shallow-water sub-basins in southern and eastern areas (i.e. Sicily and Cyprus), while in the northern and western areas evaporite-free clastic deposits dominate. Distinctive very low values of $^{87}Sr/^{86}Sr$ have been measured on both evaporites and fossils from this period (Fig. 13a) and this, combined with the widespread development of shallow-water environments characterized by brackish to fresh-water fauna and flora with Paratethyan affinities (ostracods — *Cyprideis*, *Tyrrenocythere*, Leptocytheridae, Candonids, molluscs — *Congeria*, *Melanopsis*, *Dreissena*, *Limnocardium*, dinocysts — *Galeacysta etrusca*, *Pyxidinopsis psilata*, *Spiniferites cruciformis*; Orszag-Sperber, 2006; Rouchy and Caruso, 2006; Roveri et al., 2008b), suggest a substantial dilution of surface waters, locally punctuated by episodic, precession-driven, evaporitic events (Upper Gypsum; Manzi et al., 2011). These features are consistent with the enduring concept of a Lago-Mare event (Gignoux, 1936; Ruggieri, 1967; Orszag-Sperber, 2006) and suggest that the Mediterranean basin underwent substantial palaeogeographic and palaeoclimatic modifications, leading to important hydrological changes.

Stage 3 gypsum-bearing successions commonly show well-developed rhythmic couplets, which, similar to PLG evaporitic cycles, have been interpreted as a response to high-amplitude, precession-driven climate oscillations (Fig. 6; Vai, 1997; Hilgen and Krijgsman, 1999; Krijgsman et al., 1999b; Manzi et al., 2009). Exposures on Sicily and Cyprus show alternations of meter-thick gypsum beds and marl horizons 1–10 m thick containing typical Lago-Mare flora and fauna (Bonaduce and Sgarrella, 1999; Rouchy et al., 2001; Bertini, 2006; Gliozzi et al., 2007; Londeix et al., 2007). The most complete stage 3 successions crop out in Sicily where 10 sedimentary cycles (7 with gypsum) have been recognized and correlated regionally between the top of the stage 2 deposits and the base of the Pliocene (Van der Laan et al., 2006; Hilgen et al., 2007; Manzi et al., 2009; see also Section 3.2.1). Gypsum beds show different facies and stacking pattern (number and thickness of cycles) with respect to stage 1 PLG deposits (Manzi et al., 2009), but thought to have been deposited in similar depositional settings (Manzi et al., 2009, 2011), albeit from water with a much lower ratio of ocean/continental input.

Stage 3 evaporite-free successions in the northern and western Mediterranean are well exposed in the Northern Apennines and Betic Cordillera (Sorbas and Nijar) sub-basins (Dabrio and Polo, 1995; Roep et al., 1998; Roveri et al., 1998; Krijgsman et al., 2001; Fortuin and Krijgsman, 2003; Bassetti et al., 2004; Roveri et al., 2008b, 2009; Omodeo-Sale' et al., 2012). Here, the successions comprise two distinct sequences; an upper unit (p-ev$_2$; Fig. 3) dominated by coarse-grained, siliciclastic fluvio-deltaic deposits; and an overlying a unit (upper part of p-ev$_1$; Fig. 3) usually characterized by finer-grained deposits. This also suggests a modification to the precipitation regime which started ~5.42 Ma (Roveri et al., 2008b) and allows to subdivide stage 3 into the two sub-stages 3.1 and 3.2 (Fig. 3).

The evidence of Mediterranean exchange with other water masses (e.g. Atlantic and Paratethys) during stage 3 is complex. The widespread development of hypohaline environments has commonly been related to the complete isolation of the Mediterranean from the Atlantic and to the incursion of Paratethyan waters (Orszag-Sperber, 2006; Rouchy and Caruso, 2006; Roveri et al., 2008b). In the Apennines, non-marine flora and fauna with Paratethyan affinities (e.g. the ostracod *Loxocorniculina djafarovi* and the dinocysts of the *Galeacysta etrusca* complex) appear more consistently just below the base of the upper subunit (p-ev$_2$; Roveri et al., 2008b) and this is followed by a significant increase in the abundance and diversity of the Lago-Mare fauna and flora in the upper sub-unit itself. This suggests progressively more efficient intra-basinal connections and water exchanges with the Paratethys and is consistent with a relative base-level rise throughout the Mediterranean, which has been inferred from the aggradational stacking pattern of the upper sub-unit (p-ev$_2$; Roveri et al., 2008b). The occurrence of marine fish (Carnevale et al., 2006, 2008) and long chain alkenones (Mezger, 2012) suggests the possible persistence, continuous or episodic, of an Atlantic connection at that time. This could also explain the sporadic occurrence of anomalously small or “dwarf” foraminifers in stage 3 deposits (see Iaccarino et al., 2008 and references therein).

3.2.3.5. MSC Carbonate platforms: the Terminal Carbonate Complex. An unusual type of carbonate platforms, oolite- and microbialite-dominated, with minor *Porites*, developed during the MSC, especially in the Atlantic gateway region (Betic and Rifian corridors). They are referred to as the “Terminal Carbonate Complex” (Esteban, 1979; Martin and Braga, 1994; Braga et al., 2006). These platforms record a dramatic decline of coralline reef builders and the progressive increase in microbial activity. These characteristics are commonly interpreted in terms of highly stressed conditions related to the reduction of ocean connections and salinity fluctuations (Esteban, 1979; Martin and Braga, 1994). The age of the Terminal Carbonate Complex is highly controversial, but it has been considered by many authors to postdate MSC stage 1 (Martin and Braga, 1994; Roep et al., 1998; Braga et al., 2006; Bourillot et al., 2010). However, according to others, these deposits could be at least partially coeval with the Primary Lower Gypsum unit and therefore record the onset (Cunningham et al., 1994, 1997; Cornée et al., 2006) and/or the whole duration of the MSC (Roveri et al., 2009).

3.2.3.6. The Zanclean flooding (5.33 Ma): the end of the MSC. The return to fully and stable marine conditions throughout the Mediterranean area marks the end of the Messinian salinity crisis. This event has been used to define the GSSP of the Zanclean stage and hence of the Pliocene series in the Eraclea Minoa section (Sicily; Van Couvering et al., 2000; see Section 3.2.1). In the Mediterranean, this substantial paleoceanographic change is typically recorded by a sharp lithological and paleontological boundary, suggesting a geologically instantaneous event. Commonly this has been thought to imply the abrupt collapse of the Gibraltar sill and the consequent catastrophic flood (a big waterfall or a cataract, according to different models; Hsü et al., 1973a) of Atlantic waters into the (desiccated) Mediterranean basin (Blanc, 2002; Meijer and Krijgsman, 2005; Garcia-Castellanos et al., 2009). The almost instantaneous and synchronous nature of this event has been proven by several detailed studies in many sections and cores (Iaccarino and Bossio, 1999; Iaccarino et al., 1999; Gennari et al., 2008; Roveri et al., 2008c) throughout the Mediterranean area. Whereas the planktonic foraminifer record indicates an abrupt return to marine conditions, the benthic foraminifers record of the deep Tyrrhenian and Western Mediterranean seas (Ocean Drilling Project — ODP — sites 974B and 975B; Fig. 1) suggest that the restoration of open marine condition was not instantaneous but that the circulation remained restricted for a time interval encompassing two or three precessional cycles (Iaccarino et al., 1999).

The amplitude of the sea-level rise that eventually led to the reestablishment of marine conditions in the Mediterranean basin is not known due to the lack of reliable paleodepth indicators. Shallow to relatively deep marine deposits (Trubi Fm., Argille Azzurre Fm.) sharply overly the MSC stage 3 Lago-Mare deposits. The boundary is often characterized by a cm- to dm-thick transitional interval typically enriched in organic matter (the “black layer”; Roveri et al., 1998, 2008b) whose meaning is not fully understood.

Please cite this article as: Roveri, M., et al., The Messinian Salinity Crisis: Past and future of a great challenge for marine sciences, Marine Geology (2014), http://dx.doi.org/10.1016/j.margeo.2014.02.002

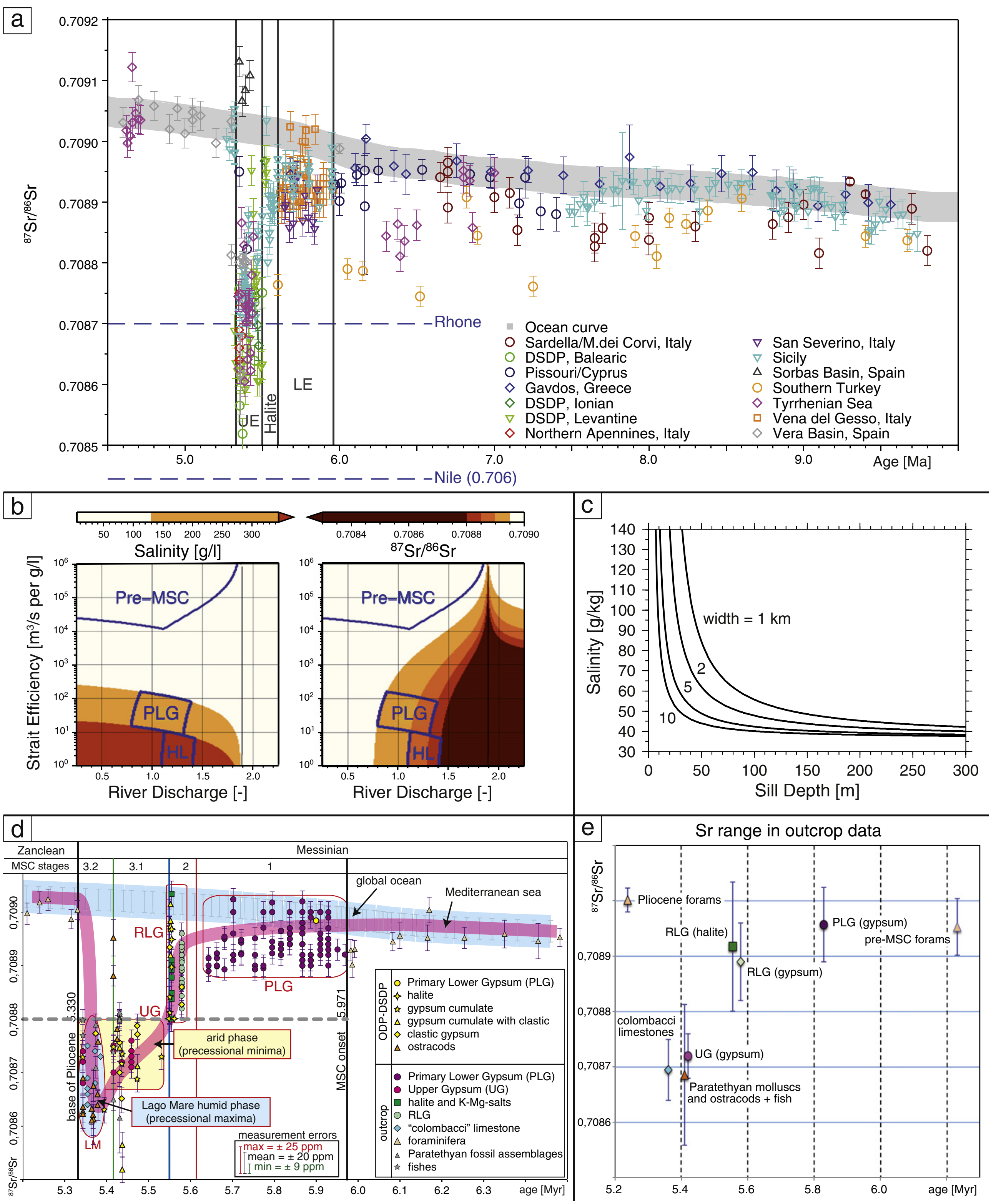

Fig. 13. (a) Late Miocene–Pliocene $^{87}Sr/^{86}Sr$ data from various localities in the Mediterranean Sea, together with the global ocean curve. (b) Salinity (left panel) and Sr ratio (right panel) calculated as a function of river discharge and efficiency of exchange with the Atlantic Ocean. Strait efficiency is the factor of proportionality between water transport and density difference between basin and ocean: the larger the factor, the wider/deeper the gateway and vice versa. Superimposed on the model results are three fields that encompass the observed range of salinity and Sr ratio, divided into the pre-MSC sediments, the Primary Lower Gypsum (PLG) and Halite (HL). (c) Average salinity of the Mediterranean basin as a function of depth of the connection to the Atlantic Ocean and for different strait widths, based on the assumption that the exchange is subject to hydraulic control (Meijer, 2012). Results are for a uniform-width channel, a net evaporation of 0.5 m/yr, and assuming the interface between in- and outflow to be situated halfway the water column. (d) $^{87}Sr/^{86}Sr$ curve during the Messinian salinity crisis in the Mediterranean. (e) variability in the $^{87}Sr/^{86}Sr$ ratio for the different unit of the Messinian salinity crisis).
See Topper et al. (2011) for details and a list of data sources. River discharge is shown as a multiplication factor where unity equals the river input into the Mediterranean inferred from a global climate model simulation by Gladstone et al. (2007). Panel b is after Topper et al. (2011). Panel d is from Roveri et al. (2014).

3.3. The offshore perspective

After the pioneering studies of the 1950s to 70s (see Introduction), a renewed interest in understanding the MSC based on seismic data has persisted since the 2000s. Numerous studies illustrate the complexity of the offshore seismic MSC signal and the spatial, temporal and geometrical variability of the associated depositional units and surfaces (e.g. Hsü et al., 1973a,b; Ryan, 1976; Hsü et al., 1978; Ryan and Cita, 1978; Stampfli and Höcker, 1989; Savoye and Piper, 1991; Escutia and Maldonado, 1992; Guennoc et al., 2000; Lofi et al., 2005; Sage et al., 2005; Bertoni and Cartwright, 2006; Maillard et al., 2006; Netzeband et al., 2006; Schattner et al., 2006; Bertoni and Cartwright, 2007a,b; Gillet et al., 2007; Lofi and Berné, 2008; Bache et al., 2009; Ghielmi et al., 2010; Garcia et al., 2011; Just et al., 2011; Lofi et al., 2011a,b; Maillard et al., 2011; Obone Zue Obame et al., 2011; Bowman, 2012; Gaullier et al., 2014). This variability implies that a fully understanding of the offshore MSC record can be achieved only through basinwide studies. Only the uppermost deep basin deposits have ever been sampled by scientific drillings (see Section 3.3.4). Although commercial boreholes have been drilled through the MSC evaporitic sequence in the eastern Mediterranean, the entire Messinian succession has not been recovered and the data remain largely inaccessible to the academic world. As a result we are currently lacking lithostratigraphic control on the vast majority of the MSC deposits.

3.3.1. Updated definition of seismic units and surfaces

The seismic markers associated with the MSC in the offshore domain correspond to erosional surfaces and depositional units and associated bounding surfaces. Among them, the most characteristic are undoubtedly the widespread Messinian Erosional Surface (MES), observed on

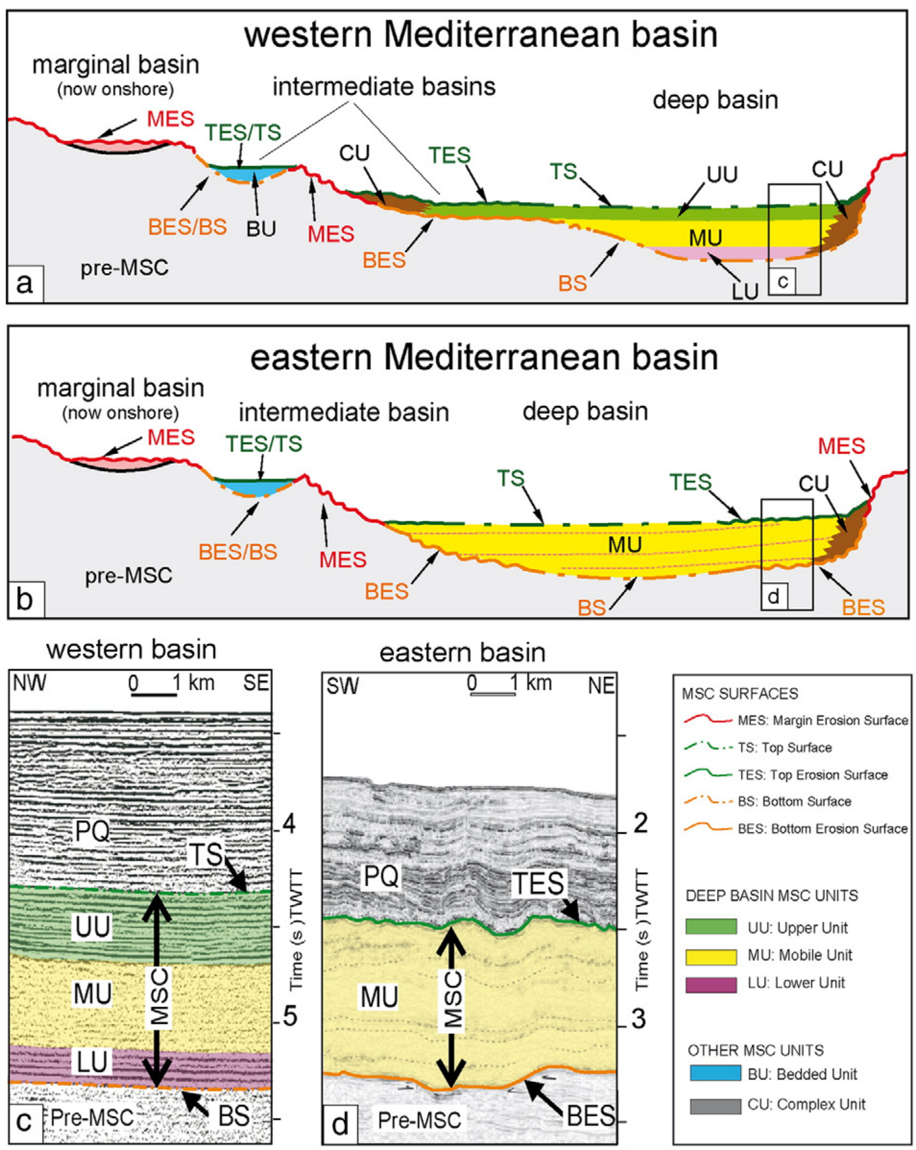

Fig. 14. Schematic conceptual sketches (not to scale) across the western (a) and eastern (b) Mediterranean basins, illustrating the organization of the MSC seismic markers from the margins down to the deep basins at the end of the MSC. In this scheme, marginal basins are considered as shallow water basins having accumulated evaporites during the first step of the MSC, before the drawdown phase (e.g. Sorbas Basin) (see text). Intermediate-depth basins containing Bedded Units (BU) are located deeper, but remain shallower compared to the deepest areas of the basins (e.g. West Corsica, Valencia and Syrthe basins) (modified from Lofi et al., 2011a and b). Seismic profiles from (c) the western (Gulf of Lions, modified from Lofi et al., 2005) and (d) eastern (Levant Basin, modified from Bertoni and Cartwright, 2005) Mediterranean deep basins illustrating the main differences in the present day MSC seismic records. In the western basin, the MSC seismic sequence consists of up to 3 distinct seismic units (seismic trilogy: UU, MU (~1000 m of halite), LU). In the eastern basin, only one seismic unit is clearly recognized (MU, >2000 m) that can be divided in 6 sub-units.

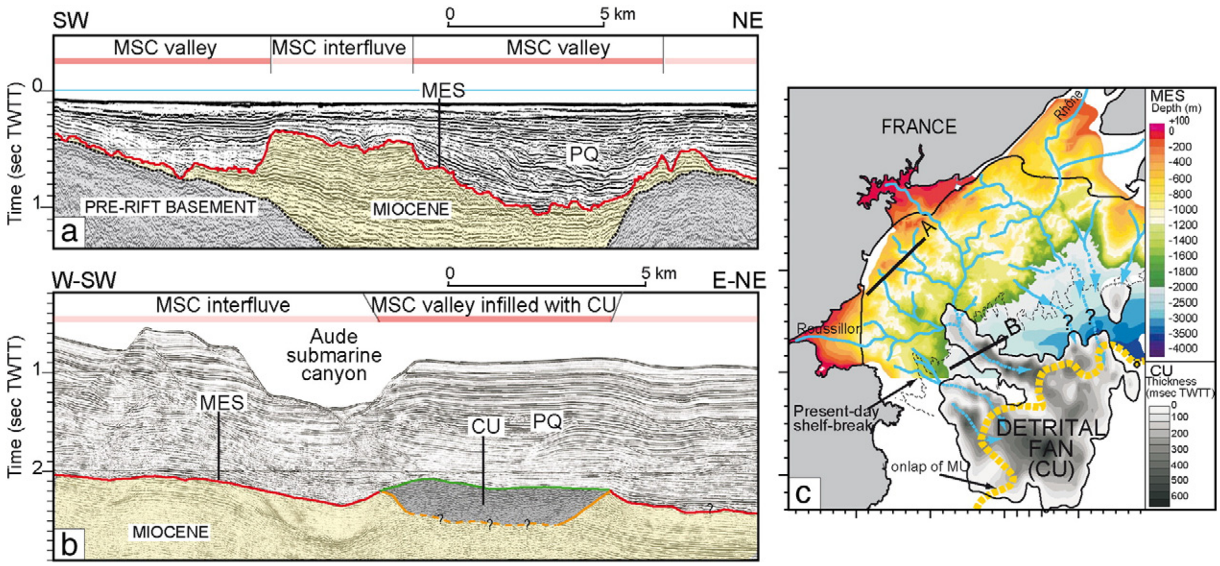

Fig. 15. Seismic profiles in the Gulf of Lions illustrating: (a) the Margin Erosion Surface (MES) buried below the Plio-Quaternary deposits beneath the inner shelf. The MES is a prominent unconformity displaying here two palaeo-MSC valley morphologies and (b) the presence of a Complex Unit (CU) infilling a MSC valley beneath the slope. (c) Present-day depth map of the MES and isochron map of CU. The MES displays drainage network and CU accumulated as fan shaped deposits in the downstream part of the network.
Panel a is modified from Lofi et al. (2011a,b). Panel b is modified from Lofi et al. (2005). Panel c is modified from Lofi et al. (2011a).

the margins and onshore (see Section 3.2.3.1), and the thick salt layer (Mobile Unit, MU) mostly limited to the deepest parts of the basins. A detailed review at Mediterranean scale of the seismic characteristics and nomenclature of those seismic markers has been recently proposed by Lofi et al. (2011a,b) (Fig. 14).

3.3.2. MSC surfaces

The MSC surfaces have been defined based on their geometrical relationship with the pre-MSC units, the Messinian deposits, and the Plio-Quaternary cover. The age of these surfaces is typically not known and it could differ from place to place. While only one widespread erosion surface is generally observed on the margins (Margin Erosion Surface), several surfaces, some erosional, some not, are observed at the bottom (Bottom Surface, Bottom Erosion Surface), within (Internal Erosional Surface) or at the top (Top Surface, Top Erosion Surface) of the MSC depositional units.

3.3.2.1. Margin Erosion Surface. This polygenic surface is the offshore prolongation of the onshore Messinian Erosional Surface; consequently the acronym MES is used in both domains. The MES is one of the most striking features, well identified on basin margins at the base of the Pliocene–Quaternary unit (Fig. 15). It typically forms a prominent reflector, associated with truncations of underlying seismic reflectors or an angular discordance between pre-MSC and Pliocene reflectors (Ryan, 1976; Ryan and Cita, 1978; Lofi et al., 2003, 2005, 2011a and references therein; Roveri and Manzi, 2006; Gillet et al., 2007). Numerous seismic investigations showed that complex Messinian drainage networks developed across the MES along the Egyptian margin (Barber, 1981; Aal et al., 2000; Loncke et al., 2006), the Gulf of Lions shelf (Gennesseaux and Lefebvre, 1980; Guennoc et al., 2000), the Ebro margin and Valencia trough (Stampfli and Höcker, 1989; Urgeles et al., 2011) and the Po Plain and Adriatic basin (Roveri et al., 2005; Ghielmi et al., 2010, 2013). In several places they extend onshore and connect with valleys or narrow incisions like the Nile (Chumakov, 1973), Rhône (Clauzon, 1973) and Sahabi canyons (Nicolai, 2008). Some of these Messinian valleys can be locally superimposed on pre-existing canyons (Bertoni and Cartwright, 2006; Lofi and Berné, 2008) or show several phases of incisions that could be related either to pre-MSC cuttings or to multi-stepped incisions during the MSC. In the Alboran Basin the MES displays several erosional features related by some authors (Garcia-Castellanos et al., 2009; Estrada et al., 2011) to the flooding event at the end of the MSC.

Offshore, the MES extends basinward to the pinch-out of the MSC depositional units accumulated in the deepest parts of the basins. There, the MES splits into at least two surfaces, extending at the bottom (BS, Bottom Surface) and at the top (TS, Top Surface) of the MSC deposits and showing evidence for local erosion (Lofi et al., 2011a).

The MES is generally interpreted as subaerial as a result of a huge sea-level drop associated with the MSC acme during stage 2. However, this surface is most likely a diachronous and polygenetic feature (Lofi et al., 2011b) resulting from the combination of several processes including early stage subaqueous large-scale mass-wasting (Roveri et al., 2001; Lofi et al., 2005), rivers downcutting as they adjust to a new base level (Clauzon, 1973; Lofi et al., 2005; Loget and Van Der Driessche, 2006; Urgeles et al., 2011) and possibly periods of sea-level stagnancy (Lofi et al., 2005; Just et al., 2011), marine abrasion (Bache et al., 2012) and/or dissolution of carbonate rocks. Some authors suggested that the MES developed during a moderate amplitude sea-level drop, mostly as a result of subaqueous processes (Roveri et al., 2001; Roveri and Manzi, 2006; Martínez del Olmo, 2011), including the cascading of dense water down the shelf (Roveri et al., in press). In the Black Sea, a single erosional surface is observed on the seismic lines (Gillet et al., 2007; Tari et al., 2009); according to some authors the surface is overlain by Pliocene deposits (Popescu, 2010) and is tentatively related to the MSC (Gillet et al., 2007).

3.3.2.2. Bottom Surfaces (BS, BES). The BS marks the base of the MSC deposits in the deep basins (Fig. 14). When associated with erosion and/or angular unconformities, the surface is labeled BES (Bottom Erosion Surface). The BES/BS is the basinward extension of the MES beneath the onlapping evaporitic units (Ryan, 1978; Ryan and Cita, 1978; Lofi et al., 2005; Bertoni and Cartwright, 2006; Maillard et al., 2006; Fig. 16a).

3.3.2.3. Top Surfaces (TS, TES). The TS is a correlative conformity surface separating MSC deposits from the overlying Plio-Quaternary units (Figs. 14 and 16) in the deepest part of the basins. In shallower areas (Valencia basin; Ryan and Cita, 1978; Escutia and Maldonado, 1992; Maillard et al., 2006); East Corsica Basins; Thinon et al., 2004; Thinon

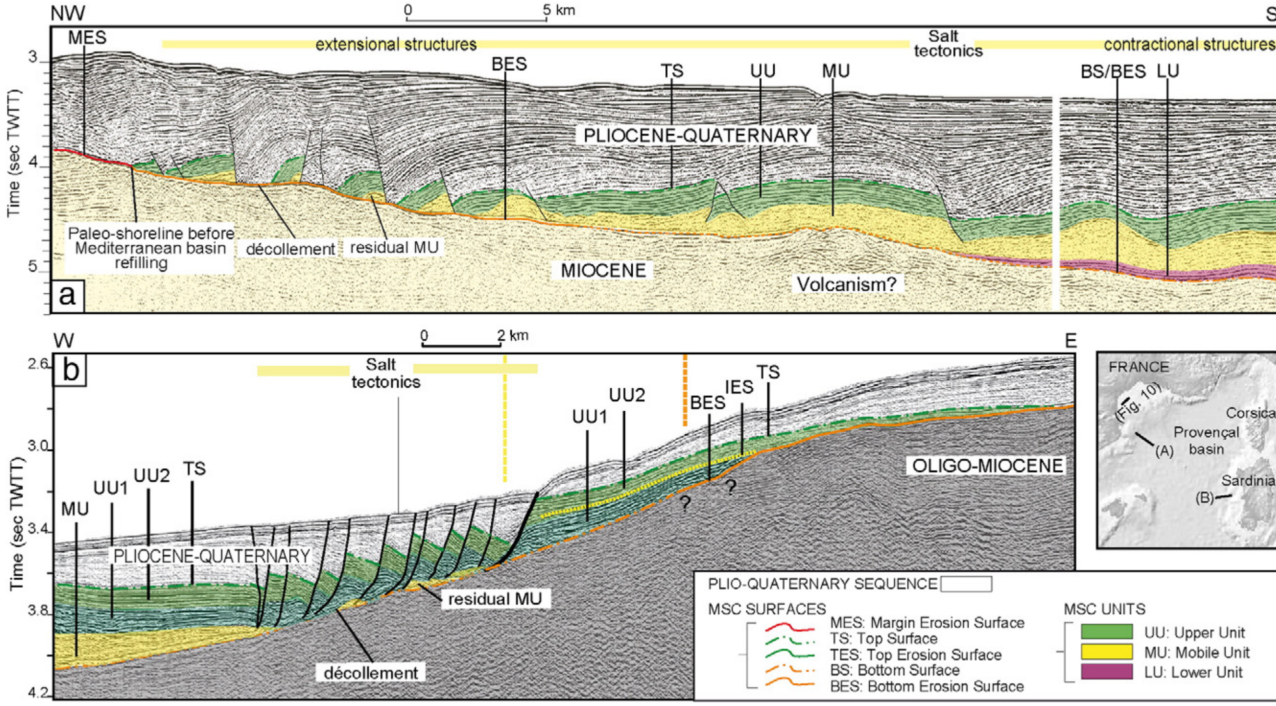

Fig. 16. (a) Seismic profiles along the Gulf of Lions margin lower slope showing the MES extending out downslope to a depth of ~4.0 s TWTT and basin basinward to the deep basin Messinian trilogy (LU, MU, UU) deposited in onlap on the margin slope. (b) Seismic profile along the Western Sardinia margin illustrating the onlap of UU and MU on the margin slope. The LU is not observed on this profile. On both margins, the MES divides into the Top Surface, at the top of UU, and into the BES, extending basinward beneath the UU, MU and possibly LU. The BES and TES pass basinward to conformable surfaces bounding the MSC trilogy and are labeled TS and BS respectively. Above the MSC trilogy, the Plio-Quaternary sequence is faulted due to salt tectonics. The most landward fault marks the initial pinch-out of MU, before salt retreat.
Panel a is modified from Lofi et al. (2005, 2011a. Panel b is modified from Sage et al. (2005) and Sage and Déverchère (2011).

et al., 2011; in the Levant basin (Ryan and Cita, 1978; Bertoni and Cartwright, 2007a; Figs. 17 and 18; and on the Cyprus Arc (Maillard et al., 2011), this surface shows evidence of erosion (e.g. gullied morphology or erosional truncations) and is thus labeled Top Erosion Surface (TES) and is generally interpreted as resulting from a phase of subaerial erosion during the last stage of the crisis (Maillard et al., 2006; Bertoni and Cartwright, 2007a; Lofi et al., 2011b).

3.3.2.4. Internal Erosional Surfaces (IES). Locally some intermediate unconformities are observed within the MSC deposits (Sage and Déverchère, 2011; Thinon et al., 2011; Bowman, 2012). They are labeled IES and show locally gullied morphologies, such as in the Valencia (Maillard et al., 2006) and in the East Corsica basins (Thinon et al., 2004, 2011) these internal erosion surface display gullied morphologies.

3.3.3. MSC depositional units

The seismic nomenclature for the MSC units is based on the seismic facies of the units and/or their geometrical relationships with the Mobile Unit (MU). Thus, the Lower Unit (LU) and Upper Unit (UU) underlie and overlie the MU respectively. In the deep Western Mediterranean Basin, these three distinct units, which are known as the "Messinian trilogy" (Montadert et al., 1970), form an aggrading sequence that onlaps

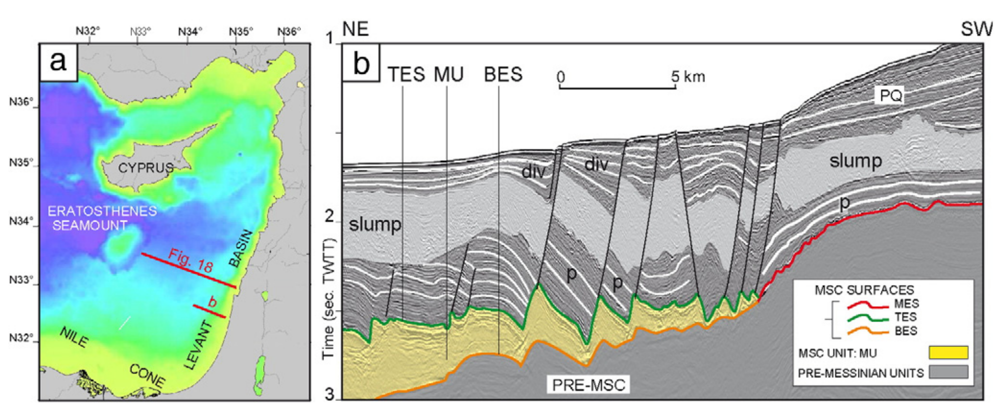

Fig. 17. (a) Map of the Levant basin (courtesy: John K. Hall). The black lines indicate the position of the seismic lines. (b) Time-migrated seismic section from the Levant coast. Extensional thin-skinned salt tectonics led to formation of salt rollers. The Israeli Slump Complex (Martinez et al., 2005) separates the supra-salt strata into a parallel-layered pre-kinematic unit beneath (p) and a syn-kinematic units characterized by divergent reflections (div). The listric faults pierce the seafloor proving active tectonics.
Panel b is after Hübscher and Netzeband (2007) and Hübscher and Dümmong (2011).

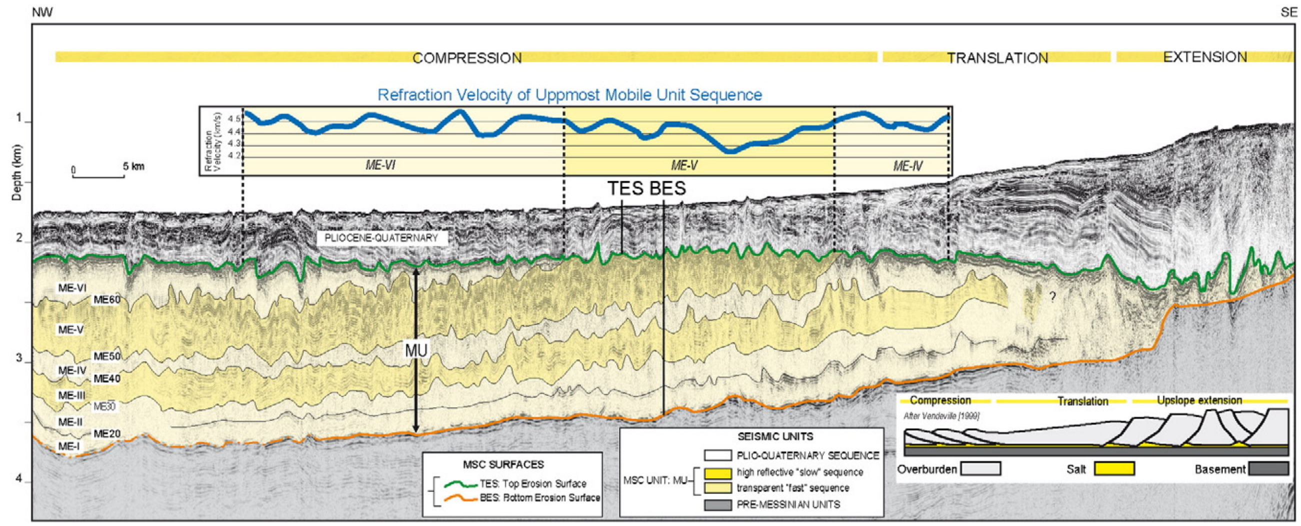

Fig. 18. Depth-migrated seismic section in the Levant basin showing the MU containing several reflector packages exhibiting intra-evaporitic deformation. The thrust angle reach 14°. Refraction velocity of upmost mobile unit sequence shows that some of the units within the MU have lower velocities than expected from pure Halite. See profile location in Fig. 17. After Netzeband et al. (2006) and Hübscher and Dümmong (2011).

the margins and infills the topographic lows (Figs. 1 and 14). Only the Mobile Unit is clearly observed in the Eastern Basin and it contains distinct internal sub-units (Fig. 14). Other seismic units (BU — Bedded Unit and CU — Complex Unit) are observed only locally, the latter generally in association with MSC drainage outlets (Fig. 15).

3.3.3.1. Lower Unit (LU). LU is the lowest MSC seismic unit. It is clearly identified in the Provençal Basin (Western Mediterranean) onlapping the Miocene margin (Fig. 16a) and comprises a group of continuous high-amplitude reflections. These are well imaged in the Gulf of Lions (Montadert et al., 1970; Lofi et al., 2005) where they are partly correlated laterally with CU. The extent and thickness of LU throughout the Western Basin is still poorly constrained. LU has never been drilled and its lithology and age remain unknown. Based on their observations in the Gulf of Lions, however Lofi et al. (2005) proposed that these deposits were up to ~450 to ~600 m thick (using an interval velocity of 3.5 km/s m) and that turbidites and debris flows may partly account for the observed parallel reflectors. In terms of depositional processes (and maybe timing), LU could be an equivalent of the onshore RLG unit (see Section 3.2.3.3; Roveri et al., 2001; Manzi et al., 2005) and/or partly of the PLG unit (De Lange and Krijgsman, 2010). In the Gulf of Lions, Bache et al. (2009) suggested that an additional ~1 km thick unit below the LU (LU0) is also related to MSC events.

3.3.3.2. Mobile Unit (MU). This unit corresponds to the Messinian salt, onlapping the Miocene margins (Figs. 14, 16 and 17). MU is typically the easiest Messinian unit to recognize on seismic profiles due to its characteristic transparent acoustic facies which is thought to result from a succession dominated by halite (Nely, 1994). Plastic deformational structures and associated listric faults and diapirs in the overlying brittle sediments (Gaullier and Bellaiche, 1996; Gaullier et al., 2000; Dos Reis et al., 2005; Loncke et al., 2006) are also typical. MU has been observed in the Western (Provençal Basin, offshore Algeria), Central (western Tyrrhenian Basin, Ionian Basin) and Eastern (Levant Basin, Cyprus Arc, Cilicia–Adana Basin, Mediterranean Ridge) Mediterranean basins. In the Western Basin, apparently undeformed MU is <~1.2 km thick (using a 4.5 km/s velocity) when non-deformed. In the Eastern Basin, MU is thicker (up to ~2.1 km) assuming a 4.2 km/s velocity (Netzeband et al., 2006) (Fig. 18). Some authors have suggested that MUs relatively reflection-free seismic facies results from a very high rate of halite precipitation and/or that salt accumulation started at an early stage, prior to the low-stand phase (Lofi et al., 2011b and references therein). MU locally contains some clastics, although in lesser quantities than the CU, LU and UU. In the Levant Basin the MU contains several discontinuous reflection packages apparently separating six sub-units (Bertoni and Cartwright, 2005, 2006; Netzeband et al., 2006; Hübscher et al., 2007; Dümmong and Hübscher, 2011; Gvirtzman et al., 2013; Figs. 17, 18). Four sub-units are seismically transparent and are presumed to be mainly halite (Bertoni and Cartwright, 2006; Hübscher and Netzeband, 2007). The uppermost transparent sub-unit ME-VI has refraction velocities between 4.4 and 4.6 km/s which is consistent with this interpretation (Dümmong and Hübscher, 2011; Fig. 18). The other sub-units are characterized by lower internal velocities that could reflect vertical variation in the salt facies (gypsum or carnallite) or intercalated layers of limestone, clastics and/or trapped fluids (Mart and Ben-Gai, 1982; Bertoni and Cartwright, 2006; Hübscher et al., 2009).

3.3.3.3. Upper Unit (UU). The UU forms the uppermost unit of the Western Mediterranean MSC trilogy (Fig. 14) and has also been observed in the Tyrrhenian (Lymer et al., 2013; Gaullier et al., 2014) and, possibly, in the Ionian basins (Gallais et al., 2012). Its thickness is up to ~500–900 m (using an interval velocity of 3.5 km/s), with an aggrading geometry that onlaps the margin (Fig. 14). This geometry (Figs. 14 and 16) is often interpreted as reflecting the shoaling of the basin floor when the base level was drastically lowered (Sage et al., 2005; Maillard et al., 2006; Lofi et al., 2011b). UU consists of a group of parallel and fairly continuous reflections of relatively high amplitude (Maillard et al., 2006), generally bounded by a conformable Top Surface (TS). Internal layering becomes rougher when approaching the margin, especially at canyon mouths, reflecting an increase in clastic content (e.g. Provençal, Ligurian and Sardinian margins (Sage et al., 2005; Obone Zue Obame et al., 2011; Sage and Déverchère, 2011). UU also displays an important internal vertical variability, and two sub-units can be locally distinguished (e.g. Fig. 16B; Sage et al., 2005). In the Valencia basin, UU shows internal erosion surfaces (Maillard et al., 2006; Lofi et al., 2011b) and erosion at the top (TES) (Ryan, 1978; Escutia and Maldonado, 1992; Maillard et al., 2006). DSDP-ODP drillings show that the topmost part of UU locally consists of a few tens of meter-thick terrigenous deposits

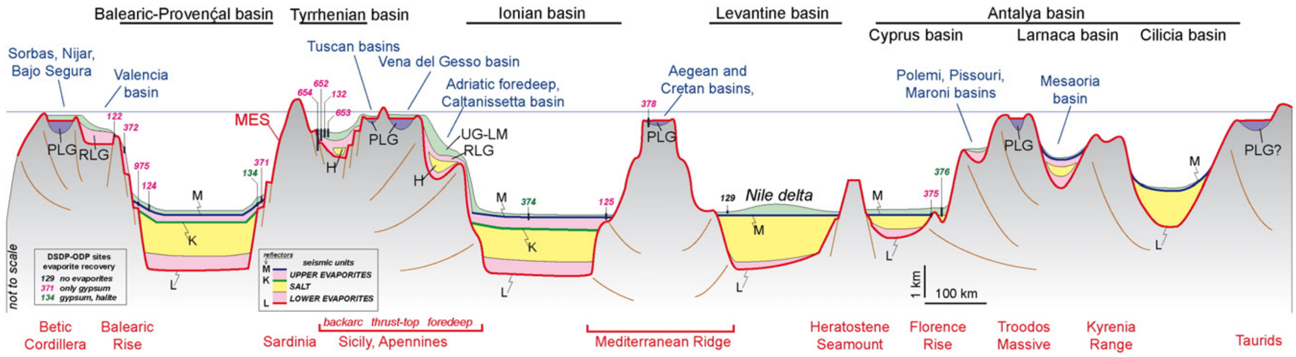

Fig. 19. Distribution of the Messinian deposits and seismic units in the Mediterranean area along a schematic W–E cross-section. The location of the main onshore outcrops and of the DSDP-ODP drilling sites are also reported; in pink the sites where only sulfate evaporites have been recovered; in green the sites where halite has been recovered.
From Roveri et al. (2014).

with rare hypohaline ostracods of Paratethyan affinity overlain by early Pliocene biogenic ooze containing faunal assemblages living in normal, bathyal seawater settings (Cita, 1973). Due to its limited thickness with respect to seismic resolution, this uppermost Messinian unit is seldom distinguished in low-resolution seismic profiles and is commonly combined with the Pliocene unit (Roveri et al., 2014).

3.3.3.4. Complex Unit (CU). Chaotic or roughly bedded, more or less transparent seismic facies (Fig. 15b) have been recognized in several areas (Déverchère et al., 2005; Lofi et al., 2005; Maillard et al., 2006; Bertoni and Cartwright, 2007b; Lofi and Berné, 2008; Bache et al., 2009; Estrada et al., 2011; Lofi et al., 2011a; Loncke et al., 2011; Obone Zue Obame et al., 2011) and labeled as CU (Complex Unit). CU is generally absent on shelves and only rarely observed on the upper part of the continental slope. Thick CU deposits are mainly recovered at the base of the continental slope, either as infill of the MSC paleo-valleys, as fan-shaped deposits (Fig. 15c) or as poorly organized bodies; they often mark the transition to the deep basin evaporites (Déverchère et al., 2005; Lofi et al., 2005; Maillard et al., 2006; Bertoni and Cartwright, 2007b; Lofi and Berné, 2008; Bache et al., 2009; Lofi et al., 2011a; Obone Zue Obame et al., 2011), but the detailed stratigraphic relationships between these units are complex (Rizzini et al., 1978; Barber, 1981; Lofi et al., 2005). The true nature of the CU is still unknown, but according to the main hypotheses, it could consist of both subaqueous and subaerial deposits, resulting from the combined action of several processes including: early subaqueous mass-wasting possibly related to sea-level lowering (Lofi et al., 2005; Bertoni and Cartwright, 2007a), margin steepening due to halite loading (Govers et al., 2009), increasing density inversion between heavier overlying brines and lighter sediment pore waters (Ryan, 2009, 2011), tectonics (Roveri et al., 2001; Manzi et al., 2007); subaerial deposits as a result of lowered sea-level (Lofi et al., 2005).

3.3.3.5. Bedded Unit (BU). Another seismic unit that consists of relatively continuous sub-parallel reflections (Maillard et al., 2006; Guennoc et al., 2011; Thinon et al., 2011) is observed in topographic lows, geometrically disconnected from the other deep basin Messinian units (e.g. East-Corsica basin (Aleria group, 1978; Thinon et al., 2004, 2011) and on the West-Corsica margin (Guennoc et al., 2011). The relative age of these units can therefore not be established and consequently they have been labeled as Bedded Unit deposits (BU; Lofi et al., 2011a). The BU is up to 350 m thick (using a mean internal velocity of 3.5 km/s) and is often bounded above by well expressed Top Erosion Surface (TES), and sometimes below by the BES. In some places, the BU also contains Internal Erosion Surfaces (IES).

3.3.3.6. Eastern vs. Western basin. The seismic records of the MSC differ greatly between the Western and Eastern basins (Fig. 14; Lofi et al. (2011a,b)). It is not possible to correlate the individual units, and in particular the MU, from one basin to the other, because the East and West Mediterranean basins are now separated by the Sicily sill. As a result the synchronous/diachronous nature of MU deposition is not defined yet. This has profound consequences for the evolution of the MSC as a whole since intermediate Mediterranean sills are critical in creating diachronous halite deposits in the Eastern (first) and Western basins (later) (Blanc, 2000; Ryan, 2008), while simple restriction of the Atlantic–Mediterranean connection create coeval Mediterranean-wide halite deposition (Topper and Meijer, 2013).

3.3.4. Deep evaporites: direct and indirect evidence

3.3.4.1. Core data. Although largely incomplete and restricted to the topmost layers of seismic units UU and/or BU, DSDP and ODP cores are the only direct source of information for the offshore deep evaporite rocks (Figs. 1 and 19). The sediments recovered were originally thought to have been in the past the result of deposition in shallow water-supratidal (sabkha) environments during the desiccation of the Mediterranean (Hsü et al., 1973a,b). Hardie and Lowenstein (2004) questioned the original facies interpretation, suggesting that most facies features may result from deep-water (below wave base) deposition. Cored Messinian evaporites can be grouped into three main groups:

1. *Primary facies* consist of selenite, cumulate laminar gypsum and halite cumulate. Massive, banded selenite facies similar to typical onshore PLG deposits (Lugli et al., 2010) have been recovered only from site 378 (Figs. 1, 19). Cumulate laminar gypsum consists of a millimeter-scale rhythmic succession of inversely graded gypsum laminae (e.g. the "balatino" gypsum of Ogniben, 1957). Halite cumulates with clastic carbonate intercalations have been recovered from sites 134 and 374 (Figs. 1, 19).

2. *Clastic evaporitic facies* include gypsrudites, gypsarenites and gypsum siltites consisting of corroded fragments of selenite crystals. Clastic and cumulate evaporites commonly occur together resembling the onshore Resedimented Lower Gypsum facies association (see Section 3.2.3.3). These sediments do not provide a clear indication of depositional depth, but unequivocal evidence of shallow water, sabkha environments or subaerial exposure has not been found.

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3. *Diagenetic evaporites* include nodular, massive, laminar and entherolithic anhydrite. These result from the burial diagenetic transformation (dehydration) of massive and laminar gypsum.

3.3.4.2. Indirect observations for evaporites buried below the deep Mediterranean basins. Several deep hypersaline brine basins thought to originate from dissolution of underlying Messinian evaporites have been discovered in the Mediterranean Ridge accretionary wedge (Eastern Mediterranean, south of Crete, Fig. 1) with research vessels Tyro, 1983; Bannock, 1984; Urania, L'Atalante, and Discovery, all in 1992–1993 (De Lange and Ten Haven, 1983; Jongsma et al., 1983; Scientific Staff of Cruise Bannock, 1984–12, 1985; De Lange et al., 1990a,b; MEDRIFF Consortium, 1995; Vengosh et al., 1998; Van der Wielen et al., 2005). In addition, several mud volcanoes in the eastern Mediterranean have pore fluids with enhanced salinity that has also been attributed to dissolution of underlying Messinian evaporites (De Lange and Brumsack, 1998; Daehlmann and De Lange, 2003). In all but a few sites, Na and Cl are the dominant ions usually occurring in a ratio that is close to 1:1, indicating halite dissolution. It is only in Discovery and the nearby Kryos deep Ionian brine basins that Na is a minor element and Mg and Cl are the dominant ions, suggesting the dissolution of the evaporite mineral bischofite ($MgCl_2 \cdot 6H_2O$). In some of the cores, also K as an ion increases. Taken together, these observations indicate that evaporites underlying deep eastern Mediterranean sediments are not limited to gypsum or halite alone, but do also contain other, late stage evaporite minerals which have not been recovered in existing cores of the UU.

3.4. Geodynamic framework and implications of the MSC

3.4.1. Geodynamic impact of MSC

The configuration of the Mediterranean during the MSC is the result of the progressive convergence of Africa with Arabia–Eurasia in combination with extension driven by roll-back of oceanic slabs beneath the Apennines and the Aegean Sea (Wortel et al., 2006). Precisely dated kinematic data are too sparse to allow for an accurate reconstruction, especially in the central Mediterranean where significant changes occurred since the Miocene (Fig. 20a; Jolivet et al., 2006; Govers et al., 2009; Manzi et al., 2012). Estimates of pre-MSC paleobathymetry are strongly coupled to the paleogeographic reconstruction and the unknown thickness of basinal Messinian sequence are a significant source of additional uncertainty.

The width of the surface load relative to the “effective elastic thickness” of the lithosphere controls whether compensation occurs by local or regional isostasy (flexure; Watts, 2001). The Mediterranean basin had relatively small horizontal dimensions during the Messinian and the evaporite/halite load appears to have been distributed over a large part of the basin (Rouchy and Caruso, 2006; Fig. 1). Lithospheric flexure is therefore the most likely mechanism in responding to MSC surface processes (Govers et al., 2009). An important consequence of flexural support is that a surface load drives vertical motions over a much wider region (tens to hundreds of km beyond the edge of the load). The deposition of massive halite results in a very significant load and predicts basement subsidence beneath the load of the continental margins, and peripheral uplift further inland (Govers et al., 2009; Garcia et al., 2011). Predicted change in erosion rate at the margins (positive values in Fig. 20b) help in constraining the thickness of the massive halite. Conversely a sea level drawdown (if it occurred; Roveri et al., 2001; Cosentino et al., 2012) would have caused the uplift of the basin and of its margins, and subsidence further inland.

The feedback between surface processes and isostasy becomes even more relevant when the gateways that connected the Atlantic and the Mediterranean are considered. Lowering of Mediterranean sea level must have resulted in isostatic uplift of the edges of the Alboran basin including the seaway(s) (Govers, 2009). Potentially, this would further diminish the water supply from the Atlantic and lead to further sea level lowering of the Mediterranean. The restoration of Mediterranean sea level in the Pliocene neutralized the isostatic uplift of the edges within a few thousand years. If there existed a seaway through the Rif and/or the Betics until the Pliocene, it probably was choked by the isostatic uplift during the Lago Mare stage of the MSC and some additional tectonic process would be required to force the reflooding of the Mediterranean. In the context of plate convergence, a possible driver for this opening would be the dynamics of the Gibraltar slab (Duggen et al., 2004; Govers, 2009). If the Strait of Gibraltar region was leaky before the Pliocene however (Hsü et al., 1973a,b), there is a distinct possibility that downcutting/erosion was capable of keeping up with the isostatic uplift following drawdown (Garcia-Castellanos and Villaseñor, 2011). In this scenario, the erosion rate (sensu lato) at Gibraltar Strait is the prime regulator of Mediterranean sea level, and there is little need for additional tectonic processes to explain the Zanclean flood. Seaways within the Mediterranean–Paratethyan realm underwent similar uplift and subsidence in response to surface processes (Bartol and Govers, 2009). With its multiple shallow gateways between sub-basins, the connectivity of Paratethys realm became particularly sensitive to sea level variations (Leever et al., 2011; Csato et al., 2013; Vasiliev et al., 2013).

3.4.2. Atlantic gateway dynamics

The configuration of the Mediterranean–Atlantic gateway is a major control on exchange of both water and salt and has consequences for circulation in both basins (e.g. Reid, 1978). Today, the Mediterranean loses more fresh water by evaporation than it receives from precipitation and fluvial run-off (e.g. Bryden et al., 1994). This freshwater deficit generates a salinity contrast between the Atlantic and Mediterranean (Boyer et al., 2009) which results in a bottom current of warm, saline Mediterranean Outflow Water flowing over the sill of Gibraltar while less saline Atlantic water is drawn in at the surface.

In pre-MSC times, two marine gateway systems connected the Mediterranean and Atlantic (e.g. Benson et al., 1991): the Iberian or Betic Portal (southern Spain), and the Rifian Corridor (northern Morocco). Most of the Late Miocene gateway sediments that are now exposed on land have been intensively studied, but the timing of corridor closure is still subject to significant uncertainty (Benson et al., 1991; Martin and Braga, 1994; Krijgsman et al., 1999a; Martin et al., 2001; van Assen et al., 2006; Hüsing et al., 2010, 2012). This is largely because the area which first blocks the marine connection is, by default, the area that experiences the longest uplift and erosion and the sedimentary succession preserved is therefore always incomplete. Most studies of these unconformity-bearing successions, however, suggest that the gateways were closed well before the onset of the MSC (Benson et al., 1991; Kouwenhoven et al., 1999; Martin et al., 2001). The Betic Corridor is thought to have been shut first around 6.3 Ma (Martin et al., 2001), while the southern strand of the Rifian Corridor became emergent sometime before 6.0 Ma (Krijgsman et al., 1999a). This pre-MSC cessation of flow through Morocco is consistent with and refined by several other datasets from the region, e.g. mammal exchange between Africa and Europe (before 6.1 Ma; Agustí et al., 2006); sedimentological changes elsewhere in the corridor (6.58–6 Ma; van Assen et al., 2006) and Nd-isotope records tracking Mediterranean outflow to the Atlantic (Ivanovic et al., 2013a). This result is problematic because the precipitation of thick evaporite deposits in the Mediterranean during the MSC requires a significant supply of salty water. Modeling studies suggest that only a very narrow (~1 km) and shallow (~10 m) connection is required (Meijer and Krijgsman, 2005), but from the perspective of linking gateway evolution with the development of the MSC succession, identifying the location and geometry for this gateway is critical.

3.4.3. Mediterranean–Paratethys connectivity

The presence of many fossils of Paratethyan affinity in the latest Messinian Mediterranean sequences (Suc et al., 1999; Gliozzi et al., 2002; Bertini, 2006; Grossi et al., 2008; Popescu et al., 2009) supports the hypothesis that Paratethys fed brackish water to the Mediterranean

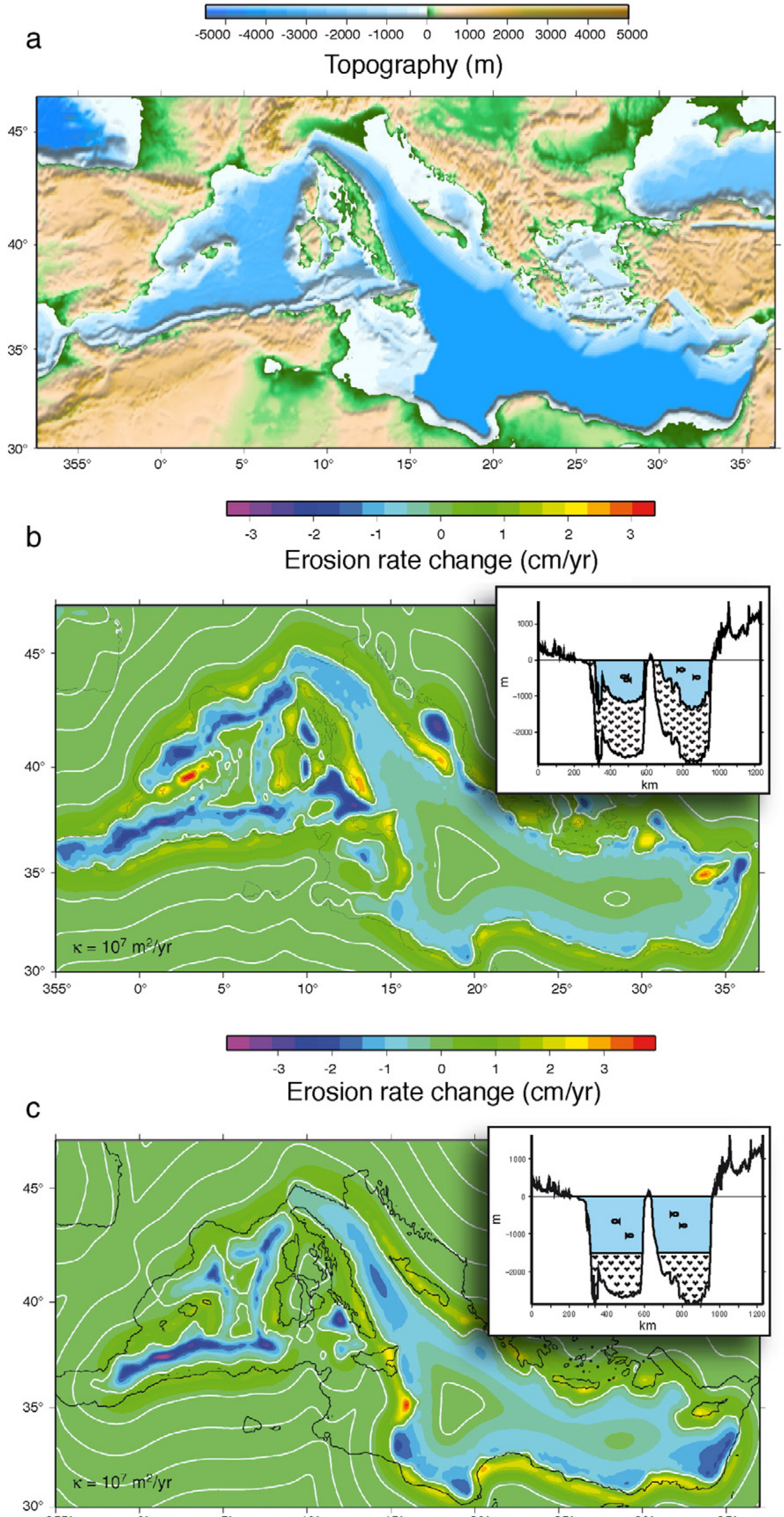

Fig. 20. (a) Reconstructed geometry and bathymetry of the Mediterranean region at the onset of the MSC. Relative to the present-day situation, Italy and Calabria were not yet in place and Mesozoic oceanic lithosphere still needed to be subducted beneath the Apennines (Di Stefano et al., 2009). In the eastern Mediterranean, the Aegean region was smaller and the Mediterranean Ridge was relatively thin. (b) Change in erosion rate in response to two alternative scenarios for loading of the Mediterranean basement by a layer of Lower Evaporites and massive halite. The top panel shows the consequences for a uniform 1500 meter thick layer. The bottom panel shows the result for a layer which is deposited 1500 m below sea level. This layer is more than 1500 meter thick in the deepest part of the basin. Predictions are markedly different for the Balearic Islands, which are not very much affected by tectonic motions during this period. Panel a is modified from Govers et al. (2009). Panel b is modified after Govers et al. (2009).

during MSC Stage 3, resulting in the establishment of Lago-Mare facies all over the Mediterranean Basin (Cita et al., 1978; Orszag-Sperber, 2006; Roveri et al., 2008b). The location of such a marine connection, however, has, so far, not been unequivocally demonstrated (Popov et al., 2006; Krijgsman et al., 2010). Inadequate stratigraphic correlations and insufficiently robust age control on Paratethyan successions have long hampered a thorough understanding of Paratethys–Mediterranean exchange during the MSC. Paratethyan time scales comprise regional stages (e.g. Maeotian–Pontian–Dacian or Kimmerian within the Mio-Pliocene interval; Fig. 3) and substages (e.g. Odessian–Portaferrian–Bosphorian within the Pontian; Fig. 3) based on endemic ostracod and mollusc species. In the last decade, numerous integrated magnetobiostratigraphic studies have been performed on the Messinian sedimentary successions, which have resulted in a revised chronological framework for Eastern Paratethys (Vasiliev et al., 2004, 2011; Stoica et al., 2007, 2013; Krijgsman et al., 2010). This framework allows high-resolution stratigraphic correlations between the individual Paratethyan subbasins and the Mediterranean (Fig. 3).

The Messinian pre-evaporite successions correspond in time to the upper Maeotian deposits of the Eastern Paratethys (Fig. 3). The onset of the MSC in the Mediterranean slightly post-dates the Maeotian–Pontian transition, which is magnetostratigraphically dated at 6.04 ± 0.01 Ma (Krijgsman et al., 2010; Vasiliev et al., 2011). An interval containing marine foraminifera occurs at the base of the Pontian suggesting that a marine flooding event took place in Paratethys, probably as a result of a connection with the Mediterranean (Stoica et al., 2013). The lowermost Pontian (Odessian/Novorossian) substage, which mainly correlates with the PLG unit of the MSC (Fig. 3), is a base level highstand in the Paratethys. Interbasinal and Mediterranean connections are indicated by the presence of marine alkenones in the Black Sea basin (Vasiliev et al., 2013). A marked change in Paratethyan ostracod faunas, suggesting re-stabilization of palaeoenvironmental conditions, coincides with the onset of MSC evaporites in the Mediterranean (Stoica et al., 2013). This suggests that changes in Paratethys–Mediterranean connectivity may have generated the conditions for gypsum to precipitate in the Mediterranean basin (Krijgsman et al., 2010).

The MSC acme interval corresponds to the base of the Black Sea's Kimmerian Stage and to the mid-Portaferrian in the Dacian Basin of Romania (Stoica et al., 2013; Fig. 3). A subsequent fall in Paratethyan water level corresponds in age to the glacial cycles TG12–14 (5.59–5.52 Ma) and therefore coincides closely with the Mediterranean isolation-event of MSC stage 2 (Fig. 3). The extent of this sea level fall is still vigorously debated, with figures ranging from only a relatively minor (~50 m) drop to the level of the paleo-Bosphorus (Popov et al., 2006; Krijgsman et al., 2010) to large-scale desiccation (>1000 m) of the entire Black Sea (Hsü and Giovanoli, 1979; Gillet et al., 2007; Popescu, 2010; Munteanu et al., 2012). A major climatic change towards more humid conditions is likely to have produced a more positive hydrologic balance which may have generated a second transgression that occurred in the Paratethys during MSC stage 3 (Bosphorian), although its relationship with the Pliocene flooding of the Mediterranean cannot be completely ruled out (Krijgsman et al., 2010).

3.5. The impact of MSC on marine biota

The progressive closure of its Atlantic connections changed the physical and chemical properties of Mediterranean water and consequently had a profound impact on its ecosystems (see also Section 3.2.3.1). Stronger water column stratification at around 6.7 Ma (Sierro et al., 2003; Van Assen et al., 2006) promoted significant changes to the shallow-water macrobenthic assemblages. Coral reef and reef complexes dominated by *Porites* developed all over the Western Mediterranean, especially around the Gibraltar Arc in SE Spain, the Balearic Islands and the northern margin of Morocco and Algeria, as well as in northern and southern Italy (Esteban, 1979; Dabrio et al., 1981; Braga and Martin, 1996; Perrin and Bosellini, 2012). Coral diversity was not very different from that of the Tortonian because other genera, such as *Tabellastraea* and *Siderastrea*, vermetids and serpulids as well as algal bioherms were also present

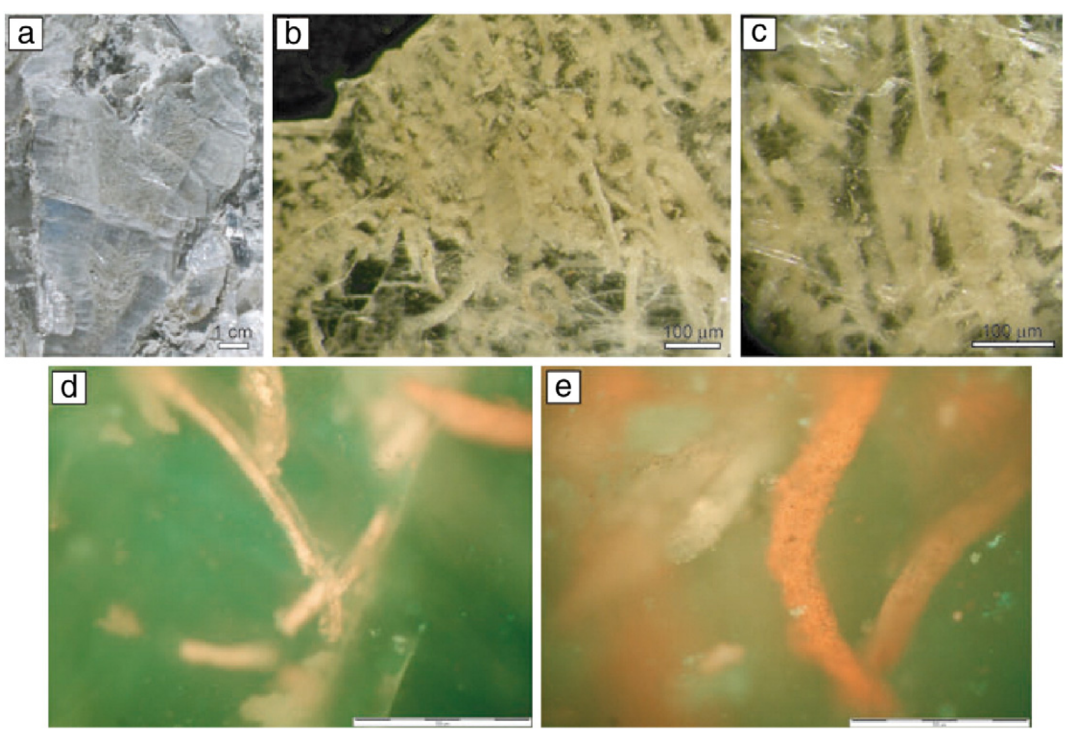

Fig. 21. The bacterial signature in the Primary Lower Gypsum evaporites. (a) 'Arrow-head' selenite crystal characterized by a dense fossil filaments in the inner dark core. Monte Tondo quarry, Vena del Gesso basin, PLG cycle 3. (b, c) Photomicrograph of cleavage fragments of a selenite crystal showing single, unbranched filaments, appearing as whitish tubes causing a cloudy appearance in the crystals (natural, transmitted light). (d, e) Photomicrograph of microbial filaments preserved in a gypsum crystals under UV light; chlorophyll presence in the filaments is revealed by the red autofluorescence. Gypsum crystals without visible filaments did not show any fluorescence under the UV light.
From Panieri et al. (2010).

Please cite this article as: Roveri, M., et al., The Messinian Salinity Crisis: Past and future of a great challenge for marine sciences, Marine Geology (2014), http://dx.doi.org/10.1016/j.margeo.2014.02.002

(Saint Martin et al., 2007; Guido et al., 2012). These reef complexes are thought to have developed at times of Northern Hemisphere summer insolation maxima when sea-surface temperatures (SST) remained warm throughout the year and warm-subtropical calcareous planktonic assemblages were abundant in the open sea (Sierro et al., 1999; Sanchez-Almazo et al., 2007).

The first evidence of extreme conditions at sea surface is observed at around 6.3 Ma (Sierro et al., 1999), when planktic foraminifers began to disappear during precession maxima. A similar pattern in calcareous nannofossils followed, each group of microplanktic organisms responding to the significant hydrologic deficit (Blanc-Valleron et al., 2002; Sierro et al., 2003; Manzi et al., 2007).

After the onset of evaporite deposition only some *Porites* patches coexisting with oolitic and non-oolitic stromatolites and thrombolites survived ("Terminal Carbonate Complex", see Section 3.2.3.5; Esteban, 1979; Riding et al., 1991; Braga et al., 1995; Feldmann and McKenzie, 1997; Cornée et al., 2004; Roveri et al., 2009; Arenas and Pomar, 2010; Bourillot et al., 2010). Fossil cyanobacteria and ribosomal RNA gene fragments have been reported in gypsum crystals of the Lower Gypsum (Panieri et al., 2010; Fig. 21). Microbialites with sulfide-oxidizing bacteria have been observed in anoxic deposits representing the PLG deep-water equivalent (Dela Pierre et al., 2012). Typically, at the onset of evaporite deposition, while calcareous plankton assemblages in marginal and deeper basins were either significantly in diversity or completely absent, suboxic or completely anoxic waters hampered the survival of deep-water benthic organisms (Manzi et al., 2007; Dela Pierre et al., 2012). These anoxic conditions and lack of benthic organisms also persisted after MSC stage 1 (Sampalmieri et al., 2010).

In the pelites cyclically interbedded with evaporites of Primary Lower Gypsum and Upper Gypsum, planktonic foraminifers are either absent or present only as dwarf specimens dominated by *Turborotalita quinqueloba*, *Turborotalita multiloba* and small globigerinids, suggesting very stressful conditions at surface. The occurrence of a diverse assemblage of calcareous planktonic and benthonic microfauna in some of these pelites as well as in deep-sea records is very controversial as it has been interpreted by some authors as a proof of the incursion of normal open marine conditions (Aguirre and Sánchez-Almazo, 2004; Braga et al., 2006) or the result of reworking (Cita et al., 1978, 1990; Iaccarino and Bossio, 1999; Bassetti et al., 2006). Fish fossils found in the pelites interbedded with the PLG in Italy suggest the existence of normal marine conditions in the open sea (Carnevale et al., 2008).

Information about marine fauna during halite deposition is scarce (e.g. Bertini et al., 1998). A major change to hypohaline fauna is observed between cycles 4 and 5 of the Upper Gypsum (Lago-Mare) unit (Rouchy and Caruso, 2006; Roveri et al., 2006; Manzi et al., 2009) on Sicily. In those marginal settings where the Upper Gypsum was not deposited, the post-evaporitic units are typically characterized by freshwater-oligohaline to mesohaline benthic organisms, the so-called "Lago-Mare" Paratethyan fauna (see Section 3.2.3.4). These freshwater species have also been found in the Upper Unit at deep-sea sites such as in ODP sites 654 (Cita et al., 1978, 1990), 968 (Blanc-Valleron et al., 1998), 974, 975 and 978 (Iaccarino and Bossio, 1999) and in DSDP sites 375 and 376 (Fig. 1).

3.6. Paleoclimate and paleoceanographic reconstructions

3.6.1. Peri-Mediterranean climate

Climate proxy data from both the terrestrial (e.g. Pound et al., 2011, 2012) and oceanic (e.g. Zachos et al., 2001, 2008) realms suggest that, globally, the Late Miocene was warmer and wetter than it is today. Regrettably, the dataset which provides quantitative estimates of these differences is not evenly distributed and the terrestrial dataset, in particular, is strongly spatially biased towards Europe and Asia (Bradshaw et al., 2012). Nonetheless, these proxy data suggest that climatic difference from the present day was greater for the Tortonian period than it was for the Messinian, particularly with respect to palaeoprecipitation (Bradshaw et al., 2012). Palaeo-CO_2 reconstructions are available from a variety of proxies for this period. There is considerable uncertainty about the absolute concentration during the Late Miocene, but estimates are generally within error of pre-industrial levels (280 ppm). Climate modeling of this period typically struggles to simulate the warm, wet conditions indicated by the climatic proxy data with such low CO_2 concentrations (e.g. Steppuhn et al., 2006; Micheels et al., 2007) and Bradshaw et al. (2012) concluded that late Miocene atmospheric CO_2 concentrations are likely to have been near the high end of the range of reconstructions e.g. ~350 ppm.

Palaeoclimatic reconstructions based on the Late Miocene Mediterranean successions (e.g. Suc and Bessais, 1990; Suc et al., 1995b; Bertini et al., 1998; Suc et al., 1999; Bertini, 2006; Fauquette et al., 2006, 2007; Iaccarino et al., 2008; Menichetti, 2011; Jimenez-Moreno et al., 2013) document longitudinal and latitudinal climatic gradients over the Mediterranean, as well as significant local variability. However, the MSC is not marked by the dramatic fluctuations seen in sediments younger than 2.6 Ma during glacial–interglacial cycles (Bertini, 2010).

Palynological data suggests that prior to the MSC, wet, subtropical to temperate forest taxa dominated the north of the region, whereas to the south, herbaceous taxa typical of dry open environments expanded (Suc et al., 1995a; Bertini, 2006; Bertini and Martinetto, 2008, 2011; Gennari et al., 2013). Sedimentary cyclicity (sapropel–marl–diatomite) is reflected in the pollen records (Suc et al., 1995a; Iaccarino et al., 2008; Menichetti, 2011), but there is no major change from moist to dry conditions at the onset of evaporite precipitation and consequently, alteration in the climate is not considered to be a major trigger for the MSC (Suc and Bessais, 1990; Suc et al., 1995a,b; Bertini et al., 1998; Bertini, 2006). This broad vegetation pattern persisted during MSC stage 2, but the presence of sub-arid as well as some tropical taxa indicates that climatic conditions in the south were similar to those around the Red Sea today (Bertini et al., 1998). In the lower portion of stage 3, an expansion of open vegetation is indicated by the spread of subdesertic plants, e.g. *Lygeum*, which reach the latitude of Ancona (Marche Region, central Italy; Bertini, 2006). During the Zanclean, the southwest Mediterranean area had an open subdesertic landscape while west equatorial Africa was covered with tropical forest and the vegetation of western Europe and the northwest Mediterranean was largely subtropical forest.

Although the climate shows no major variations during the MSC, the existence of large-scale rivers draining north Africa west of the Nile (Griffin, 2002, 2006), is likely to have resulted in a significantly different hydrologic budget in the Mediterranean (Gladstone et al., 2007). Large-scale fluvial features crossing northern Chad and Libya, linked the Chad Basin to the south with a prominent fluvial incision preserved near the Gulf of Sirt on the Libyan coast (Griffin, 2002). During the Quaternary, when the precipitation–evaporation relationship permitted, a lake formed in the Chad Basin (Pachur and Altman, 1997) and Griffin suggested that something similar occurred in the Neogene coincident with subsidence in Chad Basin area (Genik, 1993). This idea is consistent with modeling studies, which show migration of the Inter Tropical Convergence Zone, such that Lake Chad would have experienced far greater precipitation as a result of monsoonal activity (Gladstone et al., 2007). Consequently, it is likely that the Mediterranean received considerably more fresh water during the Miocene than today, principally through the north Africa drainage system, which is dry today (Gladstone et al., 2007).

3.6.2. Modeling palaeoclimate and palaeoceanography

Faced with a geological problem as poorly constrained by observations as the MSC, the application of theory and modeling provides a valuable additional means of gaining insight, exploiting the laws of physics and chemistry. This theoretical approach was pioneered by Debenedetti (1982), Jauzein and Hubert (1984) and Sonnenfeld (1984). More recently, Blanc (2000, 2006) quantitatively analyzed scenarios proposed for the entire MSC.

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The specific stages of the MSC and the processes thought to dominate them are also amenable to this approach. Meijer and Krijgsman (2005) and Meijer (2006; see also Krijgsman and Meijer, 2008) focussed on, respectively, the desiccation and refilling of the basin and the situation of (partially) restricted outflow to — but continuous inflow from — the Atlantic Ocean. These relatively simple box model calculations indicate, for example, that the Mediterranean basin responds to restrictions in the Atlantic gateway on the time scale of 3–5 ka and they confirm earlier suggestions that, in order to have gypsum deposited but stay below halite saturation, some outflow must have persisted.

Further constraints on the hydrological budget were provided by combining indicators for paleo-salinity with Sr isotope data (Flecker et al., 2002; Flecker and Ellam, 2006; Fig. 13a). Because of the peculiarities of the bedrock of the circum-Mediterranean drainage areas, rivers flowing into the sea are characterized by an $^{87}Sr/^{86}Sr$ ratio that is lower than that of the Atlantic Ocean. When the oceanic inflow is sufficiently restricted so that it does not dominate the Mediterranean water signal, the Sr isotope ratio offers an additional constraint on the relative proportions of ocean and river waters. Flecker et al. (2002) concluded that the onset of evaporite precipitation required an influx of additional salty water, which they interpreted as a transgression. The combined salinity-Sr isotope approach was developed into a box model by Topper et al. (2011, Fig. 13b), which suggested that the onset of the MSC can be explained by restricting Atlantic exchange alone and the change from Lower Evaporites into halite may have involved an additional increase in river discharge.

Disregarding Sr-isotopes but introducing the predecessor of the Strait of Sicily and allowing for the possibility of seawater stratification, Topper and Meijer (2013) addressed the conditions responsible for the similarities and differences in the evaporite record of the western and eastern Mediterranean basin and of the marginal versus the deep basins. A counterintuitive and therefore important result of the calculations is that differences in halite thickness between the western and eastern sub-basin can be explained without invoking a significant restriction of the proto Strait of Sicily.

In the models discussed so far, the water exchange across the strait is parametrized in terms of an efficiency coefficient. This leaves unanswered the fundamental question as to how wide and deep the gateway was during the different stages of the MSC. To supply the missing link between strait cross-sectional geometry and throughflow, Rohling et al. (2008) turned to the theory of hydraulic control in sea straits. Thus being able to calculate basin salinity as a function of strait depth explicitly, they show that sea-level variations might explain the alternation of gypsum and non-evaporitic marls of the PLG. The application of hydraulic-control theory was explored more extensively by Meijer (2012) also considering the transient response of the basin. The model results indicate that salinity and strait depth are related in a nonlinear way (Fig. 13c): even a slow gradual decrease in sill depth is expected to give an event-like rise in basin salinity. For strait width of several kilometers, strait depth has to be reduced to a few tens of meters to achieve gypsum saturation.

Today, the higher density Mediterranean Overflow Water is thought to contribute to North Atlantic circulation through its influence on Atlantic Meridional Overturning Circulation; Bigg and Wadley, 2001; Artale et al., 2002; Voelker et al., 2006; Rogerson et al., 2010; Penaud et al., 2011; Rogerson et al., 2012). Because this density contrast was enhanced by the salinity extremes of the MSC, it is reasonable to expect some reflection of the changing volume and properties of Mediterranean Overflow Water in Late Miocene records of North Atlantic circulation and climate. Most General Circulation Model experiments that explore the impact of Mediterranean outflow either block Mediterranean–Atlantic exchange (Rahmstorf, 1998; Chan and Motoi, 2003), or specify only small perturbations to modern Mediterranean outflow strength and salinity (e.g. Rahmstorf, 1998; Kahana, 2005). The modeled impact on climate is small. Sensitivity studies using more extreme salinities consistent with Mediterranean evaporite precipitation however, produce significant regional climate anomalies in the North Atlantic, Labrador and Greenland–Iceland–Norwegian Seas (Ivanovic et al., 2013a).

Both atmosphere and vegetation models have also been used to explore other significant MSC drivers. Favre et al. (2007) evidenced a strong vegetation contrast between the Northern (forest vegetation) and Southern (open vegetation) Mediterranean regions, but do not take account of any significant sea level fall associated with the MSC. Murphy et al. (2009) atmosphere only model experiments suggest that, except when salinity is a substantial barrier to evaporation, a partially filled basin cannot exist in equilibrium and the Mediterranean must either be partly connected to the Atlantic Ocean or completely desiccated.

4. Open key questions

Despite considerable progress over the last 15 years and the consensus achieved on some important aspects, several fundamental MSC questions still remain unanswered. The most relevant are discussed briefly here.

4.1. Offshore–onshore correlation

A comprehensive stratigraphic scenario of Messinian events is still unavailable because of the difficulties in combining onshore and offshore data and observations. The fundamental problem is that onshore and offshore successions are everywhere physically disconnected, which makes establishing reliable correlations problematic. The missing information required to make these correlations is the nature of the Late Miocene rocks and sediments now lying buried beneath Pliocene and Quaternary sediments in the deep Mediterranean basins. Scientific drilling in the deep Mediterranean basins could, in principle, provide a complete sedimentary record of the main seismic units found in the Western and Eastern Mediterranean (see Section 5).

Several key surfaces identified in onshore successions have been tentatively correlated offshore. However, identifying the MES unequivocally, even onshore is challenging. For example, some authors place the MES on Sicily either in the uppermost part of the Upper Gypsum, at the base of the Trubi marls (Clauzon et al., 1996) or at the base of the terrigenous unit (Arenazzolo) (Clauzon et al., 2005; Londeix et al., 2007; Bache et al., 2012), which precedes the Messinian/Zanclean boundary, i.e. within stage 3 deposits. In the Sorbas Basin (Spain) the presence of an erosional surface at the base of the Yesares Fm. has long been proposed (Riding et al., 1999; Braga et al., 2006; Soria et al., 2008), although other authors suggest that its development is higher in the stratigraphy at the boundary between the Sorbas and Zorreras members (Fortuin et al., 2000; Krijgsman et al., 2001). Recent studies (Roveri et al., 2009; Lugli et al., 2010; Manzi et al., 2013) have suggested that the MES is actually located at the top of the Yesares Formation.

Tracing the basinward extension of the MES is equally problematic and different stratigraphic positions for it result in strikingly different scenarios for the MSC. For example, it has been suggested that the Messinian Erosional Surface cutting MSC stage 1 and older deposits onshore may be traced offshore to the base of LU or MU (Roveri et al., 2001; Lofi et al., 2005; Fig. 22). Given that the Mio-Pliocene boundary can be rather more confidently identified throughout the basin, these two surfaces imply that the deep basin evaporites and associated deposits belong to MSC stages 2 and 3. However, Bache et al. (2009) proposed that the basinward extension of the MES in the Gulf of Lions (i.e. the Bottom Surface) should be traced to a much lower position. What follows from this is that the pre-Mobile Unit deposits (LU + LU0) reach up to ~1500 m in this area and that the deep basin evaporites accumulated under shallow water conditions. This interpretation is still subject to debate (Lofi et al., 2005; Lofi and Berné, 2008; Bache et al., 2009; Ryan, 2011).

Relating the seismic units to onshore-derived stages is also problematic. Distinguishing between stage 2 and stage 3 deposits in the offshore

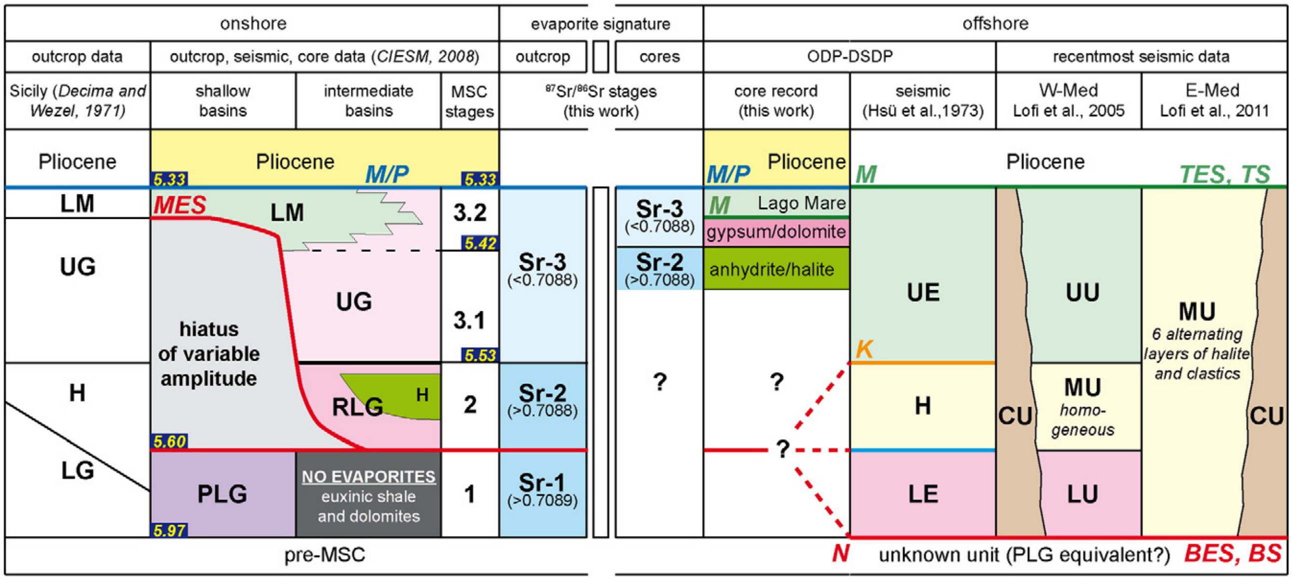

Fig. 22. Synthesis of nomenclature assigned in the literature to onshore and offshore Messinian stratigraphic units. Onshore units: LG, Lower Gypsum; PLG, Primary Lower Gypsum; H, halite; RLG, Resedimented Lower Gypsum; UG, Upper Gypsum; LM, LagoMare; MES, Messinian erosional surface. M/P, Miocene–Pliocene Boundary. MSC stages: 1, 2, 3.1, 3.2. Strontium Isotope stages: Sr-1, Sr-2, Sr-3. Offshore units: LE, Lower Evaporites = LU, Lower Unit; H, Messinian Salt = MU, Mobile Unit; UE, Upper Evaporites = UU, Upper unit; CU, Complex Unit. Offshore surfaces: MES, marginal erosional surface; horizon N = BES, basal erosional surface/BS, basal surface, base of Messinian evaporites; horizon M = TES, Top erosional surface/TS, Top Surface: top of Messinian evaporites.
From Roveri et al. (2014).

record is not currently possible, but where the Messinian trilogy is well developed, it has been suggested that MU and UU units may correspond to stages 2 and 3, respectively (CIESM, 2008). However, on the basis $^{87}Sr/^{86}Sr$ data from the UU unit Roveri et al. (2014) have suggested that most of the UU may actually belong to MSC stage 2 (see figure 10 in Roveri et al., 2014). Entirely different onshore–offshore correlation scenarios are envisaged by other authors who suggest that onshore MSC deposits (i.e. including the PLG unit) are younger than the deep offshore evaporites and formed during the Mediterranean reflooding following the main drawdown (Riding et al., 1998; Braga et al., 2006; Soria et al., 2008). Until data is available that can test these scenarios, these contradictions and controversies will persist.

4.2. Mediterranean base-level changes during stages 2 and 3

One of the most intriguing and enduring issues is the extent to which Mediterranean base-level changed during the MSC. At the heart of this controversy is the interpretation of the water depth in the central basins, which are currently based largely on MSC seismic markers geometry (Lofi et al., 2005; Maillard et al., 2006; Bache et al., 2009; Urgeles et al., 2011; Bache et al., 2012) and again awaits core-derived palaeobathymetric information.

A Mediterranean base-level fall ≥ 1500 m is envisaged in the Western basin (Ryan, 1976; Blanc, 2002; Lofi et al., 2005). This is based on (1) interpreting the shallow water nature of the offshore Upper Unit, inferred from its aggradational geometry onlapping the basin margins and the erosion at its top (Maillard et al., 2006) and (2) the subaerial origin of the widespread erosional features and associated drainage patterns observed along the Mediterranean slopes (Ryan, 1978; Lofi et al., 2005) which have been correlated with the MES onshore (see Section 3.2.3.2). (3) In addition, deep incision of long canyons and valleys such as the Rhone and Nile is thought to have been driven by the adjustment of river profiles to an exceptional base-level fall in the Mediterranean (Chumakov, 1967, 1973; Barber, 1981; Clauzon, 1982). An indirect argument supporting the idea of a high-amplitude drawdown is the development of giant pockmark fields caused by pore-fluid overpressure and large-scale fluid venting, recently discovered at the base or within the Messinian salt unit in the Levant basin (Lazar et al., 2012; Bertoni et al., 2013).

However, based on a different interpretation of some evaporitic facies (see Section 3.3.4.1), the occurrence of high-amplitude base-level changes during the MSC has been questioned by several authors (Busson, 1990; Martinez del Olmo, 1996; Roveri et al., 2001; Roveri et al., 2008a, b,c,d, 2009, in press; Manzi et al., 2005; Hardie and Lowenstein, 2004). In the Eastern basin, lower estimates of base-level fall (800 m) have been proposed for the Levant margin (Druckman et al., 1995; Cartwright and Jackson, 2008). Lugli et al. (2013) pointed out the presence of fully subaqueous clastic evaporites in the infill of the main Messinian canyons (including the Afiq canyon) and suggested that the estimates of sea-level drop previously proposed may be not correct.

The timing of Mediterranean sea-level fall has changed through the time (Ryan, 2009). Following the model of Blanc (2000), Lofi et al. (2005) suggested a two-step base level fall in the Western basin controlled by a Sicily–Tunisia sill. In this model, the first step in the order of 400–600 m caused slope instability and the deposition of gravity-flow deposits which formed a large part of the Lower Unit. The thick evaporites of the Mobile Unit and Upper Unit then formed respectively during and immediately after a second, higher amplitude drop in base-level. Many authors converge on placing the maximum sea-level fall at the end of the deposition of the Mobile Unit in the deep western and eastern Mediterranean basins (Bertoni and Cartwright, 2007a; Ryan, 2009; Lofi et al., 2011b), i.e. *at the end* of MSC stage 2 (Fig. 4). Other authors (i.e. Bache et al., 2009, 2012) consider the whole deep basin evaporitic suite as having been deposited *after* the main sea-level drop, requiring a continuous input of marine waters to an almost desiccated basin.

Exactly when the base-level fall occurs has important implications for the origin and nature of the basinal evaporite units and again various different hypotheses endure. Ryan (2008, 2009) proposed that strong net evaporation concentrated Mediterranean seawater prior to drawdown, resulting in the rapid precipitation of halite during sea level fall. In this scenario, the Mobile Unit would have started accumulating in a relatively deep-water setting and ended its deposition in an almost desiccated basin (Lofi et al., 2011b). The Top Erosion Surface observed in the Levant Basin (Ryan, 1978; Bertoni and Cartwright, 2007a) at the top the Mobile Unit is interpreted by some as a phase of subaerial erosion during the last stage of the crisis (Bertoni and Cartwright, 2007a). Others (Roveri et al., 2001, 2014a,b), suggested that halite accumulated in a fully subaqueous environment.

The rise of the sea/lake-level after the MSC peak is considered to have occurred almost instantaneously (Meijer and Krijgsman, 2005; Garcia-Castellanos et al., 2009), more progressively with large-scale fluctuations related to precession-controlled hydrological changes (Fortuin and Krijgsman, 2003) or in a stepped way (Lofi et al., 2005; Bache et al., 2009; Just et al., 2011; Lofi et al., 2011a,b; Bache et al., 2012). Lofi et al. (2005, 2011b) suggested that the Sicily–Tunisian sill may have generated temporary base-level stillstand in the western Mediterranean during the refilling phase. A stepped base-level rise during the final MSC stage has also been suggested by Just et al. (2011), based on the observation of terraced surfaces at constant depth along Messinian paleoslopes. These feature would record a sea-level stillstand in the western Mediterranean which persisted until the water level in the eastern basin transgressed the Sicily–Tunisia sill. Bache et al. (2009, 2012) consider a moderate initial rise in base-level, associated with the deposition of the deep basin evaporites and the development of a transgressive ravinement surface, followed by a rapid sea-level rise (up to 900 m) resulting from the collapse of the Gibraltar channel. In their model, the rapid sea-level rise occurred well before the Mio-Pliocene boundary, at around 5.46 Ma, i.e. during the Lago-Mare phase. Were this to be the case, the eustatic sea-level rise associated with the latest Messinian deglaciation (see Section 3.2.2) could explain, at least in a part, the transgressive trend observed in the onshore stage 3 deposits, and this would indicate a Mediterranean base-level already rising before the Zanclean (Section 3.2.3.4).

4.3. Evaporite facies and related depositional processes

Understanding the depositional environments of the Mediterranean's evaporite facies is critical to producing a reliable interpretation of the MSC. Unfortunately, due to the lack of modern analogs for most evaporites found in the geological record, a large number of important questions still need to be addressed. They concern the definition of:

a) the littoral and/or continental equivalents of the subaqueous PLG stage 1 selenite unit;
b) the true water depth of the laminar cumulate gypsum facies;
c) the contribution of continental and marine source waters to Mediterranean brines during precipitation of the Primary Lower Gypsum and Upper Gypsum (Flecker and Ellam, 2006; Lugli et al., 2010; Natalicchio et al., 2014);
d) the origin of the hydrological change that triggers the appearance of branching selenite from the 6th cycle of the Primary Lower Gypsum units (Lugli et al., 2010);
e) the detailed correlation of selenite gypsum–shale cycles in the shallow sub-basins with the shale–carbonate cycles at greater depth (Lugli et al., 2010; Dela Pierre et al., 2012);
f) the biological control on evaporite, particularly gypsum precipitation e.g. cyanobacteria (Panieri et al., 2010) and sulfide-oxidizing bacteria (Dela Pierre et al., 2012);
g) the origin of high to very-high frequency cyclicity in the selenite and microscopic growth bands in halite (weekly or even daily wind-driven evaporite cycles?).

4.4. Impact on global climate

An important open question that involves both ocean circulation and climate is whether the MSC did, or did not, have an impact on global climate. An ongoing paradox is the recognition that while the Mediterranean Overflow Water (MOW) influences North Atlantic Circulation today, no clear record of the impact of the MSC on Late Miocene climate has been identified. There are several possible reasons for this:

1. Climate modeling studies that assess the impact of varied MOW are hampered by the crude parameterization of Mediterranean–Atlantic exchange in most General Circulation Models (GCM; e.g. Ivanovic et al., 2013b), and by the lack of constraints on the timing and volume of Late Miocene Mediterranean outflow (Ivanovic et al., 2014). GCM modeling experiments focused on the MSC are therefore currently restricted to sensitivity studies rather than more realistic attempts to simulate Late Miocene climate (Ivanovic et al., 2013a, 2014). This limits their role in identifying locations outside the Mediterranean that should have been sensitive to fluctuating MOW during the MSC. It is possible therefore that evidence for MOW driven climate change exits, but has yet to be recovered.
2. The timescale on which the most extreme fluctuations in MOW salinity reached the Atlantic is much shorter than the duration of the MSC stages. In order to raise salinity in the Mediterranean to concentrations consistent with evaporite precipitation, export of salt to the Atlantic must be reduced significantly (Meijer, 2006). The only time at which highly saline water might have been flushed in quantity into the Atlantic is therefore when Mediterranean–Atlantic exchange was re-established ending a phase of evaporite precipitation. As this should also have resulted in Atlantic inflow compensating for both outflow and evaporative loss, any high salinity Mediterranean brine would have been rapidly diluted driving a corresponding decline in MOW salinity and reducing the Mediterranean–Atlantic density contrast and hence reducing significantly the climate impact this high salinity MOW.
3. Detailed records of when MOW reached the Atlantic during the MSC remain elusive. Isotopic records that monitor MOW reaching the Atlantic are currently too low resolution (e.g. Abouchami et al., 1999; Muiños et al., 2008; Ivanovic et al., 2013a) to link to individual MSC stages let alone specific cycles. This is necessary to test another related open issue which is whether the evaporite–marl cycles represent salinity fluctuations either side of a precipitation threshold or the depletion of a required component of the evaporite composition, e.g. the sulfate component of gypsum (de Lange and Krijgsman, 2010). The difference between these two mechanisms has profound implications for whether or not we should anticipate MOW and a related global climate response.

5. Future directions of MSC research

The study of the Messinian Salinity Crisis and its sedimentary record offers opportunities for fundamental exploration as well as theoretical and methodological development in many geological and non-geological disciplines (e.g. life in extreme environments). Here we suggest some of the “frontier” issues that could be addressed by future MSC research.

5.1. Modern vs ancient interaction: towards sub-Milankovitch time scales

In the near future, Messinian research will require more involvement of scientists working on modern Earth processes and environments. The solution to several of the current open MSC questions requires a much deeper knowledge of modern extreme environments and conditions which undergo large amplitude areal or seasonal physio-chemical fluctuations.

To achieve this, we need to improve the stratigraphic resolution of MSC successions. Astrocyclostratigraphy produced a significant step forward in understanding Messinian events by allowing the comparison of stratigraphic records across the Mediterranean basins on a precessional time-scale. However, many apparent paradoxes persist (see Section 4) and it is possible that some of these are related to high-amplitude palaeoenvironmental changes occurring on much shorter, annual to interannual scales. Achieving this degree of stratigraphic resolution is a pre-requisite both to investigate these issues and to establish common ground with scientists working on modern processes. It would also have important implications for palaeoclimatic and palaeoceanographic modeling.

Three fields that would benefit from input from those studying modern systems are:

1. *Evaporite systems*. Saline giants like the MSC have developed episodically across Earth history and for this reason the combination of geodynamic and climatic factors controlling their formation cannot necessarily be found in modern settings. For example, the large selenite crystals typical of the Primary Lower Gypsum do not form in modern natural environments, or even in man-made salt works. The integration of modern (natural and man-made) and late Quaternary evaporite systems, with geological evaporites may provide mutually beneficial insights into environments controls.
2. *Ecology of inland seas.* Some palaeobiological paradoxes characterizing the last MSC stage would also benefit from modern insights. The apparently contemporaneous occurrence of marine fish and meso- to oligohaline molluscs, ostracods and dinocysts during the Lago-Mare event is a good example. A deeper knowledge of environments that experience large salinity fluctuations on seasonal to annual time-scales (e.g. southwestern Black, central Marmara, Caspian and Aral Seas) would help evaluate the salinity, temperature, and nutrient requirements of both recent taxa and their fossil ancestors.
3. *Deep-water seascape evolution*. A further, still virtually unexplored field, concerns the exploration of deep-sea sedimentary processes and environments. One of the main problems related to the MSC is the genesis of widespread erosional features throughout the Messinian continental margins. The subaerial *or* subaqueous nature of such surfaces, their possible evolution from one to the other during the different MSC stages and associated base-level fall remain contentious. This debate could benefit from sharing data and experience with the community studying modern deep-sea environments where such surfaces form today.

5.2. The physical oceanographic perspective

The present-day Mediterranean, despite its small size and semi-enclosed nature, is characterized by a complex pattern of physical oceanographic processes, resulting from its strongly articulated topography and the large seasonal and interannual oscillations of temperature, salinity and freshwater input. The Mediterranean Sea is an important laboratory for understanding the generic controls on circulation patterns and dense-water formation (CIESM, 2009) and the impact of such processes (e.g. the formation of cold dense waters and their cascading off the continental shelf; Shapiro et al., 2003; Ivanov et al., 2004; Canals et al., 2006; Trincardi et al., 2007) on modern sedimentation and resulting stratigraphic record (Roveri et al., in press), has begun to be explored over the last few decades. This involves innovative interdisciplinary studies integrating oceanography, marine geomorphology, sedimentology and marine biology.

Transferring this information and approach to the study of Messinian events is a challenge. Reliable topographic and climatic data for the late Miocene is sparse and model experiments replicating evaporite precipitation and Mediterranean base-level change are mostly aimed at evaluating the timing and feasibility of such processes at a large scale. In reality however, the scale of spatial and temporal variability implied by the accumulation of a thick salt deposit in a deep basin, may be much smaller, with processes and gradients varying horizontally and vertically within individual sub-basins on interannual–seasonal timescales. How were local and basin-wide circulation patterns affected by this? What was the impact of sedimentary and erosional processes driven by the dynamics of water masses? These and many other questions generate new perspectives on the oceanography of Messinian events, which requires a tight multidisciplinary collaboration.

5.3. Exploring life in extreme environments

Evaporite basins are considered to be such extreme and inhospitable environments that it seems almost reasonable to imagine the Mediterranean experiencing catastrophic conditions able to wipe out most life forms from seawater. Yet, a visit to a natural evaporite basin or commercial salt works reveals salt brines that are densely packed with a low biodiversity assemblage of organisms that are able to survive such conditions. Since evaporite minerals are able to trap and preserve large quantities of microorganisms within their fluid inclusions, it follows that by studying ancient crystals we may be able to access archives of life throughout Earth history, and possibly on other planets such as Mars as well.

The survival of ancient organisms for hundreds of millions of years within salt crystals has been widely discussed since the 1960s, prompted by the isolation of bacteria within fluid inclusions in Permian salt (Vreeland et al., 2000). Today, modern facies analysis allows us to define which halite crystal and fluid inclusion may contain primary brines and associated biological material (Powers et al., 2001; Satterfield et al., 2005). The development of accurate laboratory protocols (Gramain et al., 2011) and the microscopic analysis of fluid inclusions allows the direct in situ identification of microorganisms, before their eventual extraction and culture, excluding laboratory contamination (Schubert et al., 2009). Recent research seems to confirm that prokaryotes (bacteria and archaea) have survived for at least 30,000 years within halite fluid inclusions from Death Valley. The organisms have been in a sort of “dormant” state; their survival ensured by obtaining energy to repair DNA damage using nutrients derived from the degradation of algae originally included with them (Lowenstein et al., 2011).

In this regard, the Messinian evaporites of the Mediterranean basin represent a natural laboratory for the study of ancient life as most of the rocks have not suffered sufficient burial or deformation (e.g. high pressure and temperatures) to destroy the primary fluid inclusions and their microorganisms (Lugli et al., 1999). Solid organic inclusions in Messinian gypsum has already enabled the identification and sequencing of the oldest cyanobacterium DNA ever extracted (Panieri et al., 2010; Fig. 21). Consequently, the study of MSC archives for micro-organisms will contribute to our knowledge of the origins and evolution of life on Earth and its ability to endure and regenerate itself during periods of extreme environmental change.

5.4. Modeling

Numerical modeling experiments can contribute to all the open questions outlined earlier. Both process-oriented models targeting specific mechanisms and the integration of model representations of various processes are important. The gateway region is already an important focus of model studies, but physics-based insight into the pattern and magnitude of exchange is, as yet limited. Modeling the interaction between the gateway and the Messinian Mediterranean's water properties and circulation is also required. Such studies require high, or spatially variable resolution to capture the complexity of the Mediterranean system and long simulation times to match the geological record. This, and the need to incorporate additional model variables that are more directly comparable to the geological data available (e.g. isotopes rather than salinity) remains a challenge.

5.5. The need for deep drilling in the Mediterranean

The significant advances in understanding Messinian events achieved over the last 40 years, has been on the basis of shallow marginal deposits, exposed, almost exclusively, on land. Access to samples of the deep, basinal deposits, which represent more than 80% of the volume of MSC sediments, has not been possible. We are now at a stage when, to tackle most of the persistent controversies and outstanding questions (see Section 4), deep drilling is needed and represents the main priority for future MSC research.

Multi-site, riser drilling, allowing the safe recovery of undeformed Messinian sequences and their bounding surfaces, will permit not only offshore–onshore stratigraphic correlation, but also marginal-to-deep and East-to-West correlation within MSC deposits. Such records should also resolve the timing of halite deposition, allow MSC base-level changes to be quantified and shed light on the subaerial/subaqueous nature of Messinian erosion surfaces. The cause–effect relationship between global, regional and local Messinian climate and the MSC is dependent on recovery of high-resolution basinal records, which will also provide information on the Mediterranean's connectivity history with Paratethys.

In addition to these objectives and the fundamental exploration of ancient microbial life in extreme pressure, temperature and salinity conditions, important and complementary objectives of multi-site drilling include understanding the role of salt tectonics in post-Messinian basin evolution and the effects of deep sea halite deposition and continental margin erosion on fluid formation and migration.

6. Conclusions

The Messinian salinity crisis can be defined as an ecological crisis caused by large amplitude environmental changes which developed in the Mediterranean at the end of the Miocene as a result of coeval effects of geodynamic and climatic forcings and their feedbacks.

Despite intense study, key questions remain unanswered, many awaiting new data from the deep offshore basins.

The MSC was an extraordinary and extreme event in the history of the Earth. It warrants further study because it has implications for many scientific, technological and economic fields, some of them, as yet, not clearly defined.

In this respect the MSC will represent for the next generation of Earth and Life scientists, just as big an intellectual challenge as it has posed for those of us involved in studying it over the last decades.

Acknowledgments

M.B. Cita, W.B.F. Ryan, K. Hsü and B.C. Schreiber are greatly acknowledged for many stimulating discussions and continuing dedication to the MSC event. A special thought goes to G. Clauzon, who passed away while the preparation of this manuscript was under way. We are also grateful to all those who attended the CIESM Workshop of Almeria (2007) and the Magellan Workshops of Brisighella (2013) and Paris (2014) for the constructive discussions, and particularly to J.A. McKenzie, J.-P. Suc, J.-M. Rouchy, F. Ricci Lucchi and G.B. Vai. We also thank DSDP, ODP and ESF for supporting drilling and workshops. D.J.W. Piper, B.C. Schreiber and M. Rabineau contributed with their thorough reviews to a significant improvement of an early version of the manuscript.

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Please cite this article as: Roveri, M., et al., The Messinian Salinity Crisis: Past and future of a great challenge for marine sciences, Marine Geology (2014), http://dx.doi.org/10.1016/j.margeo.2014.02.002

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Please cite this article as: Roveri, M., et al., The Messinian Salinity Crisis: Past and future of a great challenge for marine sciences, Marine Geology (2014), http://dx.doi.org/10.1016/j.margeo.2014.02.002

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Please cite this article as: Roveri, M., et al., The Messinian Salinity Crisis: Past and future of a great challenge for marine sciences, Marine Geology (2014), http://dx.doi.org/10.1016/j.margeo.2014.02.002