The transition from Pyrenean shortening to Gulf of Lion rifting in Languedoc (South France) - A tectonic sedimentation analysis

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The transition from Pyrenean shortening to Gulf of Lion rifting in Languedoc (South France) – A tectonic-sedimentation analysis.

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Abstract:

The pyrenean orogen extended eastward, across the present-day Gulf of Lion margin. The late or post-orogenic dismantling of this orogen segment, contemporaneous with ongoing shortening in the Spanish Pyrenees, is still debated. Understanding the transition between the two geodynamic events requires to document the precise timing of the succession of the tectonic processes involved. We investigate the superposition of rifting structures over pyrenean thrusts and folds in the onshore Languedoc. Compilation and reassessment of the regional chronostratigraphy, in the light of recent biostratigraphic dating and new mapping of Paleogene basins, lead to date the transition to the Priabonian. Tectonic-sedimentation relationship in the Eocene to Oligocene depocentres are analysed in surface exposures as well as in seismic reflection surveys. Bed-to bed mapping allowed us to: i) characterise an intermediate sequence of Priabonian age, bounded at the base and the top by unconformities; ii) evidence syn-depositional deformation within the Priabonian; iii) define the axes of Priabonian deformation. Interpretation of seismic reflection profiles, across the onshore basins covered by syn- and post-rift sequences, reveal the existence of an intermediate sequence displaying similar features, and that is correlated to the Priabonian. Syn-depositional deformation of some Priabonian basins correspond to extensional structure, whereas neighbouring, contemporaneous basins, reveal compressional deformation. The distribution of such apparently conflicting observations across the studied area provides evidence for left-lateral strike-slip deformation between two major regional faults (Cévennes and Nîmes faults). Left-lateral strike-slip along NE-trending faults accommodates E-W extension of the West European Rift (ECRIS) and part of the ongoing N-S shortening in the Central and Western Pyrenees. Priabonian clastic sedimentation and deformation in Languedoc witness the initial stages of the dismantling of the Languedoc-Provence Pyrenees, prior to Oligocene-Aquitanian back-arc rifting.

Keywords:

Pyrenees, Gulf of Lion, strike-slip basins, syn-depositional deformation, Priabonian

Résumé (français)

L'orogène pyrénéen s'étendait vers l'est à l'emplacement actuel du Golfe du Lion. Contemporain de la poursuite de la compression dans les Pyrénées franco-espagnoles, le démantèlement tardi- ou post-orogénique de ce segment de l'orogène est encore mal compris. La compréhension de la transition entre l'orogénèse Pyrénéo-Provençale et le rifting du Golfe du Lion nécessite de préciser le calendrier des évènements ainsi que la succession des différents processus tectoniques mis en jeu. Nous analysons la superposition des structures

du rifting sur les structures compressives Pyrénéennes exposées à terre, dans le Languedoc. Tout d'abord, nous compilons et révisons la chronostratigraphie régionale, à la lumière de datations postérieures à la publication des cartes au 1/50.000- du BRGM, et de la cartographie de plusieurs bassins paléogènes. Nous datons cette transition du Priabonien. Nous analysons les relations tectonique-sédimentation des dépocentres Eocène à Oligocène sur le terrain et en sismique réflexion. La cartographie banc-par-banc permet de : i) caractériser une séquence intermédiaire d'âge Priabonien encadrée par deux discordances, ii) mettre en évidence une déformation syn-dépôt dans le remplissage Priabonien, iii) définir les axes de la déformation priabonienne. L'interprétation de profils de sismique réflexion à travers les bassins paléogènes à terre, recouverts par les séries syn-et post-rift, révèle l'existence d'une séquence intermédiaire présentant des caractéristiques similaires et que nous corrélons avec le Priabonien. La déformation syn-dépôt de certains bassins est extensive, alors que dans des bassins voisins contemporains, elle est compressive. La distribution régionale de ces observations apparemment contradictoires est interprétée comme liée à un couloir de déformation en décrochement senestre entre les failles majeures des Cévennes et de Nîmes, d'orientation NE-SW. Ce décrochement accommode l'extension EW des rifts ouest-européens et contribue au raccourcissement N-S des Pyrénées franco-espagnoles, toujours actif au Priabonien. La sédimentation syn-tectonique clastique du Priabonien du Languedoc enregistre les premiers stades du démantèlement de la chaine pyrénéo-provençale, plusieurs millions d'années avant l'initiation de l'extension d'arrière-arc.

Mots-clé:

Pyrénées, Golfe du Lion, bassins sur décrochement, déformation syn-sédimentaire, Priabonien,

1. Introduction

The Pyrenean orogen results from the inversion of Mesozoic extensional basins, mostly formed during Mid-Cretaceous rifting and mantle denudation (Chelalou et al., 2016; Cochelin et al., 2018; Lagabrielle and Bodinier, 2008; Lagabrielle et al., 2010; Ternois et al., 2019 ; Tugend et al., 2014; Vergés and Garcia-Senz, 2001). The orogenic shortening was accommodated in the upper crust by reactivation of the inherited Cretaceous extensional faults, as thrust faults (Debroas, 1990), while the middle crust and exhumed mantle were subducted beneath the European crust (Teixell et al., 2018; Teixell et al., 2016). The orogen extended eastwards (Ford and Vergés, 2020), in Corbières (Lamotte et al., 2002; Viallard, 1987), Languedoc (Arthaud and Laurent, 1995) and southern Provence (Bestani et al., 2015; Lacombe and Jolivet, 2005), across the present-day Gulf of Lion (Arthaud and Séguret, 1981; Gorini et al., 1994). It was later dismantled in Oligocene time to give way to the Gulf of Lion rifted margin (Arthaud et al., 1981; Arthaud and Séguret, 1981; Gorini et al., 1994; Guennoc et al., 2000; Lacombe and Jolivet, 2005; Mascle and Vially, 1999; Mauffret and Gorini, 1996; Séranne, 1999; Séranne et al., 1995). In the onshore Corbières area, upper crust extension is characterised by “negative inversion” of the Pyrenean thrusts, associated with deposition of Oligocene-Aquitanian syn-rift sediments in the hanging-wall (Gorini et al., 1991). The crustal roots of the Pyrenees disappear eastwards (Chevrot et al., 2018) and the orogen brutally gives way to the Gulf of Lion rift and margin, within only tens of kilometres (Jolivet et al., 2020; Mauffret et al., 2001). Although documented both onshore and offshore, tectonics and kinematics of the superimposition of Pyrenean structures by later rifting (Fig. 1) remains incompletely understood. Transition of the compression to extension could be driven by collapse of the elevated mountain range (Séranne et al., 1995), it could be related to the southern propagation of the European Cenozoic Rift System (ECRIS) (Dèzes et al., 2004), or to the onset of back-arc extension due to the NW-dipping subduction roll-back of the old Tethyan lithosphere (Jolivet et al., 2020; Séranne, 1999). Better constraints on tectonics and precise timing of the transition from the Pyrenean orogeny to the Gulf of Lion rifting are needed in order to better understand the origin and driving mechanism of the transition between the two major geodynamics events. Such data are also needed to constrain paleotectonic restorations (e.g. (Bestani et al., 2015; Christophoul et al., 2016), and kinematic reconstructions of the Mediterranean (e.g. (Romagny et al., 2020).

Since most of the collapsed orogen is presently buried beneath the thick Neogene sequence of the Gulf of Lion margin, the onshore Languedoc provides the only geological outcrops that are relevant to address this question. There, the rare available stratigraphic data suggested that the Pyrenean orogeny lasted until “early Oligocene”, as reported on the Saint-Martin-de-Londres 1/50.000-scale geological map (Philip et al., 1978), while rifting took place in the late

Rupelian according to the Montpellier map (Andrieux et al., 1971). Consequences of such a short delay implies: i) that Pyrenean compression was polyphase, including a last compressive phase, following a Late Eocene period of tectonic quiescence, and ii) rifting interrupted folding and thrusting, with no transition.

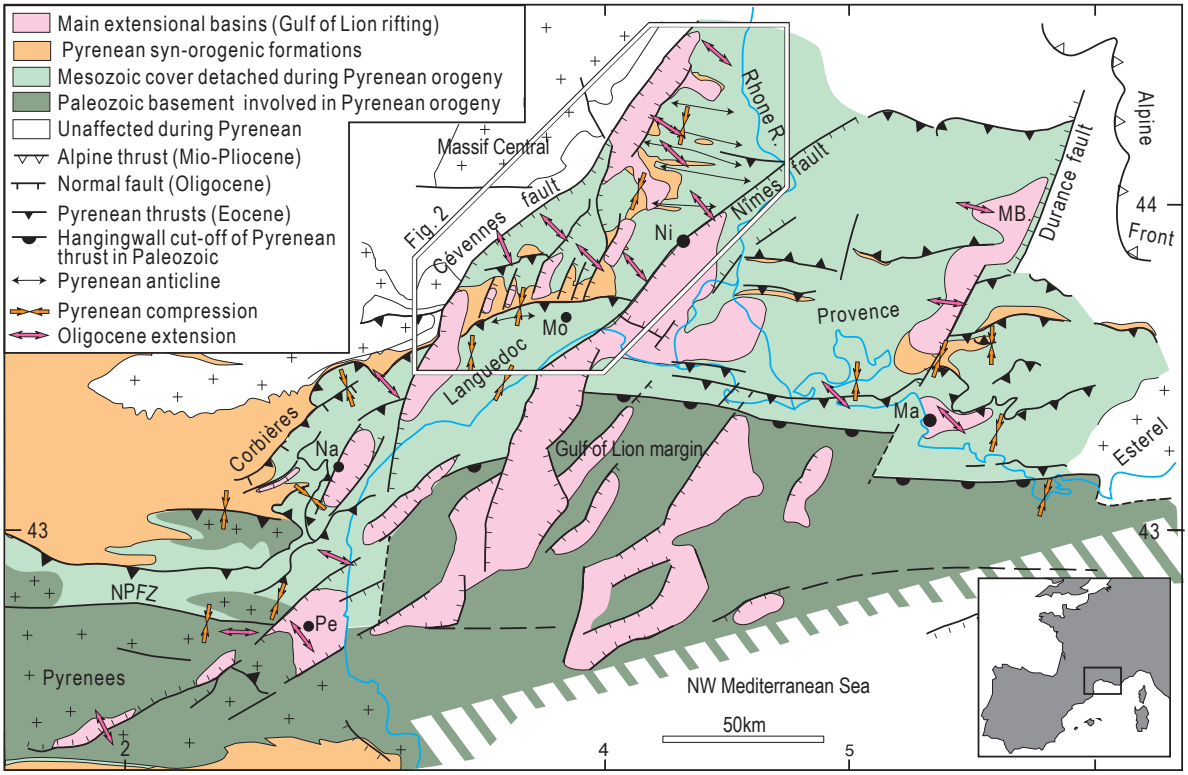


Fig. 1: *Structural map southern France showing the relation between the Pyrenees and Gulf of Lion margin; modified from (Séranne et al., 1995). Na: Narbonne, Pe: Perpignan, Mo: Montpellier, Ma: Marseille. Directions of Pyrenean shortening from (Arthaud and Laurent, 1995), (Gaviglio and Gonzales, 1987; Lacombe et al., 1992; Lamotte et al., 2002) and Oligocene extension from : (Arthaud et al., 1977; Benedicto, 1996; Hippolyte et al., 1993).*

In this paper, we examine five Paleogene basins located in Languedoc, which developed at the front of the Eocene external thrusts of the Pyrenean belt and onshore of the Gulf of Lion passive margin. The main objective of this contribution is to document, thanks to the revision of available biostratigraphic data and to detailed sedimentological, tectonic and seismic analyses, the key period of the transition between the last stages of Pyrenean orogenic (mountain building) events and the earliest stages of rifting associated with the opening of the Gulf of Lion. We aim at constraining the tectonic setting associated with the deposition of Priabonian continental units by analysing the tectonic-sedimentation relationships in 5 distinctive basins (Fig. 2). Analyses are based on field observations (Guzargues, Les-Matelles, Saint-Martin-de-Londres basins) and subsurface data (Hérault and Saint-Chaptes Gardon basins). This study also relies on: i) new original investigation in the Guzargues basin, ii)

reappraisal of published works in Les-Matelles and Saint-Martin-de-Londres basins (the latter involving additional mapping), and iii) the interpretation of ancient (1983-1992) seismic profiles tied to boreholes data in the Hérault and Saint-Chaptes Gardon basins.

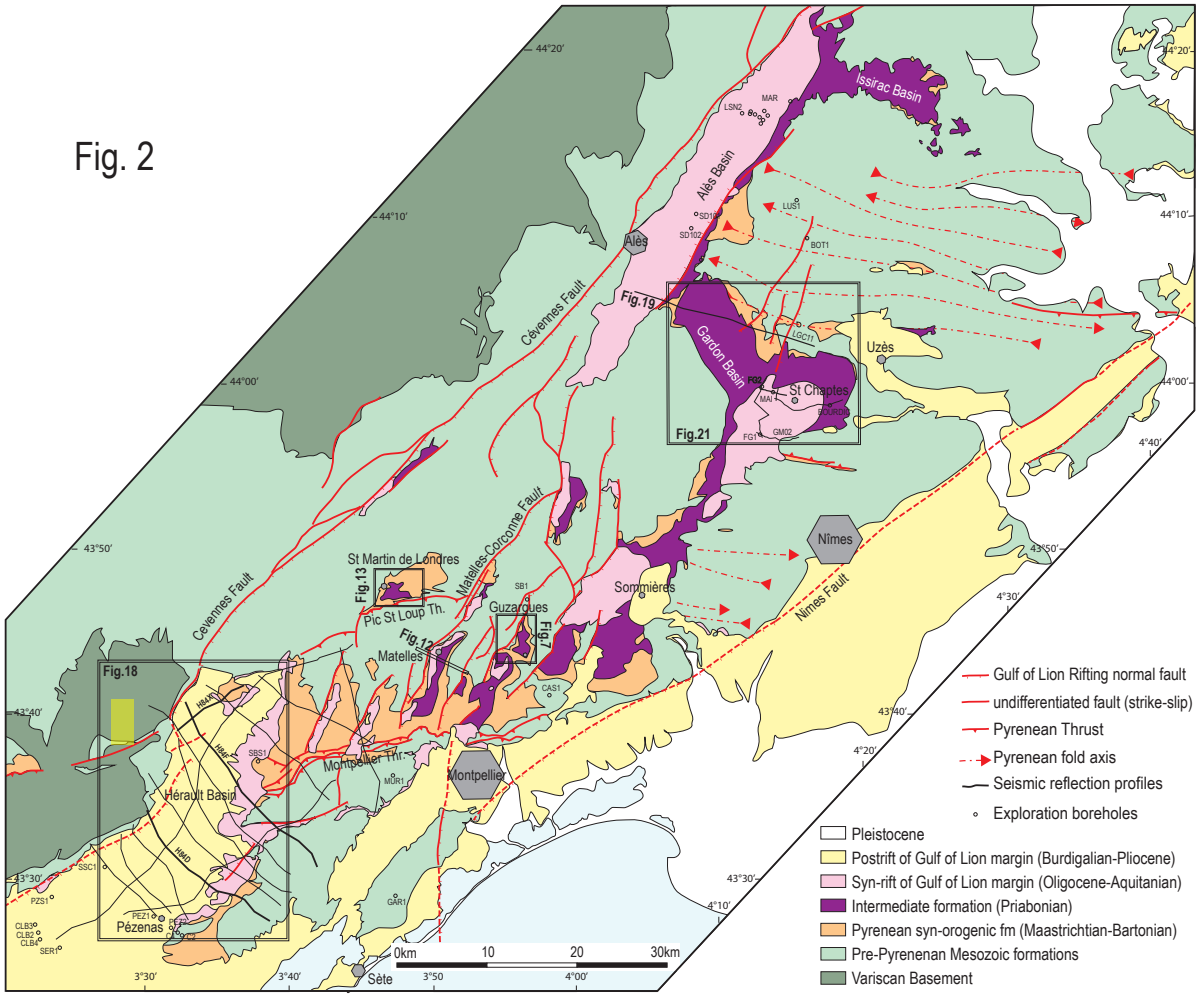


Fig. 2: *Simplified geological map of Languedoc, in the northern foreland of the Provence-Pyrenees fold and thrust belt. See location in the regional structural framework of Fig.1. Priabonian basins according to the present study, in some areas, stratigraphic attributions are different from the BRGM 1/50.000-scale geological maps (see text for explanation). Subsurface data used in the study is indicated fine black lines (seismic) and open circles (boreholes); seismic profiles shown in this contribution are in bolder black line. Location of detailed maps and sections is shown with double black line frames.*

2. Regional geological setting

2-1 Location of the studied area within the Pyrenean orogen

Languedoc, in the south of France, is located between the Variscan basement of the Massif Central and the shores of the Mediterranean Sea; it links the Pyrenees (in the SW) to the Alps (to the NE), across the Rhône Valley (Fig. 1). Together with the Corbières, Languedoc forms

a segment of the north foreland of the Pyreneo-Provençal fold and thrust belt (Arthaud and Séguret, 1981; Bestani et al., 2015).

The Paleozoic (Variscan) basement is exposed in the Massif Central to the NW, in the Maure and Esterel massifs to the east. To the south, the basement has been reactivated during the Late Cretaceous to Eocene Pyrenean orogeny. It is exposed in the Pyrenees and buried beneath the thick Neogene sequences of the Gulf of Lion margin (Arthaud and Séguret, 1981; Gorini et al., 1994; Guennoc et al., 2000) (Fig. 1). Onshore, the Variscan basement is unconformably covered by Mesozoic sequence, including at the base variable amounts of Triassic detritals and evaporites (Debrand-Passard and Courbouleix, 1984), which have been used as decollement level during later tectonic phases. The Jurassic through to Neocomian marine marls and carbonates were deposited during the Tethyan rifting and subsequent thermal subsidence, controlled by NE-trending normal faults such as the Cévennes, Nîmes and Durance faults. Thickness of the Jurassic increases southeastwards from the Cévennes fault (less than 1,5km), and maximum subsidence is recorded in the hanging-wall of the Nîmes Fault, where seismic profiles show up to 10km of Trias to Neocomian sequences (Le Pichon et al., 2010; Séguret et al., 1997).

This Mesozoic sedimentary cover has been folded and thrusted in the Pyrenean foreland, detached over Triassic decollement. Thick-skinned thrusting presumably occurred close to the present day shoreline (Arthaud and Séguret, 1981; Lacombe and Jolivet, 2005; Séranne et al., 1995), The style of the thrust and fold belt is mostly controlled by the thickness of the detached Mesozoic sedimentary cover (Arthaud and Laurent, 1995): north-verging, EW-oriented thrusts dominate areas of thin cover (≤ 2,5 to 3 km) whereas thicker cover developed series of E-W folds. Combined thin- and thick-skinned shortening exceed 20km in the Languedoc (Arthaud and Laurent, 1995) and 40km in the Provence segment of the orogen (Bestani et al., 2016; Bestani et al., 2015). The Pyrenees-Provence fold and thrust belt is segmented by NE- to NNE-trending faults which separate areas of distinct structural styles and amounts of shortening. Left-lateral strike-slip along the Cévennes Fault accommodated shortening of its hanging-wall while the NW footwall remained mostly undeformed during the Pyrenean orogeny (Arthaud and Laurent, 1995). In Provence, the Durance Fault separates thin- and thick-skinned Pyrenean deformation (Bestani, 2015; Bestani et al., 2016).

2-2. Location of the studied area within the Gulf of Lion rift and passive margin

The studied area is affected by NE-trending normal faults bounding syn-rift Oligocene basins (Fig. 2). At least, some of these faults - if not all - are reactivated previous Pyrenean left-lateral strike-slips faults (i.e. the Cévennes fault, Matelles Fault (Mattauer, 2002). Seismic reflection profiling has shown that, during Oligocene rifting, such faults detached in the Triassic

decollement (Benedicto, 1996; Benedicto et al., 1999; Husson, 2013; Sanchis and Séranne, 2000). The onshore syn-rift basins NW of the Nîmes Fault were formed by thin-skinned extensional tectonics (Séranne, 1999). The basement-cover decollement ramps down in the Nîmes basement fault (Benedicto et al., 1996) and into the basement ramp of the Montpellier thrust to the south and southeast of the study area.

3. Revised chronology of events

The chronology of the transition between Pyrenean orogeny and the Gulf of Lion rifting is still discussed. The stratigraphy of the 1/50.000-scale geological maps of the study area is confusing: all maps are consistent in identifying the Bartonian (e6) as a Pyrenean syn-tectonic sequence, but they greatly diverge in dating the end of the orogeny and the onset of the rifting. The comparison of stratigraphy of the adjacent maps of the study area reveals contradicting ages for identical and spatially continuous formations (Fig. 3 a).

Another unconsistency arises from the outstanding question: what was the geodynamic context during the time interval between the Pyrenean syn-orogenic deposits and the syn-rift formations? Adjacent maps carry contradicting interpretations: the « g1 » is a continuity of the syn-orogenic Bartonian on the Saint-Martin-de-Londres map (Philip et al., 1978), whereas the Montpellier map correlates the « g1 » with the early stages of the Gulf of Lion rifting (Andrieux et al., 1971).

The mammalian reference levels developed for the Paleogene continental record of Europe (Sigé and Legendre, 1997), allowed to improve the stratigraphy and to revise the tectonic agenda of the area, especially around the transition from orogeny to rifting. Figure 3b uses chronology of (Vandenberghe et al., 2012) and mammal reference levels of (Aguilar et al., 1997).

The age of the lacustrine limestone was defined as Lutetian or slightly older. In the Pic-Saint-Loup foreland, fauna allowed to place this formation between MP12 and MP13 reference levels of the mammal reference levels (Crochet et al., 1988). In different localities of Les-Matelles basin, several lignite beds provided a diversified fauna, which correlates with the Lutetian (MP13) (Crochet et al., 1997). A distinctive marly facies including 1 to 20cm diametre oncolithes, occurs at the top. It has yielded fauna ascribed to the MP14 reference level (Crochet et al., 1997) of latest Lutetian age. This facies passes southwards and upwards to the Bartonian syn-orogenic breccia, exposed along the Montpellier thrust (Andrieux et al., 1971). In the Sommières and Saint-Chaptes basins the reference sites of Robiac provided a late Bartonian age for the latest Pyrenean syn-tectonic formations (Remy, 2015).

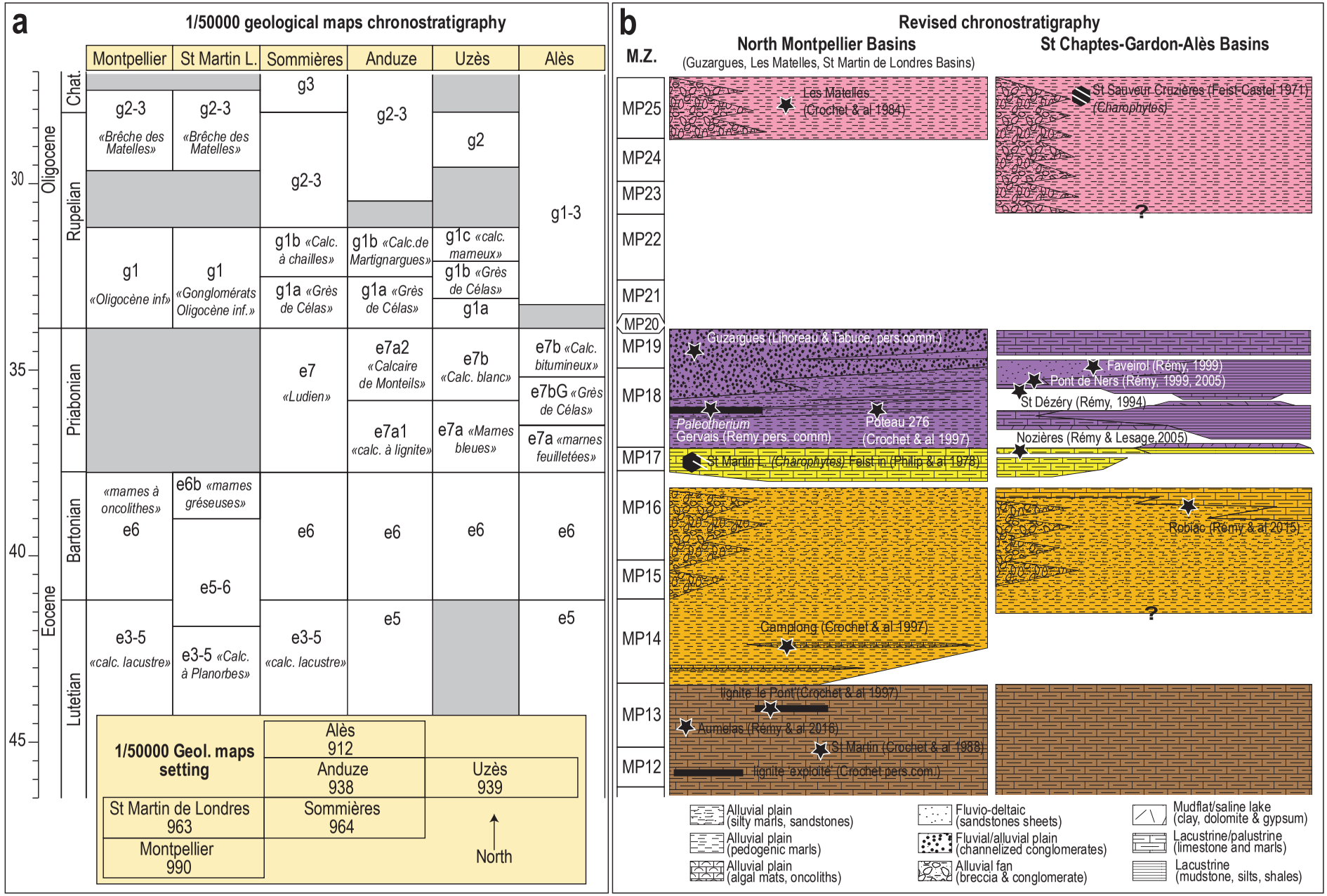

a
1/50000 geological maps chronostratigraphy
Montpellier
St Martin L.
Sommières
Anduze
Uzès
Alès
Oligocene
Eocene
Chat.
Rupelian
Priabonian
Bartonian
Lutetian
30
35
40
45
g2-3
«Brêche des Matelles»
g3
g2-3
g2
g1-3
g1b «Calc. à chailles»
g1b «Calc.de Martignargues»
g1c «calc. mameux»
g1b «Grès de Célas»
g1a «Grès de Célas»
g1
«Oligocène inf»
«Gonglomérats Oligocène inf.»
g1a
e7b «Calc. bitumineux»
e7
«Ludien»
e7a2 «Calcaire de Monteils»
e7b «Calc. blanc»
e7bG «Grès de Célas»
e7a1 «calc. à lignite»
e7a «Marnes bleues»
e7a «marnes feuilletées»
«marnes à oncolithes»
e6
e6b «marnes gréseuses»
e5-6
e3-5
«calc. lacustre»
e3-5 «Calc. à Planorbes»
e5
1/50000 Geol. maps setting
Alès 912
Anduze 938
Uzès 939
St Martin de Londres 963
Sommières 964
Montpellier 990
North
b
Revised chronostratigraphy
M.Z.
North Montpellier Basins
(Guzargues, Les Matelles, St Martin de Londres Basins)
St Chaptes-Gardon-Alès Basins
MP25
MP24
MP23
MP22
MP21
MP20
MP19
MP18
MP17
MP16
MP15
MP14
MP13
MP12
Les Matelles (Crochet & al 1984)
St Sauveur Cruzières (Feist-Castel 1971) (Charophytes)
Guzargues (Lihoreau & Tabuce, pers.comm.)
Faveirol (Rémy, 1999)
Pont de Ners (Rémy, 1999, 2005)
St Dézéry (Rémy, 1994)
Paleotherium Gervais (Remy pers. comm)
Poteau 276 (Crochet & al 1997)
Nozières (Rémy & Lesage,2005)
St Martin L. (Charophytes) Feist in (Philip & al 1978)
Robiac (Rémy & al 2015)
Camplong (Crochet & al 1997)
lignite 'le Pont' (Crochet & al 1997)
Aumelas (Rémy & al 2016)
St Martin (Crochet & al 1988)
lignite exploité (Crochet pers.com.)
Alluvial plain (silty marls, sandstones)
Alluvial plain (pedogenic marls)
Alluvial plain (algal mats, oncoliths)
Fluvio-deltaic (sandstones sheets)
Fluvial/alluvial plain (channelized conglomerates)
Alluvial fan (breccia & conglomerate)
Mudflat/saline lake (clay, dolomite & gypsum)
Lacustrine/palustrine (limestone and marls)
Lacustrine (mudstone, silts, shales)

Fig. 3 a: *Correlation chart of the Eocene to Oligocene interval, and inferred ages, according to the BRGM 1/50.000-scale geological maps of the study area. The inset in the lower part gives the position of the BRGM 1/50.000-scale geological maps.* ***b:*** *Revised chronostratigraphy of the basins north of Montpellier (covered by the geological maps of Montpellier, St Martin-de-Londres and part of Sommières and Anduze) and of the St Chaptes-Gardon basins (maps of Alès and part of Anduze and Uzès). This is based on new mapping of the Paleogene basins (this study) and on the compilation of biostratigraphy data, including recent discoveries of mammal fossils (references in the figure and in the text). Biostratigraphy scale according to mammal reference levels (M.Z.). Black stars and black hexagons correspond to mammal and charophyte sites, respectively. Simplified lithology of the formations is indicated. Colour code as in the following maps and sections of the Paleogene basins analysed in this study. White areas corresponds to non-depositional and erosional hiatuses; the Priabonian interval from St Chaptes-Gardon-Alès basins is documented in (Lettéron et al., 2018).*

They pass upward to continental marls and limestones dated to the Bartonian-Priabonian transition (MP17) (Remy, 2015; Rémy and Lesage, 2005). In the Saint-Martin-de-Londres basin, similar lithologies correspond to time equivalent, as suggested by charophytes (Feist in (Philip et al., 1978).

The overlying formation mapped as « g1 » in Les-Matelles basin, displays polygenic conglomerates and sandstones within yellow marls and silts, including several lignite beds. They are interpreted as alluvial plain deposits (Benedicto et al., 1999; Egerton, 1996). The age of this formation was subjected to discussion due to the controversial location of the discovery by Gervais, in the late XIXth Century, of a *Lophiodon* skull (see review in (Hartenberger et al., 1969). Reappraisal of the specimen, exhumed from Lyon university collections by J. A. Remy (*personal communication*), relates the specimen to MP 18 (middle Priabonian). Another *Lophiodon* specimen has recently been discovered in the correlative « g1 » formation in the Guzargues basin, and has been related to the MP19 reference level (R. Tabuce and F. Lihoreau, *personal communication*). The age of this conglomerate formation is therefore attributed to the Priabonian. As first suggested by (Andrieux et al., 1971), it is correlated with the detrital continental formation « *Grès de Célas* » of upper Priabonian, in the Sommières and Saint-Chaptes - Gardon basins, northeast of the study area, which spans the MP18 to MP19 (Rémy, 1985; Remy and Fournier, 2003; Rémy and Lesage, 2005).

At Les-Matelles and Saint-Vincent-de-Barbeyrargues localities, yellow to orange marls interfingering with syn-rift breccia, contain charophytes (Grambast, 1962), teeth of small mammals (Thaler, 1962) and gastropods (Rey, 1962) that yielded mid- to late- «*Stampian*» age, which corresponds to the Late Rupelian. This age was later confirmed by the discovery of small mammals in Les-Matelles half-graben, that relate to MP25 mammal scale (Crochet, 1984). In addition, breccia pipes and dykes of intracontinental alcaline basalt, that intrude Lutetian and Priabonian formations, were dated 24 to 28 Ma (Gastaud et al., 1983), and have

been related to the Late Rupelian rifting event (Dautria et al., 2010; Séranne, 1999). In the northern basins, this unconformable formation has not yet provided mammal remains, but Charophytes from the alluvial environments of Saint-André-de-Cruzières in the north of Alès Basin yielded an Oligocene age (Feist-Castel, 1971).

Reappraisal of ancient specimens and new discoveries thus indicate that the sequence made of marls, sandstones beds and polygenic channelized conglomerates, which is exposed in separated basins across the study area, belongs to the same interval, dated to the Priabonian. In particular, the fluvial conglomerate and yellow sandy marls (reported on the published maps as “g1”, Fig. 3a), correlate with, and are the proximal equivalent of the lacustrine delta sandstones “*Grés de Célas*” of the Alès Basin (Lettéron et al., 2018). This formation is intermediate between the syn-orogenic Pyrenean (Bartonian) and the syn-rift (Late Rupelian) deposits. Furthermore, it is bounded by two marked angular unconformities, the upper one corresponding to a several-million-years hiatus.

In the following, we examine key field evidence in the north-Montpellier area (Guzargues, Les-Matelles and St-Martin-de-Londres basins, respectively) where the tectonic sedimentation relationships allow to characterise this intermediate formation. It can be interpreted as the record of the transition between the Pyrenean orogeny to the Gulf of Lion rifting.

4. Tectonic and sedimentation in the Guzargues Basin

4-1. Lithostratigraphy of Eocene deposits in the Guzargues area

The Guzargues basin (Fig. 2) has not been the topic of any specific study and remains poorly documented. The Paleogene sedimentary succession contains 4 main lithological units representative of the regional geodynamic history since the early stages of the Pyrenean compression to the main rifting phase (Fig. 4). When possible, all of them were precisely mapped to analyse the geometrical relationships of Priabonian deposits with under- and overlying units.

(1) Paleocene to Lower Eocene facies are represented by fluvial marls with interbedded sandstones and palustrine limestones (“*Marnes infra-lutétiennes*”), only few tens of metres thick in the Guzargues area. They are usually correlated to alluvial fan syn-tectonic breccias, reworking the Jurassic succession at the front of the Montpellier thrust, and thus deposited during the early stages of the Pyrenean compression.

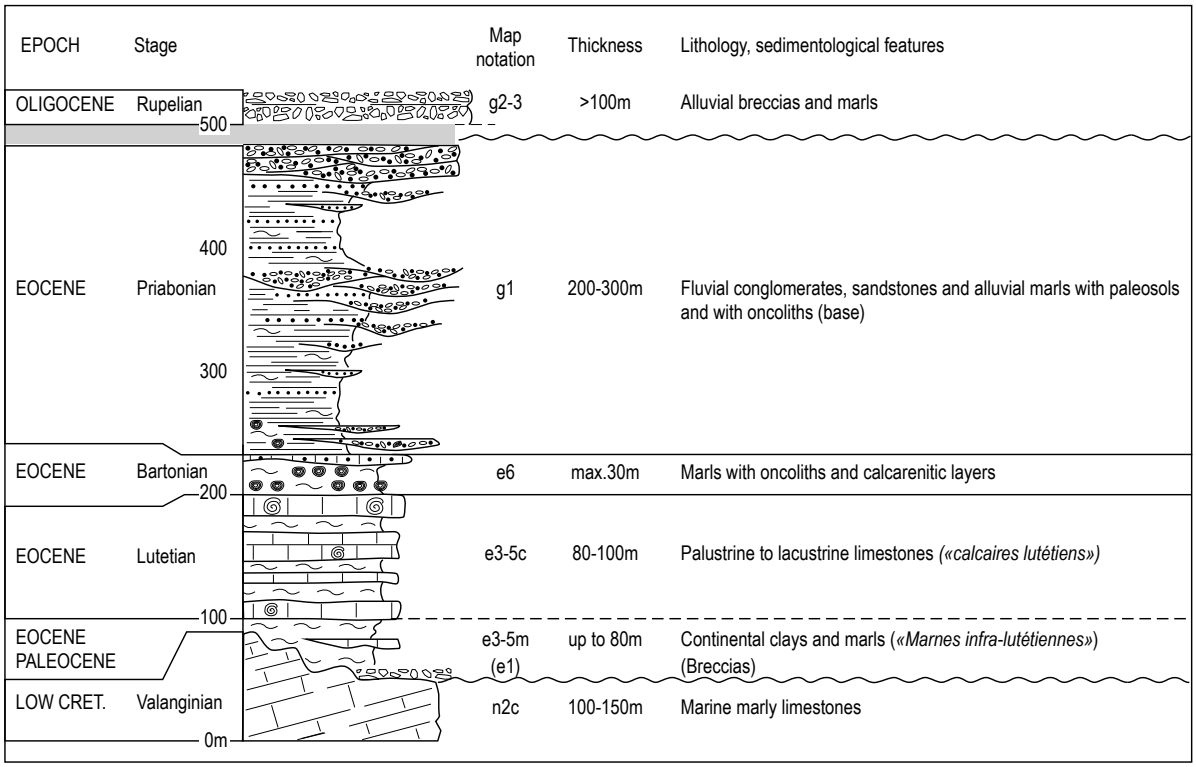


Fig. 4: *Generalised lithostratigraphy of the Late Mesozoic and Paleogene in the Guzargues Basin and immediate surrounding.*

(2) Massive whitish Lutetian limestones, up to 100m thick in the area, form an easily recognised, continuous unit displaying folds produced by late stages of the Pyrenean compression.

(3) Lutetian limestones are conformably capped by a singular unit, up to 30m thick, made of lacustrine to shallow marine, marly facies containing oncoliths and calcarenitic layers, Bartonian in age.

(4) Topmost deposits of the Paleogene succession in the Guzargues area correspond to Priabonian fluvial deposits showing alternations of conglomerates, sandstones and marls, up to 300 m thick. Pebbles are supplied both from distant Pyrenean sources and from local Mesozoic carbonates. These deposits rest unconformably above the Pyrenean folds consisting of Lutetian to Bartonian units.

One kilometre to the south, the northern edge of the neighbouring Assas basin shows a sedimentary unit of strong geodynamic significance, which is missing in the Guzargues basin. Close to the Assas village, Oligocene breccias and marls, usually interpreted as syn-rift deposits, unconformably overlie Priabonian continental facies.

Thus, the Priabonian deposits of the Guzargues basin, bounded below and above by unequivocal syn-tectonic units, represent a key unit to study the transition between two main regional geodynamic events.

4-2. Sedimentological features of Priabonian deposits in the Guzargues basin

As observed in the 10 sections studied in the Guzargues basin, Priabonian deposits are basically made of eight lithofacies units distinguished by their lithology and sedimentary structures. Each lithofacies is described and interpreted in terms of depositional processes (Table 1), and most are illustrated on Figure 5.

Lithofacies units are often associated together, defining three main facies associations (FA1, FA2, and FA3), the distribution of which varies depending on the spatial location in the Guzargues basin and on the vertical position into the Priabonian stratigraphic succession.

FA1 is represented by the alternations of (i) channel-shaped, poorly to weakly sorted conglomerates (Gu, Gt; Table 1), (ii) matrix-rich, stratified conglomerates (Gm; Table 1), and (iii) medium- to coarse-grained, cross-bedded sandstones (Sp; Table 1). A general trend in the vertical evolution of FA1 facies units is recognised through most of the studied sections. In the lower part of sections, predominant Gu/Gt conglomeratic channels, up to 3m thick and 3 to 5m wide, contain heterometric, sub-rounded to well-rounded clasts, mainly composed of local Mesozoic (Upper Jurassic to Lower Cretaceous) carbonates. In some places, they also show abundant oncoliths (as fragments or complete specimens) probably originated from the underlying Bartonian formation. In the upper part of sections, conglomeratic bodies in Gm facies unit are thinner (< 1m) and more laterally continuous (usually more than 10 metres in width) than in the lower part. In addition to local Mesozoic carbonates, quartz and lydite, originated from Cévennes and/or Montagne Noire Paleozoic peripheral units, and ginger (Aptian?) sandstones represent a significant proportion of clasts in channel fillings.

Gu conglomerates can be interpreted as channel lag (Miall, 1977, 1978; Nemec and Postma, 1993).The **Gm** and **Gt** facies units present strong similarities with gravel bars in braided streams (Mack and Leeder, 1999; Miall, 1977; Nemec and Postma, 1993) . **Sp** cross-bedded sandstones are typical of megaripples deposited by a unidirectional upper to lower flow regime (Simons et al., 1965; Southard, 1991) produced by subaerial to subaqueous, stream and/or sheet flows (Hampton and Horton, 2007; Mack and Leeder, 1999). Inferred FA1 depositional environments are thought to belong to an alluvial fan setting, first supplied by a local catchment area (lower part of sections), and then by a regional catchment area (upper part of sections).

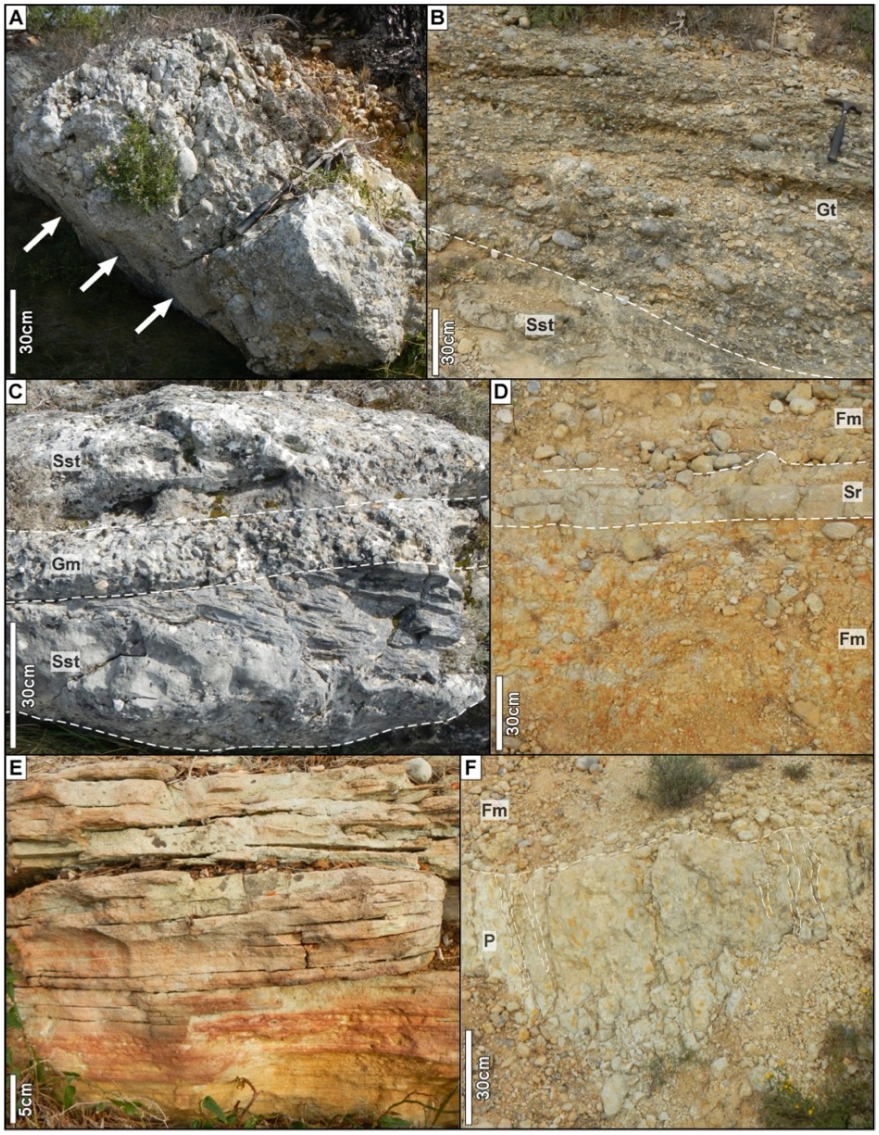


Fig. 5: *Photographs of outcrops illustrating some typical facies observed in the Priabonian deposits of the Guzargues basin.* ***A:*** *Facies Gu with carbonate clast alignment underlining tenuous cross-beddings. Note scours at the base of the channel (arrows).* ***B:*** *Facies Gt showing progressive transition between (base) unsorted, highly heterometric massive, stratified conglomerates, and (top) weakly sorted, cross-bedded conglomerates.* ***C:*** *Alternations between (Sst) cross-stratified sandstones with lenses of scattered subrounded gravels and pebbles, and (Gm) matrix-rich, graded, stratified conglomerates with weakly sorted pebbles mainly made of Mesozoic carbonates.* ***D:*** *Reddish to yellowish laminated silty claystones (facies Fm) with intercalation of a decimetric layer of massive fine-grained sandstones (facies Sr).* ***E:*** *Medium- to coarse-grained cross-bedded sandstones typical of facies Sp.* ***F:*** *50cm-thick carbonated horizon showing well-preserved, vertical root casts, probably derived from vegetalization on top of massive fine-grained sandstones (Sr).*

Note the occurrence of several pedogenetic carbonate nodules above the irregular top surface of the carbonated sandy horizon.

Facies code. Facies name	Lithological features	Sedimentary structures	Inferred depositional processes
Gu. Poorly sorted, massive, channel-shaped conglomerates (Fig. 7A)	Poorly sorted, massive, locally normally graded conglomerates. Highly heterometric, sub-rounded to well-rounded clasts (pebbles to small boulders), mainly composed of local Mesozoic carbonates	Well-defined basal erosional surfaces with gutter/scour structures. Channels displaying few clast imbrications, 50cm up to 3 meters in depth, 3 to 5 meters in width	Channel-lag deposits under subaerial flashy braided-stream flows (Barrier et al., 2010)
Gt. Weakly sorted, cross-bedded, stratified conglomerates with coarse-grained sand lenses (Fig. 7B)	Clast-supported, stratified conglomerates. Weakly sorted, heterometric, sub-rounded clasts (granules to small boulders of Mesozoic carbonates). Occurrences of poorly sorted lenses of coarse-grained sand matrix surrounding individual pebbles	Fining upward infill. Trough cross-bedding	Minor channel fills (Miall, 1978). Transverse- or diagonal-bar deposits under subaerial perennial braided-stream flows (Barrier et al., 2010)
Gm. Matrix-rich, normally graded, stratified conglomerates (Fig. 7C)	Matrix-rich, normally graded, stratified conglomerates. Poorly sorted, subangular to sub-rounded clasts (granules to cobbles) made of local Mesozoic carbonates and some siliceous elements	Some beds with horizontal alignment of clasts. Few clast imbrications into horizontal foreset strata. "Gradual" basal surfaces	Longitudinal-bar deposits under subaerial competent and perennial braided-stream flows (Barrier et al., 2010)
Sst. Moderately sorted, medium- to coarse-grained, cross-stratified sandstones with lenses of sub-rounded gravels and pebbles (Fig. 7C)	Moderately sorted, medium- to coarse-grained sandstones. Layers (and lenses) of well-sorted subrounded clasts (granules to pebbels) composed of Mesozoic carbonates, some Lutetian mudstones and common Bartonian oncoliths (as fragments or complete specimens)	Fining-up trend. Cross-stratification with grading foresets. Well-defined internal erosional surfaces	Bar deposits under subaerial sheet flows or stream flows
Sp. Medium- to coarse-grained, cross-bedded sandstones (Fig. 7E)	Weakly sorted, medium- to coarse-grained sandstones which can contain scattered granules at the base	Cross-stratification. Internal erosional surfaces. Few small scour structures	Megaripple deposits under subaerial sheet flows or stream flows
Sr. Massive, pedoturbated, well-sorted, fine-grained sandstones (Fig. 7D)	Massive, well-sorted, fine-grained sandstones	Pedoturbation (root sleeves, destructuration of primary sedimentary structures by root growth). Locally few asymmetrical ripples at the top	Ripple deposits under subaerial sheet flows or stream flows
Fm. Massive, silty claystones with planar siltstone to fine-grained sandstone layers (Fig. 7D)	Reddish to yellowish (also purplish red in some horizons of plastic texture), massive, silty claystones with few thin (mm to cm) layers of well-sorted siltstones and very fine-grained sandstones	Fining-up grading in siltstone to fine-grained sandstone layers	Deposits from suspension fallout in a floodplain. Overbank or waning flood deposits (siltstones and very fine-grained sandstones)
P. Pedogenetic carbonate nodules (Fig. 7F)	Indurated, cm to dm carbonate nodules, often scattered within plastic claystones, sometimes coalescent to form continuous carbonate horizons	Vertical root casts	Soil (Miall, 1978) or temporary vegetalized surface

Table 1*: Summary of the main facies features observed in the Priabonian deposits of the Guzargues basin, and interpretations in terms of depositional processes. Gm, Gt, Sp, Sr, Fm and P, are abbreviations from the facies code provided by (Miall, 1978). Gu and Sst are abbreviations defined by (Barrier et al., 2010).*

FA2 corresponds to the alternation between (i) moderately sorted, medium- to coarse-grained sandstones with sub-rounded clast lenses (**Sst**; Table 1), and (ii) medium- to coarse-grained cross-bedded sandstones (**Sp**; Table 1). If local (Mesozoic but also Lutetian) calcareous elements still represent an important proportion of clasts (granules, gravels, pebbles), siliceous component (quartz and lydite gravels) is more abundant than in FA1. FA2 depositional processes are dominated by tractional transport under a unidirectional upper flow regime, with episodic sheet flows. Inferred depositional environments correspond to the proximal part of a subaqueous alluvial fan to fan delta, with sediment supplied from both local and regional catchment areas.

FA3 is mainly represented by yellowish, greenish or reddish silty clays, with plastic texture at some places, in which are interbedded thin layers of well-sorted siltstones and fine-grained sandstones (**Fm**; Table 1). Carbonate pedogenic nodules locally form scarce, up to 1m thick, distinctive layers (**P**; Table 1). Interbedded massive fine-grained sandstones (**Sr**; Table 1) usually show vertical root traces and more rarely asymmetrical ripple cross-laminations several centimetres thick at the top.

FA3 deposition must be the result of vertical settling from detrital suspensions in standing water (Horton and Schmitt, 1996; Miall, 1977). Fine-grained sandstones showing asymmetrical ripples and interbedded silty and sandy layers suggest episodic flow regime, related to sheet flood events. The occurrence of carbonate nodules records occasional pedogenic processes and temporary aerial exposure. Assuming these depositional processes, FA3 should have been deposited in the distal part of a subaqueous to subaerial alluvial fan, or in a floodplain laterally to major braided distributaries.

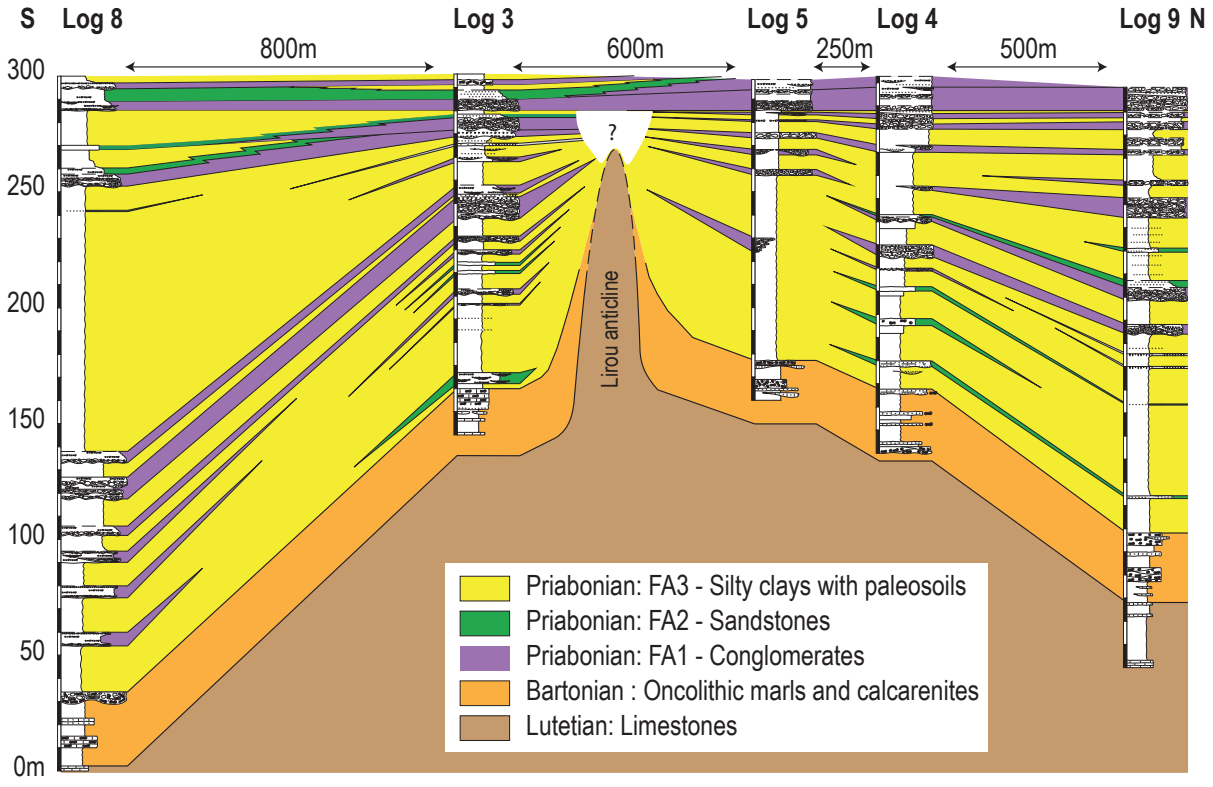


Fig. 6: *Detailed stratigraphic correlation of Priabonian deposits along a 2km-long N-S transect at the southeastern border of the Guzargues basin, showing differential subsidence and onlap geometries induced by the syndepositional growth of the Lirou anticline. See Fig. 7 for location of logs. Horizontal correlation lines set at the maximum of progradation recorded by several meters-thick, stacked conglomeratic channels that closely rests on Lutetian limestones at the apex of the Lirou anticline.*

The 10 studied sections show vertical variations in the FA succession (basically FA1-FA2 or FA2-FA3 alternations) forming elementary sequences between several metres and tens of metres thick (Fig. 6). As indicated by the various inferred depositional environments for FA, these elementary sequences, made up of alternatively proximal and distal deposits, are arranged into overall trends measuring several tens of metres thick, recording progradational and retrogradational evolutions.

At the scale of the Guzargues basin, the large FA alternations are describing a whole progradational, then retrogrational, and finally progradational evolution, which is easily

recognised on the field. These trends were used to correlate Priabonian deposits at basin scale and, combined with aerial photograph interpretation, to produce a detailed geological map that allows building cross-sections.

4-3. Tectonic-sedimentation relationships in the Guzargues basin

The detailed geological map (Fig.7) and cross-sections (Fig. 8) of Guzargues basin show sequences of Priabonian conglomerates interfingering with marls, that are involved in gentle folding of complex geometry. To the south, the basin is separated from the contemporaneous Assas Basin by a NW-SE anticline, cored by Valanginian (Fig.7 and Fig.8 - Section 1). North-south sections of Guzargues basin (Fig.8 - Sections 2, 3) display an asymmetrical syncline structure: the southern limb is steeply dipping north, whereas the northern limb extends over 3 to 4 km in a gently dipping or subhorizontal position. In addition, the basal part of the Priabonian is affected by ENE-trending anticlines, while the youngest conglomeratic Priabonian seals the anticline (Fig. 8 - Section 3).

The Lirou anticline - The Bartonian and the base of the Priabonian sequence (isolated sandstone and conglomerate beds in marls) are wedging towards an anticline with an EW-oriented, W-plunging axis. Conglomerate beds isolated in the silt and marls are observed on-lapping both flanks of the anticline (Fig.6), and they display diverging bedding away from the anticline (Fig. 7, Fig. 9a). The upper conglomeratic interval, corresponding to the final progradational trend recorded by Priabonian deposits, progrades northwards across the whole basin, over a silt and shale interval and seals the Lirou anticline (Fig. 6). Folding affects the Lutetian lacustrine limestones, and the top of the Lutetian is eroded in the crest of the anticline. Variable thickness of Latest Cretaceous to Early Eocene marls suggests a decollement with the underlying monocline Valanginian limestones. The latter is affected by a sub-vertical, NE-trending fault, which displays a left-lateral component of movement (Fig. 8 - Section 3). These observations lead to the following points: i) folding was initiated during Pyrenean shortening, and was still active during deposition of the early Priabonian; ii) folding of the Paleogene sequences is detached above the Valanginian limestone, through a detachment within the Late Cretaceous-Early Eocene marls; iii) folding is related with a NE-trending left lateral strike-slip.

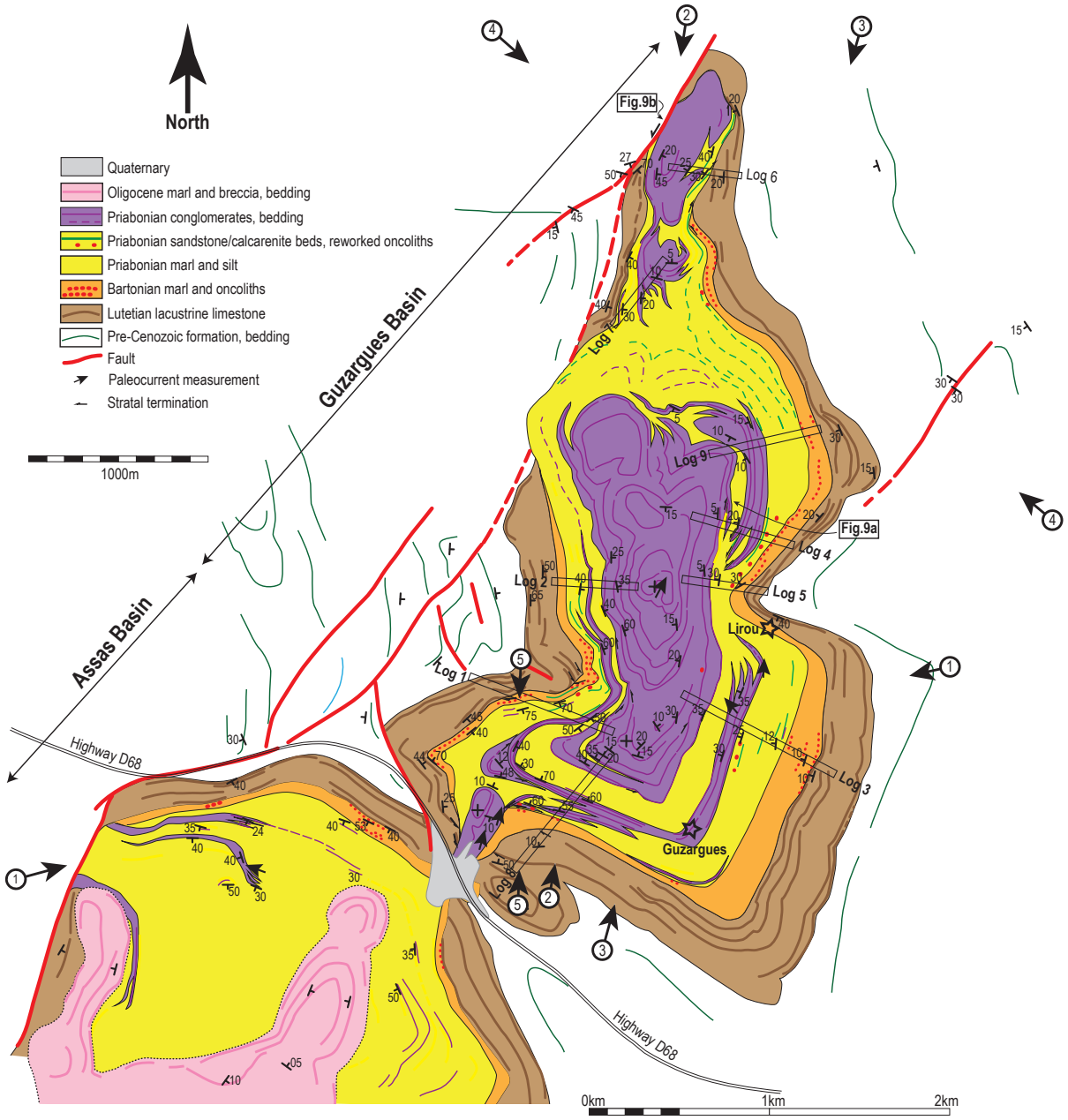
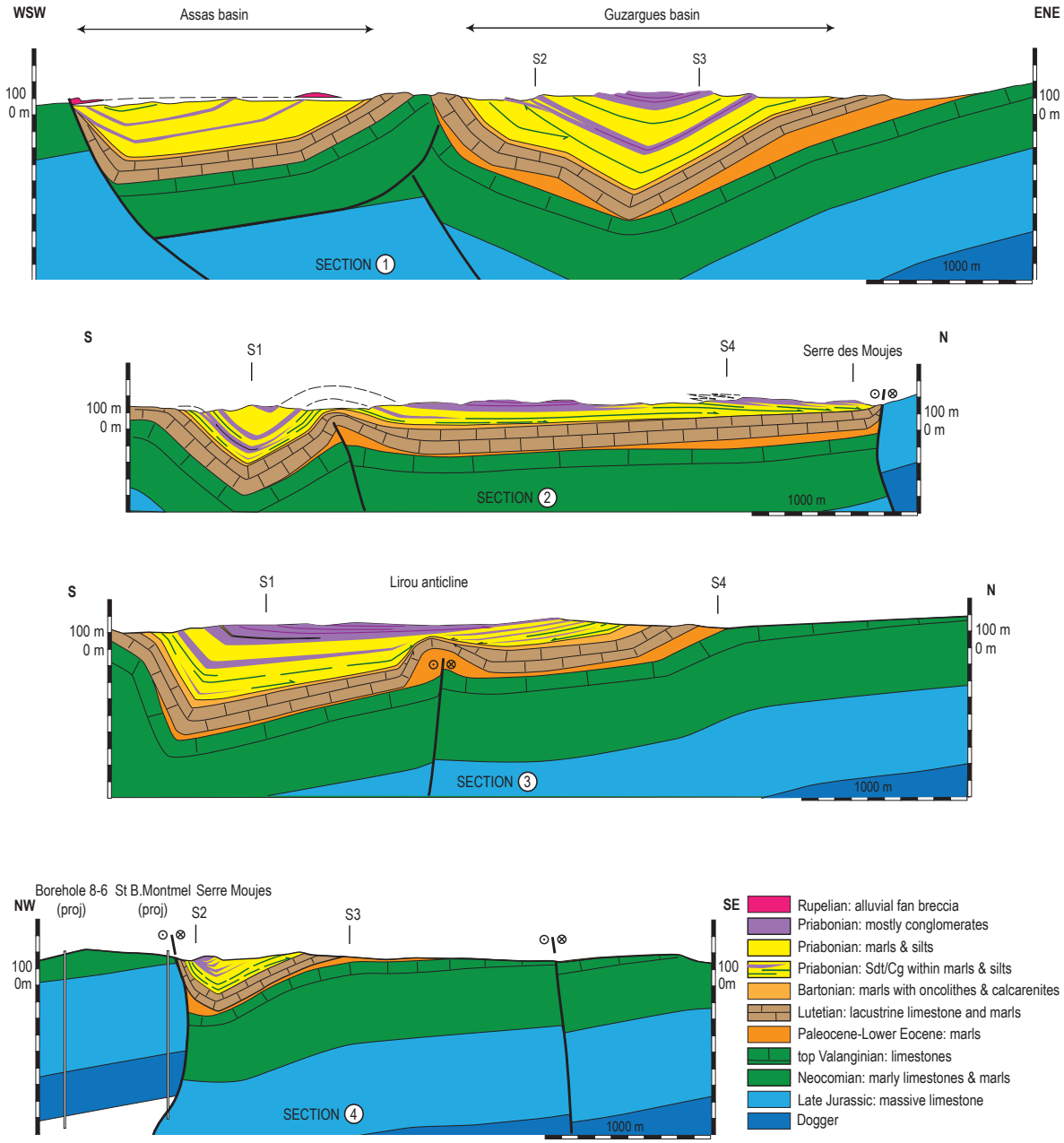


Fig. 7: *Detailed geological and structural map of the Guzargues basin (north of Highway D68, while Assas sub-basin is located south of the highway). Location of the sedimentological logs 1 to 9 reported on Fig. 6 are indicated by fine black rectangles. Position of the structural cross-sections extremities (Fig. 8 and 9) are given by black arrows.*



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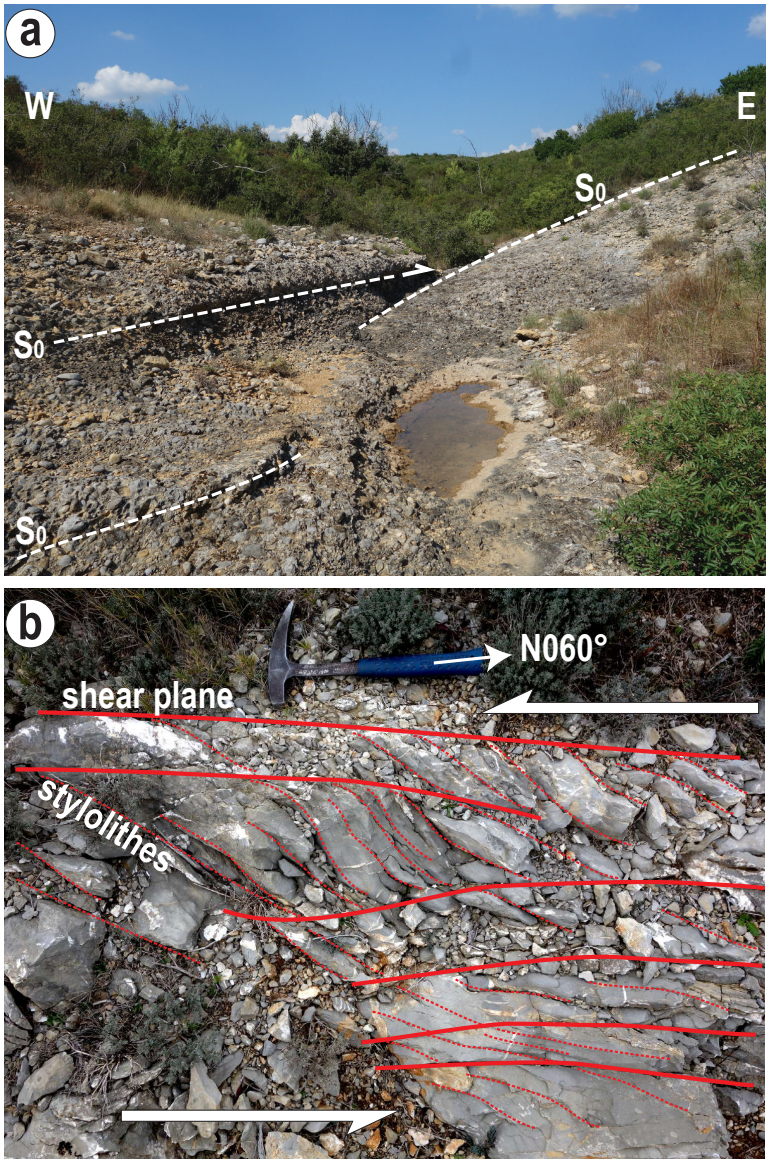
457 ***Fig. 8:*** *Structural sections across the Guzargues basin; see location on Fig. 7.*



458

459 ***Fig. 9: a:*** *internal onlap within the Priabonian conglomerates, which illustrates syndepositional folding*
460 *of the Lirou anticline. Location on map Fig. 7.* ***b:*** *Penetrativve deformation of the Jurassic limestones*
461 *along the basin-bounding fault, indicating left-lateral kinematics. Serre de Moujes, location on maps Fig.*
462 *7.*

463

464 *The southeastern boundary* – Figure 10 shows up to three stacked, angular unconformities,
465 which affect the lower Priabonian sequences along the southern flank of an EW syn-
466 sedimentary syncline. These angular unconformities pass northwards to correlative
467 conformities, while they converge southwards, towards the basin boundary, corresponding to
468 an E-trending anticline of Lutetian limestone. This anticline is locally eroded by a present-day
469 transverse valley, but mapping reveals that sub-horizontal Priabonian conglomerates are
470 exposed within this valley and that they laterally onlap onto the incised carbonates (Fig. 7).
471 Clasts imbrication within the fluvial conglomerates indicates a consistent northwards

paleocurrent in the paleo-valley. Very thin (several tens of metres) in the transverse valley, the Priabonian formations thickens rapidly northward within 1 kilometre distance, from less than 300 m to 450 m, and it exceeds 600 m in the syncline axis. In addition, fluvial conglomerate belts pass laterally and distally (northwards) to silts and marls of alluvial plain facies. These observations clearly show that: i) the syncline was formed during deposition of the Priabonian ; ii) the basin was fed by a north-flowing paleoriver, incised into the growing Lutetian limestone E-W anticline.

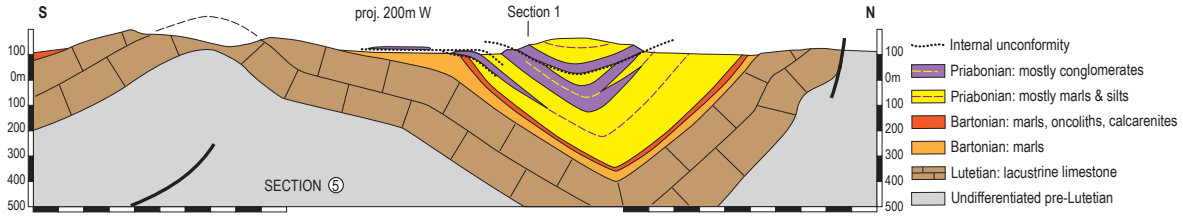


Fig. 10: *Detailed structural section in the south of Guzargues basin showing growth syncline and internal angular unconformities. See location of Section 5 on Fig. 7.*

Serre de Moujes - The northern part of the basin corresponds to a syncline trending N020°, that involves Lutetian limestones and Priabonian conglomerates (Fig. 7). The conglomerates are more developed on the steep western flank, where they unconformably rest over the steeply dipping Lutetian limestones, and they grade eastwards to distal silts and marls (Fig. 8 - Section 4). Along the eastern flank of the syncline, Priabonian formations are onlapping onto the Lutetian limestone; in addition, the Bartonian oncolith formation is transgressed towards the north by the Priabonian. Onlapping surfaces point to a pre-existing synform of Lutetian limestones prior to Priabonian deposition. The conglomerate belt is separated from that of the main part of the basin, suggesting the presence of a smaller, northern, sub-basin with a distinct feeder that supplies clastic sediments across the western boundary (Fig.8 – Section 2). Unfortunately, no clear evidence of paleocurrent were found. North-south correlations across the entire Guzargues basins shows that the northern sub-basin represents the youngest depocentre of the whole basin (Fig. 8 - Section 1). To the NW, Priabonian is directly in contact with the Mesozoic carbonates through a sub-vertical faulted boundary. In the Serre de Moujes, outcrop in the damage zone of the bordering fault reveals sets of associated fractures and dissolution surfaces, consistent with a left-lateral strike-slip motion (Fig. 9b). Unfortunately, no slickensides could be observed to constrain the respective vertical and horizontal components of slip. Saint-Bauzille-de-Montmel borehole (located 1.5 km north, along strike) constrains a curved geometry for this fault and provides evidence for a 700m vertical offset (Fig. 8 - Section 4). Considering the fault geometry and the kinematic observations on the outcrop, a dominantly strike-slip movement along this bordering fault is suggested during Priabonian and the important vertical offset implies several kilometres of cumulated lateral offset. A significant part

of such offset may be inherited from earlier (Pyrenean) tectonic phases. These observations argue for i) deposition of Priabonian syn-tectonic sediments in a hanging-wall syncline, controlled by the activity of the N020°-trending bounding fault, and ii) accommodation of a significant amount a left-lateral strike-slip, by this inherited fault.

4-4. Tectonics - paleoenvironment relationships in the Guzargues basin

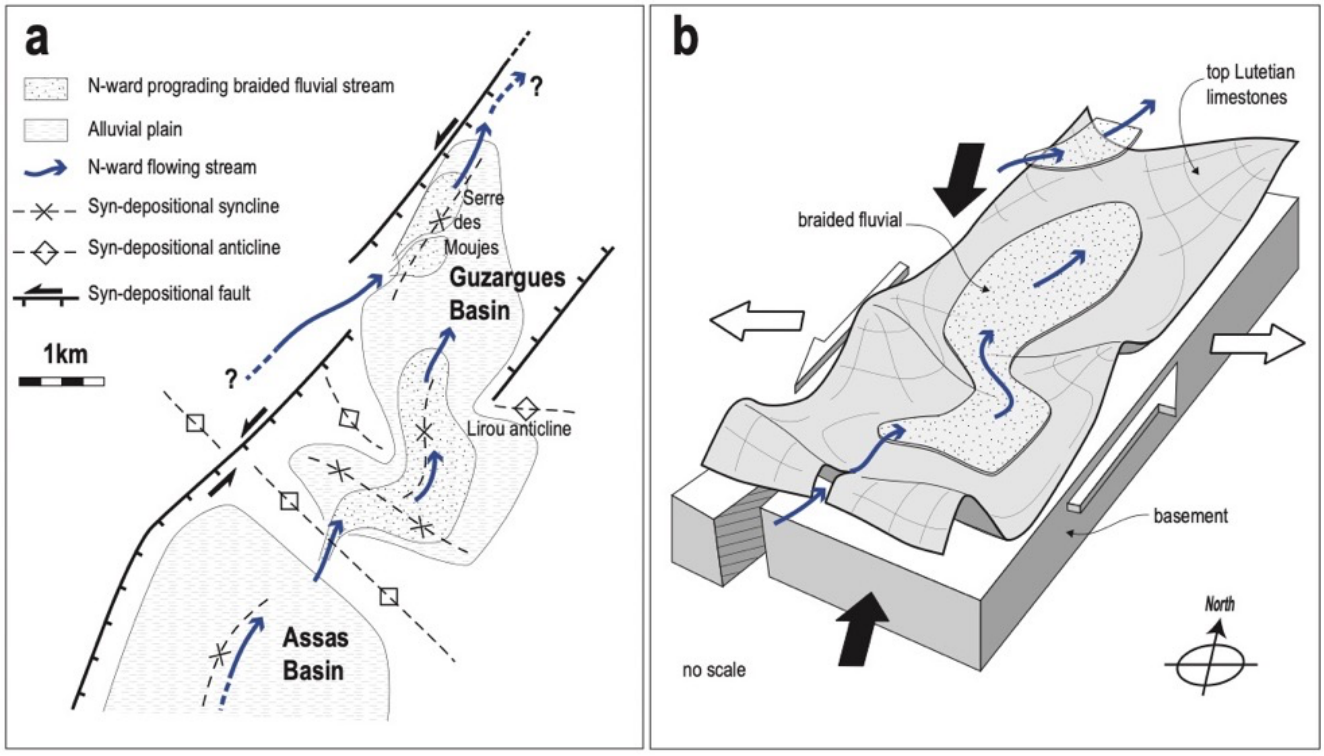


Fig. 11: a: *Sketch map of the paleoenvironments in the Guzargues basin, in relation with the Priabonian tectonics.* ***b:*** *Conceptual tridimensional representation (no scale intended) of the interaction between Priabonian kinematics and sedimentation in the Guzargues basin. Braided fluvial conglomerates accumulate in the most subsiding parts, i.e. the actively deforming synclines that affect the Mesozoic-Lutetian cover, detached over the basement. See text for explanations.*

Figure 11a provides a synthetic view of the paleoenvironment distribution across the Guzargues Basin during Priabonian. Fluvial sediments were delivered through an incised canyon across a growing anticline, which separates Assas (to the south) and Guzargues basins. The belt of northward-flowing fluvial channels preferably occupied the most subsiding parts of the basin: i) in the south, parallel to the bordering growing anticline and, ii) in the overall N-trending axis of the syn-depositional syncline. The latter was controlled by a set of NNE-trending faults which display map- and outcrop-scale evidences of normal faulting and left-lateral strike-slip. The secondary fluvial conglomerates depocenter of Serre de Moujes, in the northern part of the basin, witnesses an additional fluvial input to the basin, across a relay or

528 a jog in the bordering fault system. The syn-depositional Priabonian sequence is therefore
529 controlled by folding of the Mesozoic through the Lutetian cover detached over the Triassic
530 decollement, corresponding to an overall transcurrent, left-lateral kinematics, along the NNE-
531 trending faults (Fig. 11b). The kinematics documented in the Guzargues basin result from NS
532 shortening combined with EW extension,
533
534 **5. Les-Matelles and Saint-Martin-de-Londres Priabonian basins**
535
536 ***5-1. Les-Matelles basin***
537
538 The overall geometry of Les-Matelles basin (LMB) is that of an Oligocene syn-rift hanging-wall
539 syncline, controlled by a NNE-trending normal fault (Fig. 2), known as the Matelles-Corconne
540 Fault (Benedicto et al., 1999). The Priabonian sequence mostly consists of marls and
541 siltstones with rare conglomerate beds. Paleocurrent measurements in the channelised fluvial
542 conglomerates indicate northwest to northeast flowing channels (Egerton, 1996). Detailed
543 mapping (Benedicto, 1996) revealed that the underlying Priabonian sequence also presents
544 growth structures in an asymmetric syncline, that displays a NW limb steeper (≥ 40°) and
545 thinner than the SE limb (Fig. 12). Sequential tectonic restoration of transverse sections
546 indicates an extensional offset across Les-Matelles-Corconne Fault of some 375 m, during
547 Priabonian (Benedicto et al., 1999).
548 In the southernmost part of the basin, Priabonian unconformably overlies the syn-orogenic
549 breccia of Paleocene age, related to the activity of the Montpellier thrust (Andrieux et al., 1971).
550 In the central segment of the basin, Priabonian is conformable over the distal syn-orogenic
551 marls and oncoliths of Bartonian age. Finally, in the northern extension of this basin, Priabonian
552 is unconformable over Neocomian marly limestones, which are deformed by EW-trending
553 compressional structures. Such folds and thrusts are correlated to the Pic-Saint-Loup thrust,
554 active during Bartonian (Philip et al., 1978). Considered at the scale of the basin, the
555 Priabonian is thus unconformable over pre-Pyrenean deformed series, as well as syn-orogenic
556 Pyrenean sedimentation. Finally, the syn-rift Late Rupelian breccia are unconformably
557 overlying the Priabonian (Crochet, 1984).

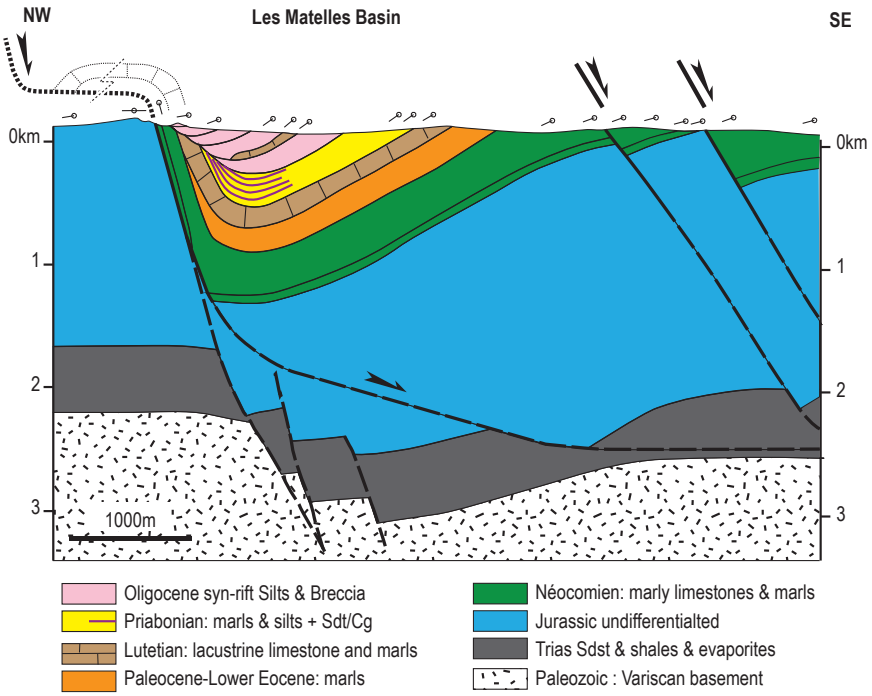
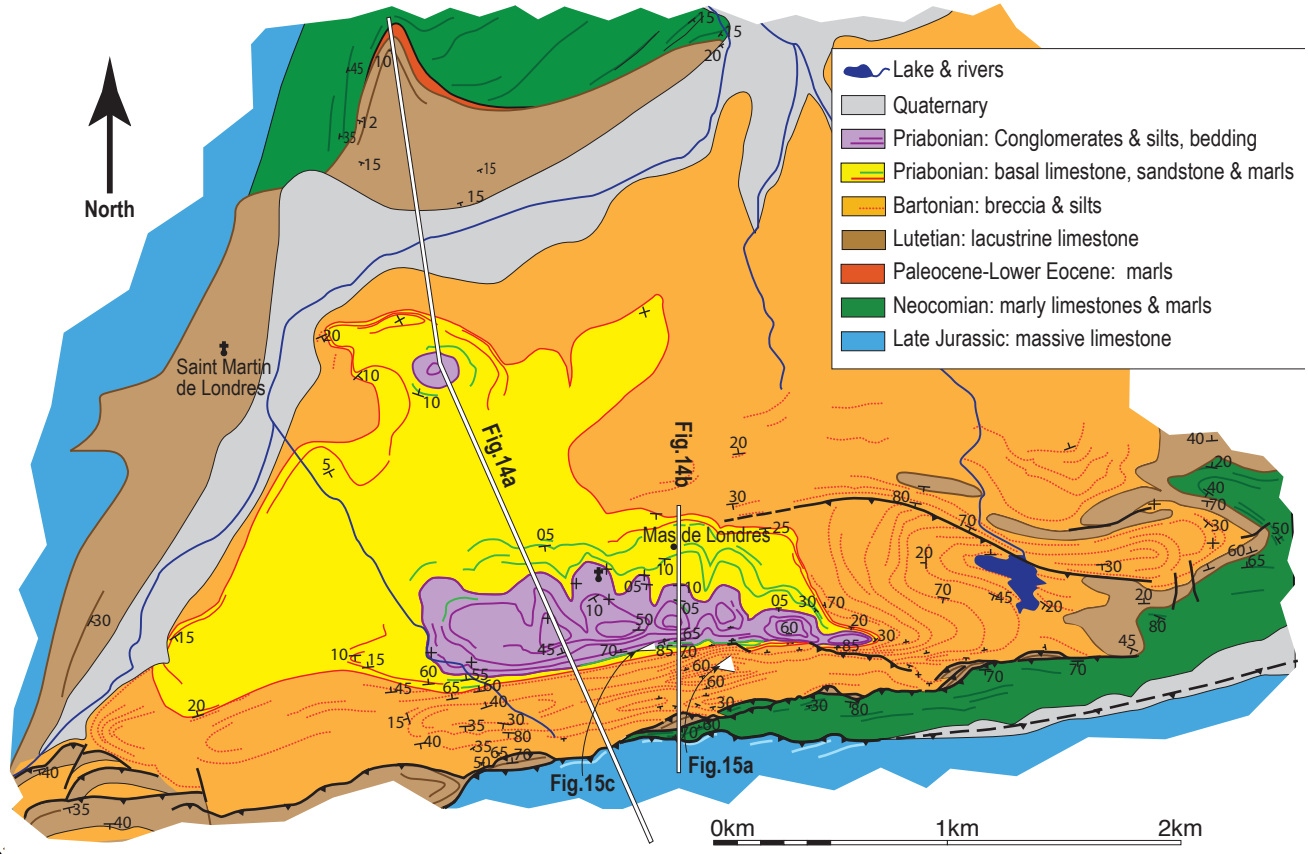


Fig. 12: *Structural section across Les-Matelles basin (from surface data and seismic profile) showing syn-extension deposition of Priabonian units. Modified after (Benedicto et al., 1999).*

5-2. Saint-Martin-de-Londres basin

Saint-Martin-de-Londres basin (SMLB) extends in an EW asymmetric syncline, immediately north of the Pic-Saint-Loup thrust, a structure of the Pyrenean fold and thrust belt (Fig. 2). Above an unconformity over the Neocomian, the basin sequence starts with continental marls and thin beds of lacustrine limestones of latest Cretaceous to Paleocene age (Crochet, 1984; Freytet, 1971; Philip et al., 1978), which are the distal correlative of the syntectonic « *Vitrollian* » breccia, deposited along the Montpellier thrust.

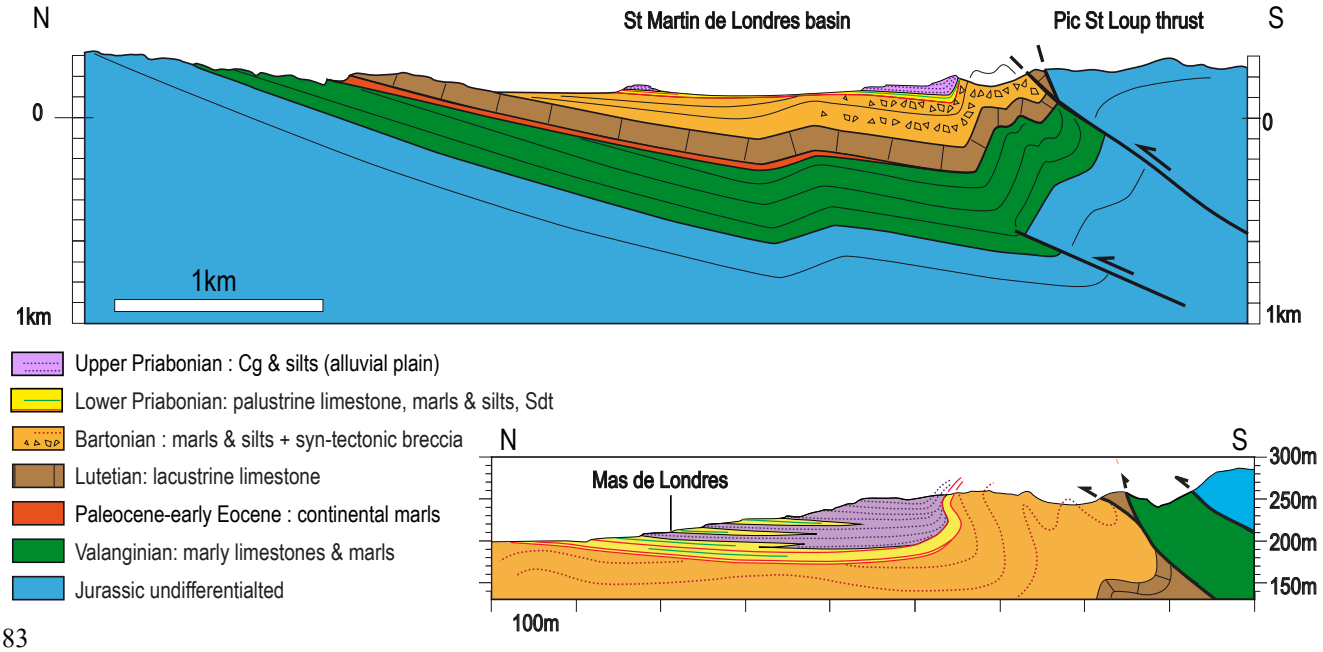
Lutetian lacustrine limestones, transgressive over Late Jurassic to Paleocene formations, form an 80 m to 100 m-thick slab that designs a wide, low amplitude, EW-oriented syncline (Fig. 13). The overlying breccia, dated to the Bartonian (Philip et al., 1978), consists of local, late Jurassic, Neocomian and Lutetian limestones, derived from erosion of the active Pic-Saint-Loup thrust front. They pass northwards to distal continental marls and silts, which present syn-tectonic growth structures. The southern boundary of the basin has been intensely folded and faulted in relation with the Pyrenean Pic-Saint-Loup thrust, during Bartonian time (Fig.14). The Priabonian sequence unconformably overlies the Bartonian breccia; to the east, it seals an intra-Bartonian thrust propagation fold (Fig.13), which requires erosion of the Pyrenean-related structure prior to Priabonian deposition.



580

581 ***Fig. 13:*** *Detailed geological map of Saint-Martin-de-Londres basin. Position of sections Fig. 14 indicated*

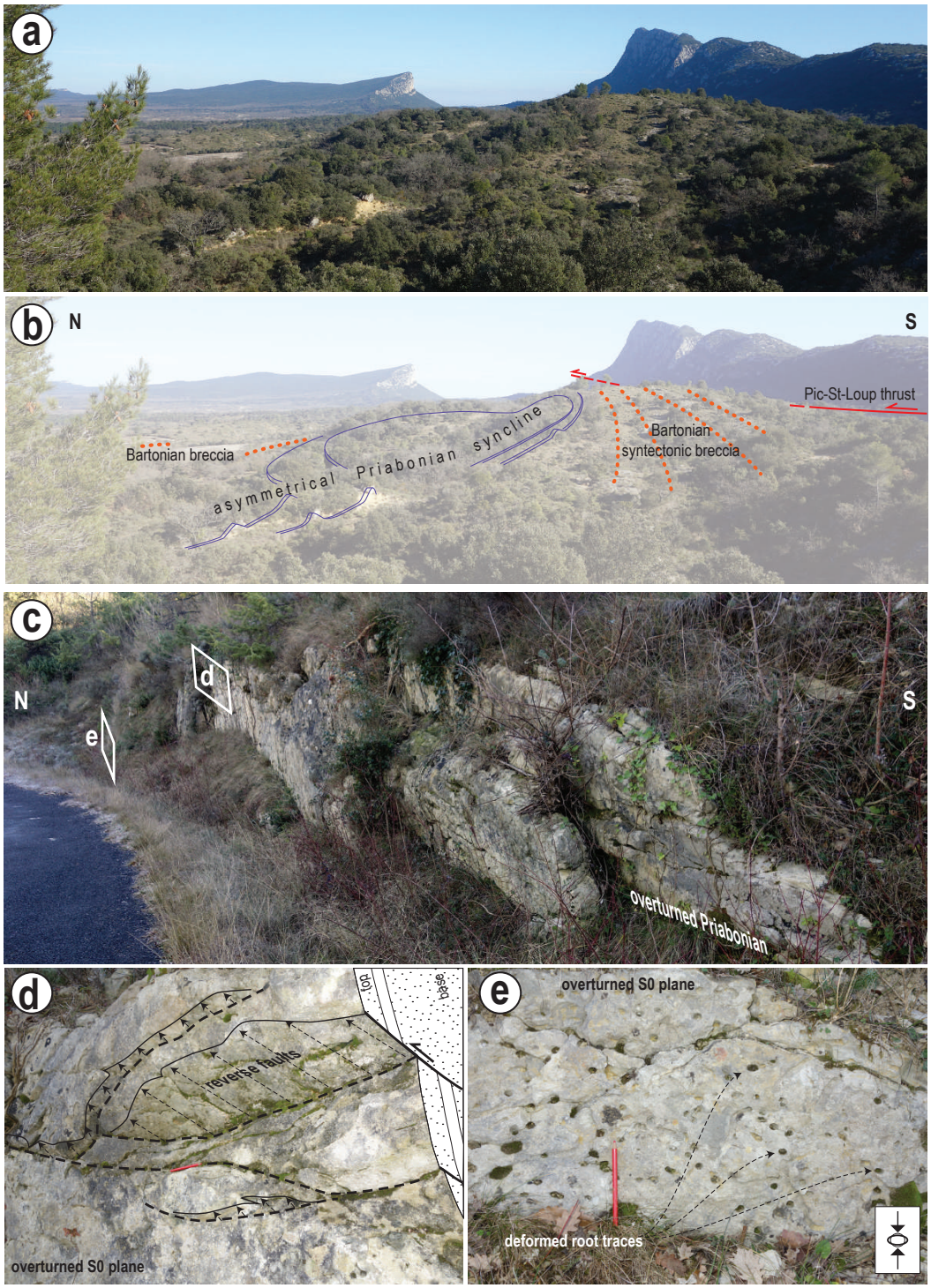
582 *as white line. The black dot and cross points to the churches of the villages.*



583

584 ***Fig. 14:*** *Structural section across St Martin-de-Londres basin:* ***a*** *- general ;* ***b*** *– detailed, showing syn-*

585 *depositional compression during Priabonian.*



586

587 ***Fig. 15:*** ***a*** *and* ***b:*** *general view of the eastern end of St-Martin-de-Londres Priabonian basin, an*
588 *asymmetrical growth syncline, unconformable over the synorogenic Bartonian breccia, which were*
589 *generated along the pyrenean Pic-St-Loup thrust.* ***c:*** *overturned Priabonian sandstone along the*
590 *southern limb of the asymmetric syncline.* ***d;*** *detail of the previous outcrop displaying small scale reverse*
591 *faults that provide evidence for N-S compressive deformation. E: deformed root traces, evidence for*
592 *pre-lithification N-S compressive deformation of the Priabonian sandstone bed, prior to its folding.*
593 *Outcrop location on Fig. 13.*

594

595 Along the southern margin of the SMLB, the Priabonian sequence designs an asymmetric
596 syncline with sub-vertical to overturned southern limb (Fig. 15a,b.). The sequence starts with
597 a thin level of finely laminated limestone that contains bioclasts and charophytes characterising
598 a lacustrine / palustrine environment. Laterally, the limestone passes to thicker and more
599 massive beds of lacustrine mudstone. The thin basal limestone is correlated to the north of the
600 basin, with 10 m to 15 m thick marls and carbonates deposited in an arid, evaporitic
601 environment. Above the limestones, along the central southern margin only, are found grey
602 marls including several thin beds of glauconitic sandstones with bioclasts and rare foraminifera
603 (Egerton, 1996). This shallow marine sandstone passes upwards to marls and siltsones
604 including thin bioturbated sandstone beds, deposited in a floodplain, similar to those described
605 in LMB. In the upright, southern limb of the syncline, these sandstone beds are affected by
606 small scale reverse faults (Fig.15c). They also display soft penetrative deformation of initially
607 cylindrical root traces (Fig. 15d), which indicate an early, pre-lithification, N-S shortening prior
608 to folding. Finally, up to 50 m stacked, channelized conglomerates, are deposited in several-
609 metres-thick levels, interbedded with siltstones and marls. The high ratio of channelized
610 conglomerates *vs* floodplain marls and siltstones is more in line with Guzargues than with Les-
611 Matelles basins. Conglomerates are clast-supported and rather well-sorted. Clasts are
612 polygenic, including significant number of quartz veins and black cherts, derived from the
613 Paleozoic deformed basement. Distinctive clasts of glauconitic sandstones and calcarenites
614 containing Albian orbitolines cannot be related to any local source, as the Mesozoic sequence
615 is interrupted above the Valanginian due to the regional erosion of the Durancian Isthmus
616 (Husson, 2013). Such clasts, also been identified in Les-Matelles and Montferrier Priabonian
617 outcrops are imbricated under the influences of north to northwestwards paleocurrents
618 (Egerton, 1996), and thus correspond to a southerly derived material (Freytet, 1971).
619 The Priabonian sequence of SMLB unconformable over Pyrenean structures, and containing
620 clasts supplied by a southerly derived sedimentation was deposited in a N-S compressional
621 setting, related to reactivation of the Pyrenean Pic-Saint-Loup Thrust.
622

623 **6. Seismic sequence analysis of the Hérault Basin**

624
625 The Hérault Basin (HB) is a NNE-SSW asymmetric half-graben developed along the Cévennes
626 Fault, formed in relation with the rifting of the Gulf of Lion (Arthaud et al., 1981; Maerten and
627 Séranne, 1995; Séranne, 1999). It covers the western part of the Pyrenean Montpellier thrust
628 (Fig. 2) and is filled by up to 1500m-thick succession of rift and post-rift sequences. The syn-
629 rift and earlier Pyrenean structures are therefore mostly covered, which lead us to investigate
630 the Pyrenean to syn-rift transition through subsurface data.
631 The basin sequence consists of (i) Rupelian to Aquitanian, continental and lacustrine

formations, and (ii) Burdigalian to Langhian, marine to fluvial post-rift deposits, forming syn-rift and post-rift sequences separated by a major unconformity. The pre-rift substratum consists of Permian to Jurassic sequence, as well as mid-Cretaceous bauxite deposits, which have been sampled by several boreholes (Berger et al., 1981; Lajoinie and Laville, 1979). In addition, syn-tectonic Pyrenean Paleocene to Eocene formations (poorly exposed along the eastern margin of the basin) and Mesozoic carbonates deformed during Pyrenean compression can be extrapolated beneath the basin, and several Pyrenean thrusts repeating parts of the Jurassic sequence are documented (Alabouvette et al., 1982; Berger et al., 1981).

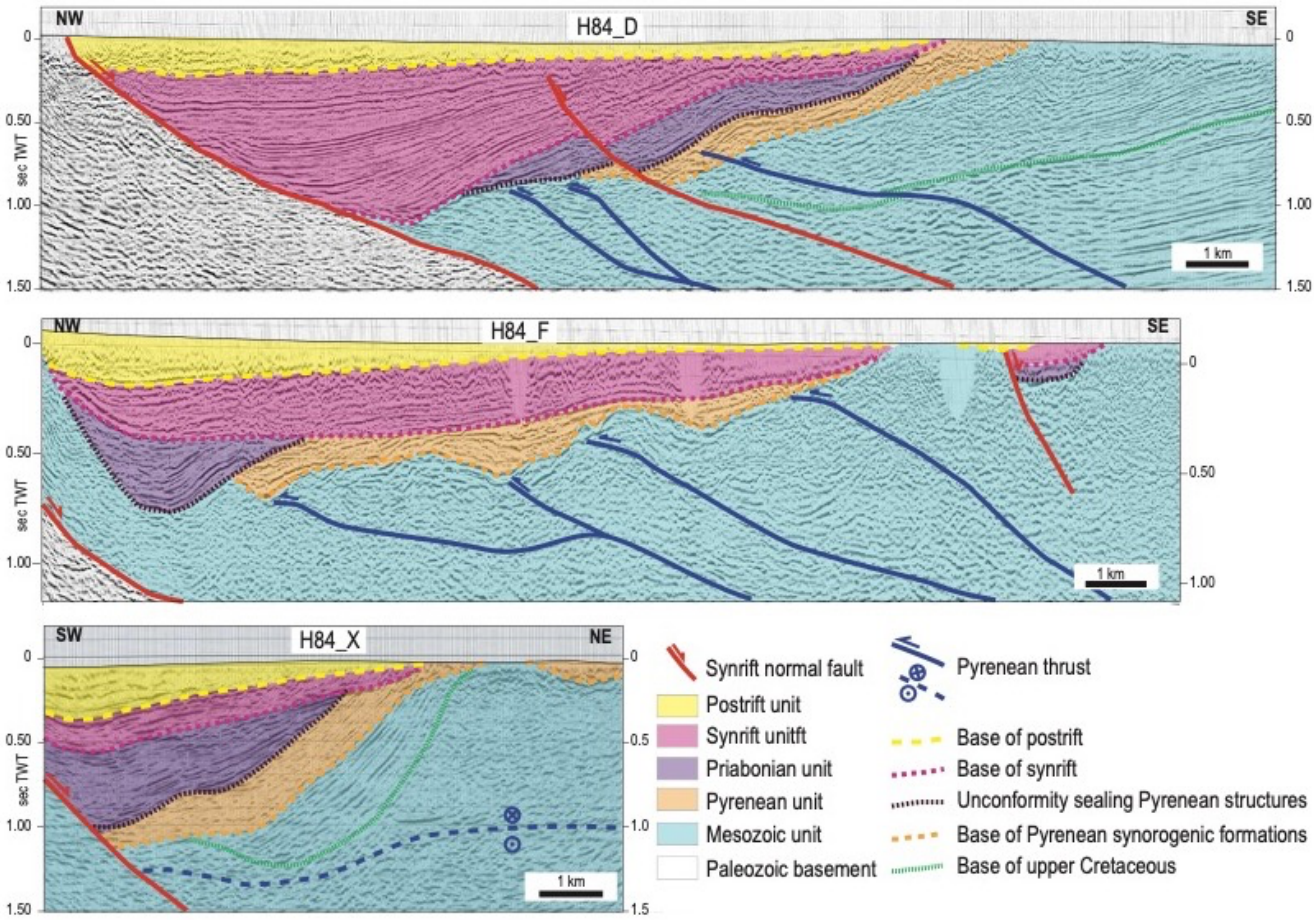
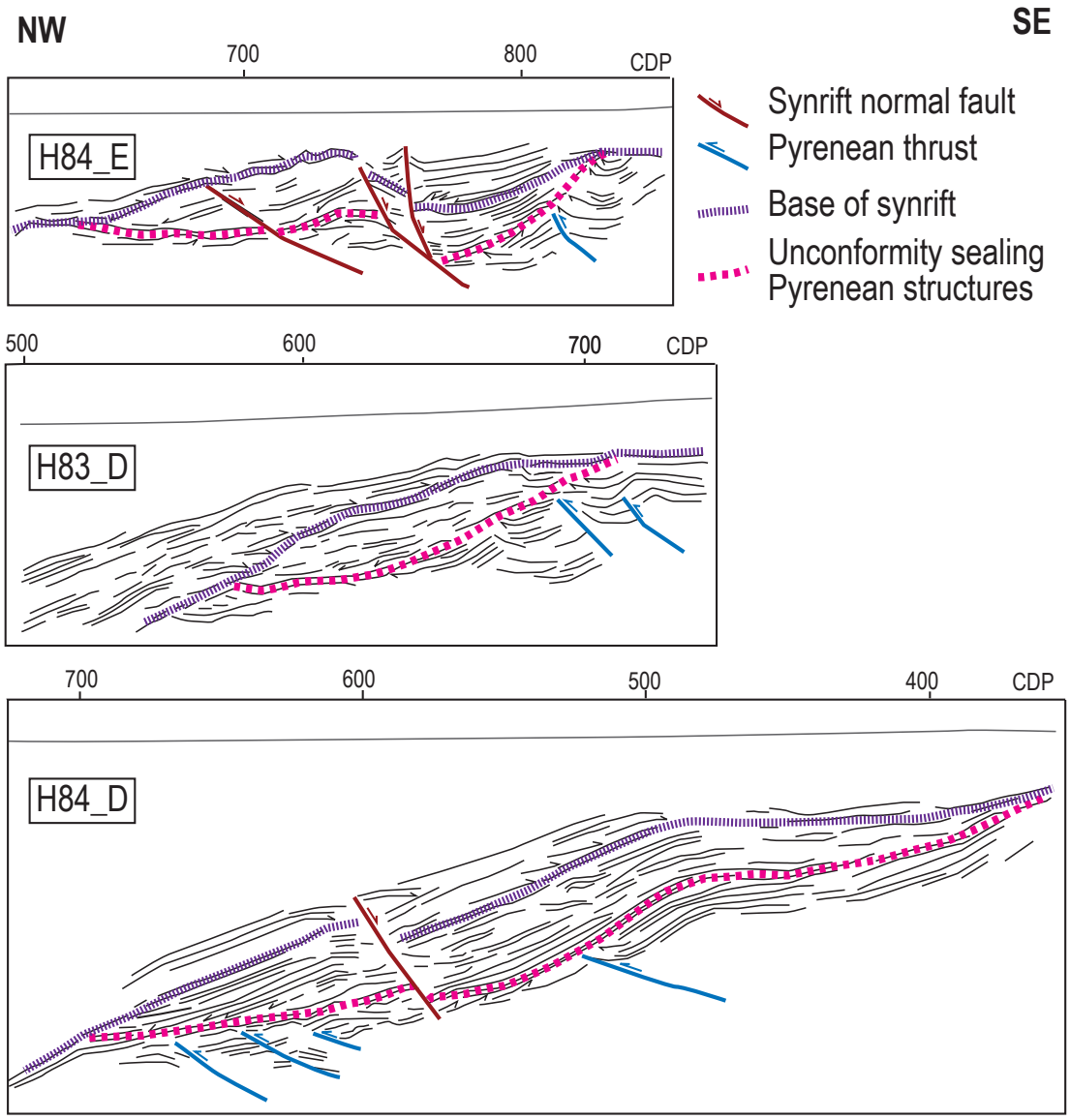


Fig. 16: *Seismic profiles representative of the Hérault basin (H84D, H84F, H84X). The purple sequence interpreted as Priabonian (see discussion in text) is intermediate between the Pyrenean synorogenic units and the Oligocene syn-rift sequence. See location on Fig. 2 and 18.*

Two geophysical surveys acquired in 1983 and 1984, then reprocessed in 2005-2008 (Serrano and Hanot, 2005), provide a regular seismic coverage (Fig. 2) of sufficient quality to examine the seismic units underlying the syn-rift sequence (Fig. 16). Some of these seismic sections were already interpreted by (Maerten and Séranne, 1995) who focused their study on the geometry of syn- and post-rift units.



650

651 ***Fig. 17:*** *Line drawings of parts of selected seismic profiles in the Hérault basin showing the lower and*
652 *upper erosional boundaries of the intermediate sequence. See location on Fig.18.*

653

654 Reinterpretation of the seismic profiles allows us to identify, in several parts of the basin, a
655 distinctive seismic unit below the syn-rift deposits (Fig.16). It is bounded by angular
656 unconformities both at the top and at the base and displays significant thickness variations and
657 complex internal geometries. Geometrical relationships with underlying and overlying seismic
658 sequences define its relative stratigraphy: (i) It is truncated by typical syn-rift sequences and
659 is affected by the extensional syn-rift faults; (ii) It unconformably rests on pre-rift units affected
660 by Pyrenean reverse faults (Fig. 17). This intermediate seismic unit, undrilled by any
661 boreholes, does not correlate with surface exposures. However, comparison with field
662 observations in the GB, SMLB and LMB point out a similar structural and stratigraphic position
663 suggestive of a Priabonian age. Mapping of this intermediate interval, of proposed Priabonian

age, across the Hérault basin shows a discontinuous sequence including two main depocentres (Fig. 18 a). One depocentre is located where the Cévennes Fault displays a significant change from an overall NE-SE direction to a SSE-trending orientation. H84X profile (Fig.16) shows evidence for Priabonian syn-depositional, dip-slip movement across the SSE-trending fault segment. Such a geometry suggests a left-lateral movement along the Cévennes fault, associated with a dip-slip across the SSE segment which represents a releasing bend (e.g. (Christie-Blick and Biddle, 1985; Cunningham and Mann, 2007) of the master fault. A second depocentre is located along the SE margin of the Hérault basin, controlled by NE-trending faults that offset the earlier Pyrenean Montpellier thrust.

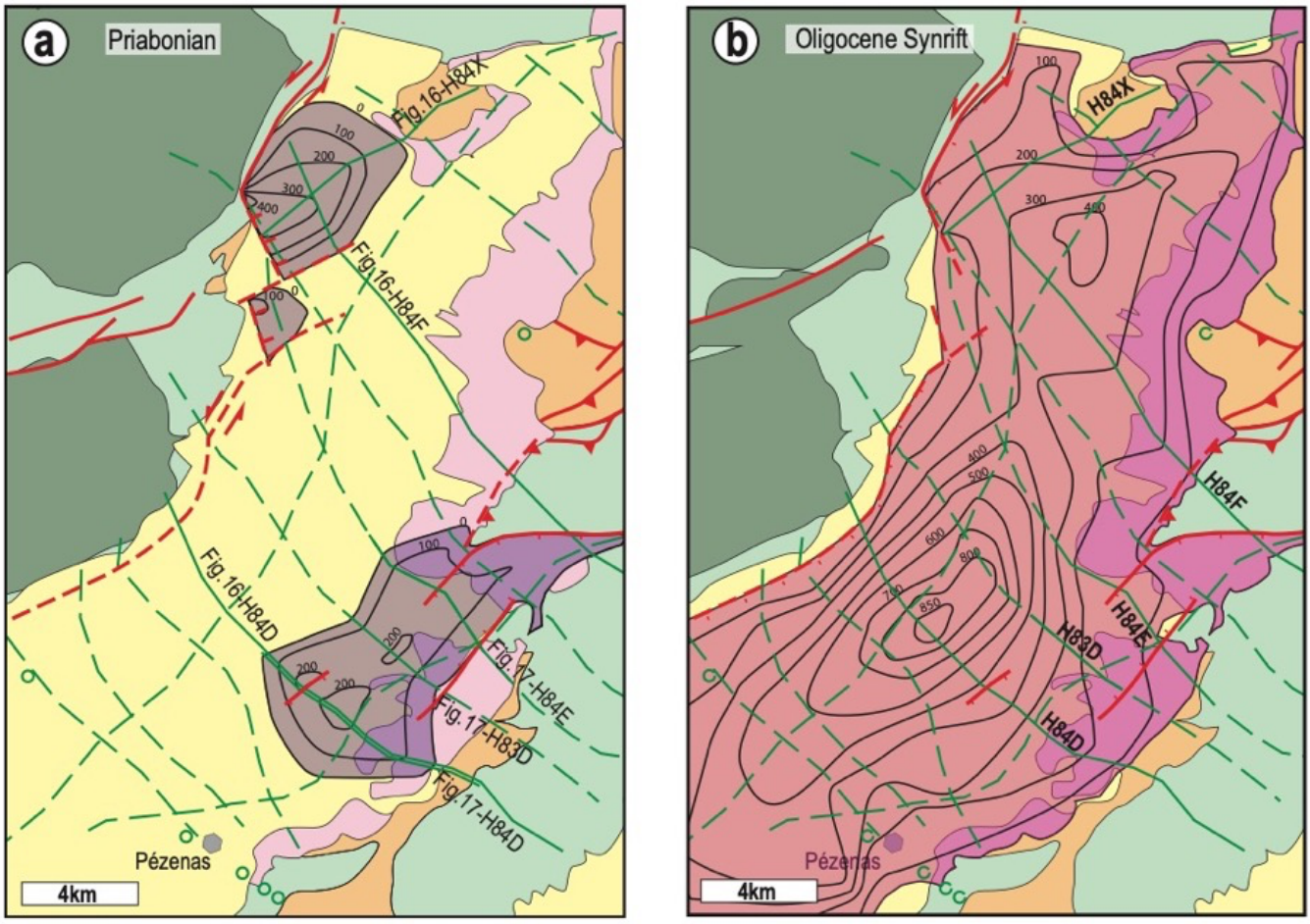
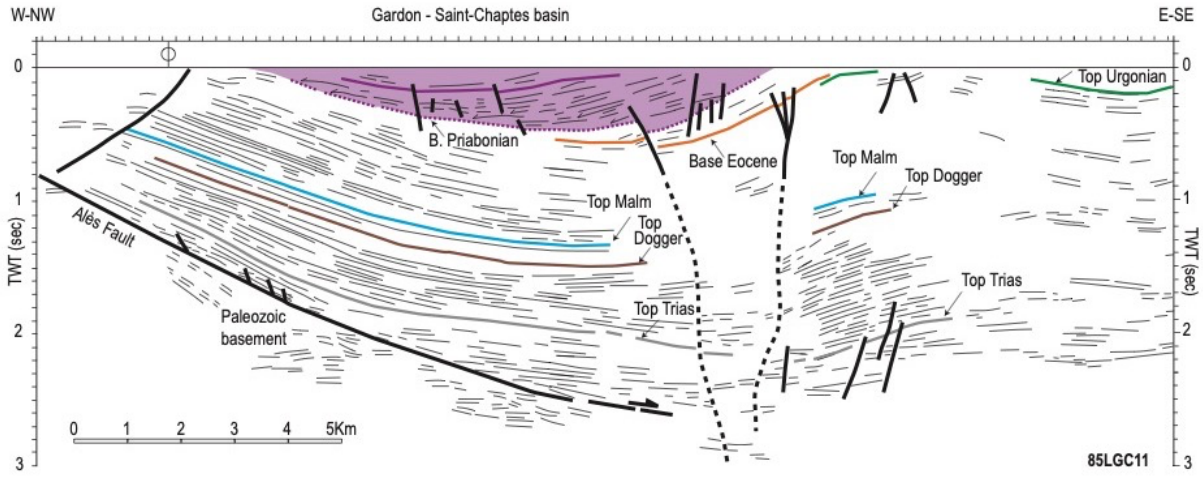


Fig. 18: *Isopach maps of Priabonian* ***(a)*** *and of syn-rift (**b**) sequences in the Hérault basin, superimposed onto a simplified geological map (same legend as in Fig. 2, and location in Fig. 2). Contouring of depocenters in metres. The seismic profiles used in the study are represented in stippled green lines; plain green lines with labels correspond to the profiles shown in Fig. 16 and 17. Faults are represented in red.*

Isobaths of the base syn-rift (Fig. 18 b) depict the depocentre of the syn- and post-rift Oligo-Miocene basin. Comparing the two maps (Fig. 18 a and b) clearly shows that the successive sequences were controlled by two distinct tectonics settings and kinematics. The distribution

683 and NE-SW overall orientation of the Oligo-Miocene depocentre, and the presence of a
684 transfer zone (roughly parallel to H83E profile) are consistent with NW-SE extension across
685 the Cevennes-fault (Maerten and Séranne, 1995). Our data do not support such a transfer
686 zone active during the Priabonian.
687
688 **7. Seismic sequence analysis in the Gardon - Saint-Chaptes basin**
689
690 The Priabonian Gardon - Saint-Chaptes basin (GSCB) sits at the junction of an almost
691 continuous NE-trending strip of Priabonian outcrops from the foreland of the Montpellier thrust
692 (Fig. 2), to the large syn-rift Alès basin, parallel to the NE-trending Cévennes Fault. Syn-rift
693 Oligocene sediments are observed, unconformable over the Priabonian, in the Alès Basins
694 (Sanchis and Séranne, 2000) and in a distinct outcrop in the southern part of the GSCB
695 (Lettéron et al., 2018). Transgression of the post-rift Burdigalian tidal sequences, restricted to
696 the E-W paleo-ria of Uzès basin (Reynaud et al., 2006), did not reached the Alès basin. The
697 Gardon segment, between Alès and Saint-Chaptes, consists of a large asymmetrical syncline,
698 with a distinctive NW-trend, contrasting with the general NE-trend of the contemporaneous
699 basins (Fig. 2).
700 Pyrenean structures in this area consist of a succession of E-W folds affecting the thick
701 Triassic to Neocomian sequence (Arthaud and Laurent, 1995), dated with syn-tectonic
702 Paleocene to Bartonien terrigenous sediments in the syncline axes. The later are buried to the
703 west beneath Priabonian and Oligocene sequences, which provides a first-order evidence for
704 a major post-Pyrenean discontinuity (Fig.2). Detailed mapping of the northeastern border of
705 the GSCB also defines a marked angular unconformity of the Priabonian over the syn-
706 pyrenean breccia sequences (Sanchis, 2000).



707
708 ***Fig.19:*** *Interpreted line drawing of a seismic reflection profile across the Gardon Basin showing the syn-*
709 *depositional migration of the hanging-wall syncline. Modified after (Sanchis, 2000). Main markers are*

identified; Priabonian interval in purple overlay; intra-Priabonian marker corresponds to the base of "Grès de Celas" Formation. Location on Fig. 2 and 21.

Chronostratigraphy is rather well constrained due to numerous mammalian fossils (Fig. 3) reviewed in (Lettéron et al., 2018). Priabonian deposits form a wide asymmetric syncline, including a thick NE and reduced SW limbs, respectively. Seismic reflection survey of the Alès basin included a southern profile (85LGC11), which reveals the structure and mode of formation of this syncline (Sanchis, 2000) (Fig. 19). In section, the GSCB displays Priabonian growth syncline migrating WNW-ward, which characterises dip-slip movement of a hanging-wall flat made of Mesozoic sequence detached above an eastward, low-dipping, decollement, above the Paleozoic basement. Such low-dipping decollement, the "Alès Fault", has been identified in the neighbouring Alès basin and beneath the Mesozoic sequence, folded during the Pyrenean (Benedicto, 1996; Sanchis, 2000; Sanchis and Séranne, 2000). This decollement at the basement-cover interface is distinct from the Cévennes basement fault, although both the Cévennes and Alès faults emerge along the same trace. This seismic profile (Fig. 19) also provides evidence for 4km E-SE migration of the depocentre, therefore suggesting similar displacement of the hanging-wall across the Alès Fault, in this direction, during Priabonian.

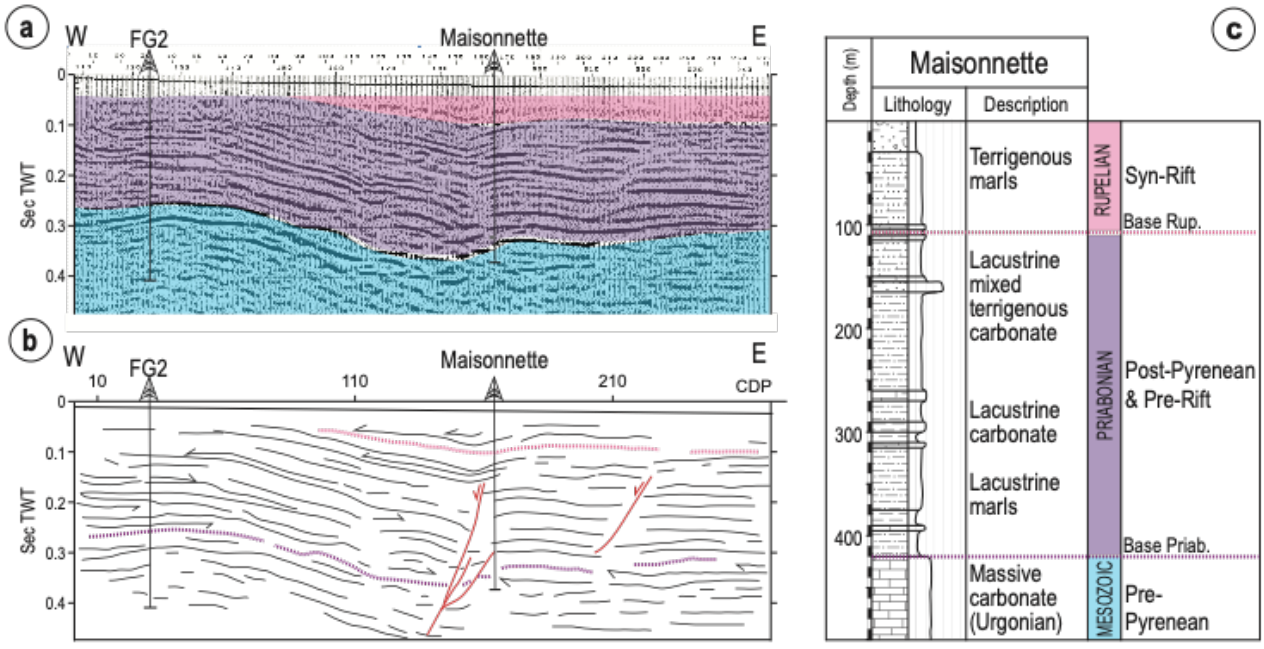


Fig. 20: *Seismic reflection profile and corresponding line drawing in the Gardon - Saint-Chaptes basin showing evidence for syn-depositional (Priabonian) extensional faulting sealed by the Rupelian sequence. Maisonnette borehole allows to interpret and to date the sequence. Modified after (Baral, 2018). Location on Fig. 2 and 21.*

The Saint-Chaptes part of the GSCB has been investigated by seismic reflection profiling, aiming at hydrogeological exploration of the covered Barremian *Urgonian* karst (Josnin, 2001). The sections tied to 4 wells sampling the Oligocene and Priabonian interval (Fig. 2) (Baral, 2018; Baral et al., 2018) and correlated with surface exposures providing sedimentology and stratigraphy controls (Lettéron et al., 2018), allowed to identify, date and map the major seismic markers across the basin (Fig. 20). In borehole, the Priabonian sequence consists of lacustrine marls and carbonates, that pass upward to sandstones. It rests on a thin interval of undated detrital continental sediment, which can be tentatively correlated with the syn-tectonic Bartonian observed in outcrops north of the borehole. Priabonian displays growth structures generated by extensional faults which ramp down into the Barremian (*Urgonian* facies) massive limestones. Such faults belong to the N-NE trending fault-network that truncates the Pyrenean folds, and is sealed by the Rupelian syn-rift sequence (Fig. 2). The dip-slip component of such faults observed in seismic, is not associated with any significant horizontal displacement, since the pyrenean fold axes (Berger et al., 1978) are not laterally offset (Fig. 2).

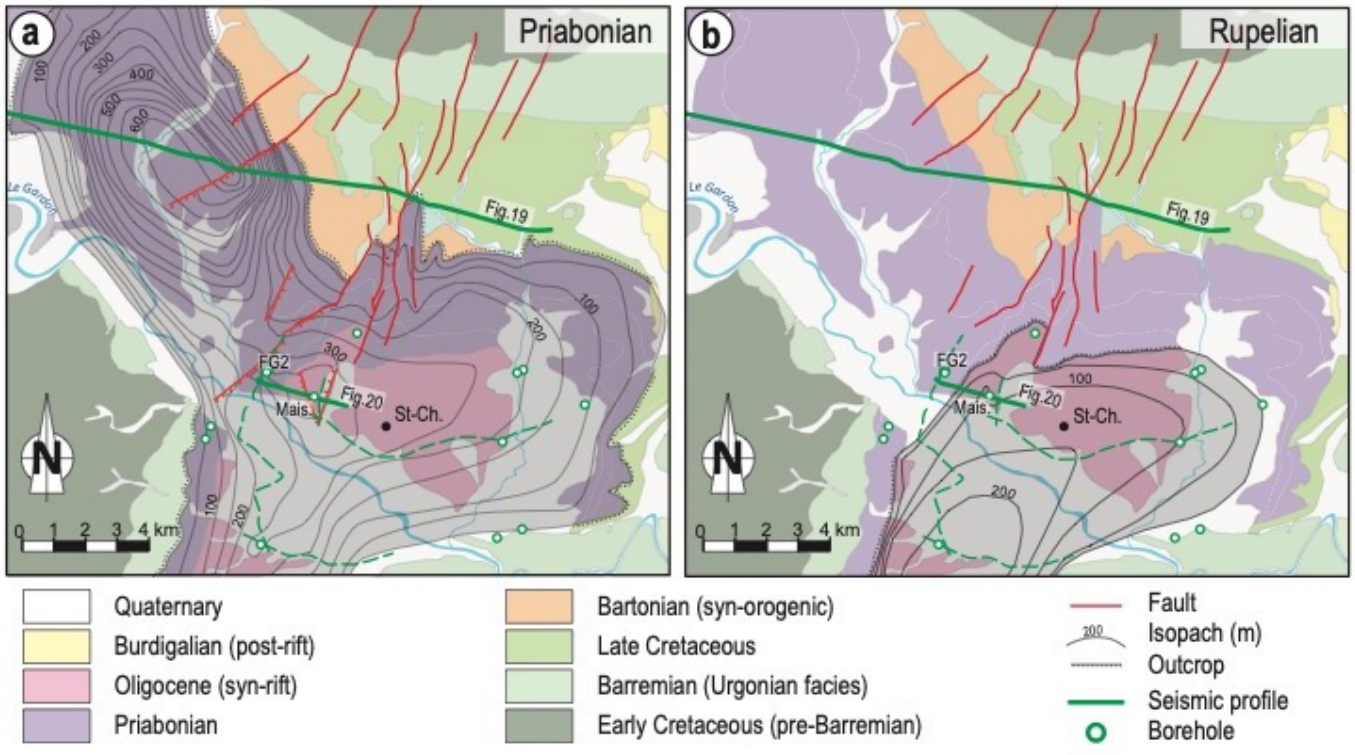


Fig. 21: *Isopach maps of Priabonian* ***(a)*** *and Rupelian* ***(b)*** *sequences in the Gardon – Saint-Chaptes basin. Note the different orientation of the Rupelian depocentre axis with respect to the previous Priabonian subsidence axis. Location in Fig. 2.*

Time-depth conversion was achieved using velocities stack provided with the seismic profiles and controlled with the borehole data (Baral, 2018), in order to make an isopach maps of the

Priabonian depocentre Fig. 21a). It provides evidence for Priabonian tectonics across N-NE-trending active faults, observed north of the basin, and southward, parallel to the Sommières basin (Fig. 2). The Gardon segment of the depocentre, although less well constrained by subsurface data, designs the NNW-trending asymmetrical syncline imaged in seismic (Fig. 19). The overlying Oligocene (syn-rift) depocentre (Fig.21b) is restricted to the Saint-Chaptes part of the GSCB and it is not affected by the earlier faults, nor controlled by any active boundary faults. It designs a wide depocentre, whose subsidence axis strikes N050, oblique with respect with the Priabonian depocentre. This suggests a Rupelian syn-rift basin formed as a hanging-wall syncline (Benedicto et al., 1999) above a detachment, which we interpret to be the Paleozoic basement-Mesozoic cover interface, i.e. the Alès Fault in this area. The structural relationships in the GSCB and the contrasted orientation of the Priabonian and Rupelian depocentres indicate two distinct events with different extension directions.

8. Regional synthesis and discussion

Field observations show that the GB, LMB and SMLB display similar Paleogene successions and stratigraphic settings. In all exposed basins, the syn-tectonic pyrenean continental formations consist of breccia along the active thrusts that pass distally, i.e. northwards, to fine continental silts and marls, or lacustrine/palustrine limestones. They are overlain by Priabonian continental sediments, which were deposited in alluvial floodplains, including channelized polygenic, sub-rounded, conglomerates. Proximal-distal facies distributions, paleocurrents and some distinctive clast lithologies are consistent with a southern source for the clastic sediments. This is related with a remaining pyrenean topography that extended in the present-day Gulf of Lion, until the Priabonian. Priabonian formations are preserved in distinct basins or sub-basins corresponding to subsiding depocentres, which display significant thickness variations and syn-tectonic growth structures. The lower boundary is an unconformity over the pyrenean syn-orogenic formations and structures, although it can locally be observed conformable at outcrop scale. The angular unconformity is better expressed where pyrenean tectonics was more developed: i.e. close to the Montpellier and Pic-Saint-Loup thrusts and around the anticlines. Priabonian sequences are eroded and unconformably overlain by Late Rupelian continental alluvial fans and lacustrine deposits, in relation with the rifting of the Gulf of Lion margin, except in the SMLB, where the stratigraphic record ends with Priabonian. The Priabonian sequences of the GSCB, observed in seismic profile and tied to borehole data, and those from the HB (although not precisely dated) provides evidence for similar structural and stratigraphic relationships, as observed in the field. Pyrenean structures and syn-orogenic sequences are unconformably overlain by a seismic sequence of highly variable thickness, which is also truncated by younger sequences displaying syn-rift growth structures against the

basin bounding normal faults. We correlate this intermediate seismic sequence with the Priabonian interval described in the field in GB, LMB and SMLB.

At variance with the observed similar structural relationship with underlying and overlying sequences, the Priabonian deposits display contrasting structures in the different basins investigated (Fig. 22):

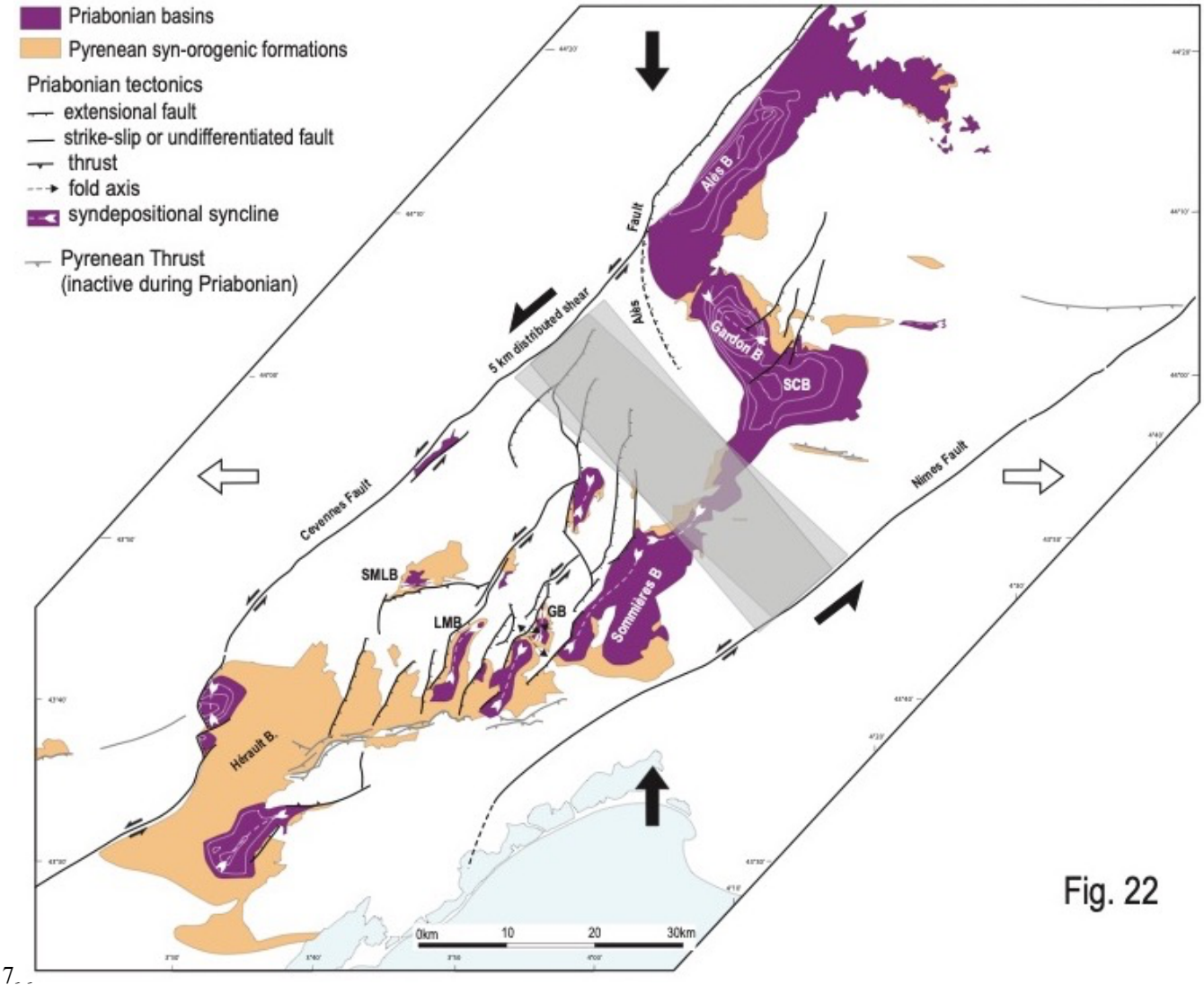


Fig. 22: *Synthetic regional map of Priabonian structures and kinematics in Languedoc. For location of specific features and localities, please refer to Fig. 2, which shares the same frame. The study area corresponds to a zone of sinistral shearing between the Cévennes and Nîmes faults, with a minimal value of 5 km (grey overlay symbolises the simple shear).*

- LMB: Syn-tectonic growth-structures are consistent with a hanging-wall syncline, resulting from extensional faulting along the NNE-striking bounding normal fault.
- SMLB: Structures are related to syn-depositional compressional folding across the E-trending Pic-Saint-Loup thrust.

- GB: The tectonic sketch derived from the detailed geological map Fig.7 displays syn-depositional folding, which characterises both NE-SW compression and EW extension; in addition, there is evidence of left-lateral strike-slip along NNE-trending faults (Fig. 11b).
- HB: Depocentre distribution with respect with controlling faults suggests left-lateral motion along the NNE-trending Cévennes Fault, which accommodates normal faulting across N-trending splays (Fig. 22).
- GSCB: The Priabonian sequence was deposited as a hanging-wall syncline controlled by a SE-trending extensional low-angle decollement of the Mesozoic cover over the Paleozoic basement.

Such apparently contradicting kinematics can be understood in an overall left-lateral deformation along the regional NNE-trending faults, such as the Cevennes, Les-Matelles-Corconne and the Nîmes faults (Fig.22). When submitted to a stress regime characterised by a horizontal, N-S, maximum principal stress $\sigma 1$ and a horizontal E-W minimum principal stress $\sigma 3$, inherited faults oriented NNE are reactivated as left-lateral strike-slip faults, with restraining bends (Christie-Blick and Biddle, 1985) along more northerly oriented segments, as in LMB, HB, GSCB. East-west oriented fault splays such as the Pic-Saint-Loup correspond to restraining bends. Syn-tectonic deposition in the syncline north of Uzès (Fig.2, Fig.22) also provide additional evidence for N-S shortening across EW oriented structures, during Priabonian.

The amount of left-lateral displacement during Priabonian is poorly constrained, as it reactivates, and is superimposed, onto Pyrenean left-lateral ramps, resulting in a combined finite strain, which hampers the distinction of the two contributions to the strike-slip offset. In addition, our study shows that i) the strain was distributed across the whole area between the Cévennes and Nîmes faults and ii) possible discrepancies between the amount of deformation recorded in the Mesozoic cover and the basement may occur. However, the Priabonian tectonic-sedimentation relationships recorded in some of the studied basins provide minimum values for the offset. The extensional growth structure in the Hérault basin (Fig. 16 c) suggests a minimum extension of some 3 km, corresponding to the minimum value of Priabonian left-lateral offset across this segment of the Cévennes fault. The migration of the Priabonian hanging-wall syncline in the GSCB (Fig. 19) indicates NE-ward displacement of 4 km. Restoration of the LMB points to 375m Priabonian extension (Benedicto et al., 1999) across a releasing bend of the Matelles-Corconne Fault. This could represent the order of magnitude of displacement across the other identified NE-trending faults that bound the Guzargues and Sommières basins (Fig. 2). We thus suggest that the Priabonian cumulated left-lateral motion distributed across the 40 km wide NE-trending wrenching zone remained moderate, with a minimal value of 4 to 5 kilometres. Even though some wrenching may have occurred without

any syn-tectonic deposits recorded, the cumulated Priabonian movement unlikely exceeded 10km. There is no data to document the offset across the major Nîmes fault.

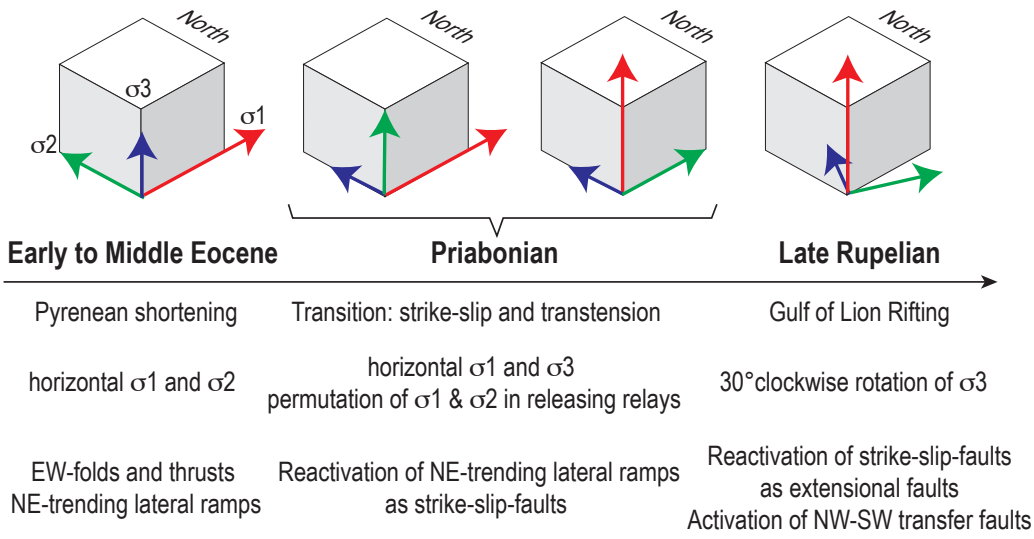


Fig. 23: *Evolution of stress regime from mid-Eocene Pyrenean shortening, through Priabonian strike-slip regime, to Rupelian rifting.*

The transition from Pyrenean compressional mountain building to rifting of the Gulf of Lion in Languedoc was achieved by a specific evolution of the stress regime (Fig. 23). From latest Cretaceous to Bartonian the N-S convergence of Sardinia and associated continental block with southern Europe was accommodated by E-W folds and thrusts, while the inherited regional NE-trending faults were activated as lateral-ramps. Such deformation results from a horizontal σ1 oriented N-S and a vertical σ3 (Arthaud and Laurent, 1995). During Priabonian, the stress regime evolved by permutation of σ2 and σ3 axes, σ1 remaining horizontal and oriented N-S. Such stress regime activated the inherited NE-trending faults as left-lateral strike-slip faults. This fault network inherited from the Late Cretaceous-Bartonian Pyrenean orogeny included restraining and releasing relays, the later correspond to Priabonian depocentres, where σ1 and σ2 permuted. The restraining bend, where compression occurred (σ1 N-S horizontal and σ3 vertical) is illustrated by the Saint-Martin de Londres basins. Contemporaneous releasing bend and relay, where local extension occurred (σ1 vertical) is examplified by the northern part of the Hérault basin, Les-Matelles basin, or Gardon Saint-Chaptes basin. (Fig. 22). Permanence of N-S horizontal σ1 from Late Cretaceous through to Late Eocene was also documented in southern Provence, defined by both strike-slip or reverse faults (Lacombe et al., 1992). N-S shortening along the regional NE-trending faults accommodated transpression of the oblique Africa-Eurasia convergence in the Languedoc-Provence-Corsica domain (Lacombe and Jolivet, 2005) and lasted until Priabonian. During Late Rupelian, the stress regime rotated clockwise around the vertical σ1, by 30°(Séranne,

1999), which activated the inherited NE-trending strike-slip faults as extensional faults. This Rupelian change in stress orientation reflects a change at plate boundaries, which we relate to the onset of subduction roll-back and back-arc rifting above the Apennine subduction (Jolivet and Faccenna, 2000; Jolivet et al., 2015; Réhault et al., 1984; Romagny et al., 2020; Séranne, 1999). Southeast-trending transfer faults zones are a distinctive feature of the rifting of the Gulf of Lion (Canva et al., 2020; Gorini et al., 1993; Guennoc et al., 2000; Maillard et al., 2020; Mauffret et al., 2001). However, there is no evidence of active SE-oriented structure in the onshore study area, until the Oligocene rifting episode (Maerten and Séranne, 1995; Sanchis and Séranne, 2000). The Priabonian strike-slip deformation documented in the present study therefore constitutes a distinct tectonic phase, corresponding to different kinematics and driving forces than the rifting of the Gulf of Lion (Séranne, 1999). We suggest that the development of SE trending transfer zones in the back-arc basin, as from the Rupelian, dates the initiation of slab tearing in the underlying subduction zone (Romagny et al., 2020).

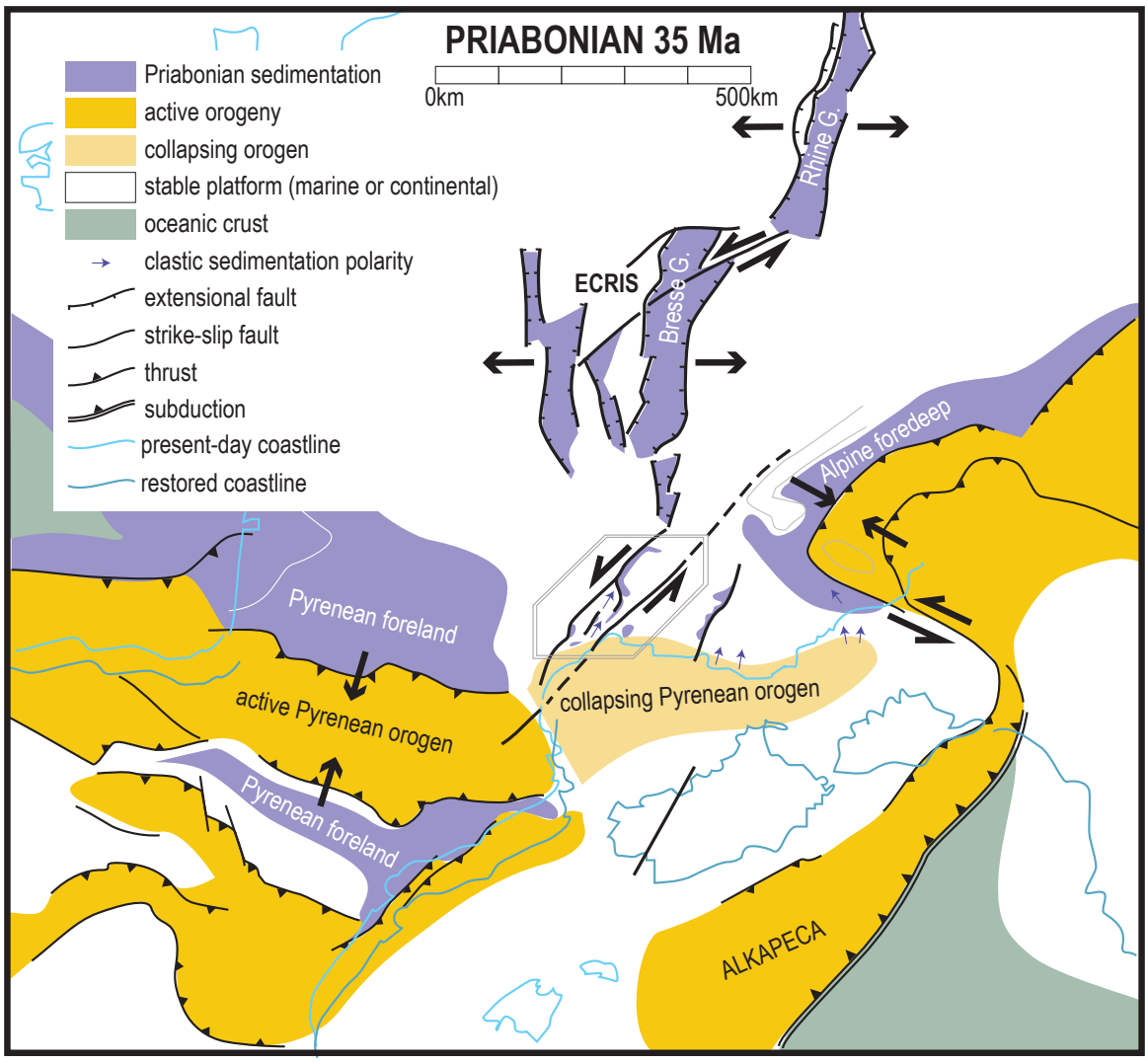


Fig. 24: *Simplified map Pyrenees / European Cenozoic Rift System /Alps during Priabonian.*

Paleogeographic reconstruction at 35 Ma from (Romagny et al., 2020); Position of the North Alpine foreland from (Dumont et al., 2012; Ford and Lickorish, 2004) and of the ECRIS from (Dèzes et al., 2004).

Priabonian left-lateral wrenching in Languedoc links the active Pyrenean orogen to the E-W rifting of the European Cenozoic Rift System (ECRIS) (Dèzes et al., 2004) and the North Alpine Foreland Basin (Ford and Lickorish, 2004). Such left-lateral wrenching during Priabonian most probably extended southward in the Corbières (Fig. 1); however, lack of Priabonian sedimentary record in this area does not allow to confirm this. Eocene northward motion of the Adriatic plate was accommodated by left-lateral oblique convergence along the SW Alps until continental collision in Oligocene (Dumont et al., 2012; Ford et al., 2006) (Fig. 24). Convergence between the Iberia and Europe respectively lasted until early Miocene in the Pyrenees e.g. (Mouthereau et al., 2014; Teixell et al., 2018) and *circa* 85km shortening is computed for the Late Eocene to Early Oligocene interval (Teixell et al., 2016). Pyrenean shortening in Corbières, Languedoc and Provence terminated during the Bartonian (Alabouvette et al., 2001; Bestani et al., 2015; Séguret and Benedicto, 1999). Northward flowing clastic sedimentation pattern in Languedoc, Provence and the SE Alps shows that this segment of the orogen, located in the present-day Gulf of Lion, was being dismantled at that time (Egerton, 1996; Guieu and Roussel, 1990; Joseph and Lomas, 2004).

The left-lateral movement along the Nîmes and Cévennes fault system extended northward to the Jura region where sinistral strike-slip is well documented during Eocene (Homberg et al., 2002). It is also consistent with the transfer zone between Rhine and Bresse grabens (Bergerat et al., 1990; Lacombe et al., 1993; Madritsch et al., 2009), which both record a distinctive Priabonian pulse of sedimentation predating a mid-Oligocene extensional tectonics (Sissingh, 1998). Crustal extension across these rift basin amounts to 7 km (Ziegler and Dèzes, 2005), a value in agreement with the offset measured across the strike-slip faults.

During the Priabonian, the Languedoc-Provence segment of the Pyrenean orogen was thus surrounded by two active convergence zones, while it was undergoing collapse. This different behaviour may be due to the interaction with the west-European rifting (Dèzes et al., 2004), which propagated southward beneath the Languedoc-Provence-Pyrenees.

9. Conclusion

The transition from the Pyrenean orogeny to the Gulf of Lion rifting in onshore Languedoc occurred during Priabonian. Prior to Priabonian time, every structural and tectonic-sedimentation evidence argue for pyrenean compressional deformation. In contrast, after

Priabonian, all structures indicate NW-SE extension related to the Gulf of Lion rifting. This makes the Priabonian interval a key intermediate period of kinematic change. Reassessment of the chronostratigraphic evolution of the area allowed us to constrain the calendar of events. This period is recorded by deposition of a distinctive sedimentary sequence, which is documented both by surface analyses and seismic profiling of Priabonian basins. Such intermediate sequence is bounded at the base by an angular unconformity over the Pyrenean structures, and at the top by the Oligocene syn-rift unconformity, respectively. Continental sedimentation ranges from lacustrine/ palustrine limestone and evaporitic marls at the base, to marls and silts of alluvial plain, including variable proportion of sandstone sheets and channelised fluvial conglomerates, toward the top. The Priabonian series progrades north- or northeastward and is derived from a southern source, presently located offshore, which indicates the existence of a remaining pyrenean topography in the present-day Gulf of Lion.

Mapping and seismic analyses of the tectonic-sedimentation relationships in the Priabonian basins give evidence for syn-depositional deformation against bounding faults and in growth synclines. Contrasted modes of deformation are observed, within short distance, in the Priabonian basins: i) controlled by extensional movement of NNE to NNW-trending faults or segments of faults or ii) controlled by E-W-oriented thrusts and folds. Regional mapping of these structures shows a distribution of the compressional and extensional structures which are consistent with a left-lateral shearing of the area comprised between the major, NE-trending, Cévennes and Nîmes faults. Such sinistral strike-slip of moderate offset (± 5 km) accommodates the E-W extension of the West European rifts basins that are located to the north of the study area. To the south, the strike-slip faults are connected to the ongoing shortening in the Spanish Pyrenees. Priabonian deformation and sedimentation records the initial stages of the dismantling of the Languedoc-Provence segment of Pyrenees, located between two active orogens: the Spanish Pyrenees and the Western Alps. The late orogenic dismantling of the eastern Pyrenees (Jolivet et al., 2020) therefore started up to 5 My earlier than the Rupelian onset of Apennine subduction retreat and associated back-arc NW-SE extension.

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