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## Review Article

# Sediment routing and palaeogeographic evolution of the Western Alpine Foreland Basin during the early collisional stage

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## ABSTRACT

In the Western Alpine Foreland Basin (WAFB), Late Eocene and Miocene periods were characterized by longitudinal sediment routing systems: The first one was situated within the turbidite basin during the underfilled phase and exhibited a northward orientation toward the Swiss Basin, whereas the second was located in the Rhône Valley during the overfilled phase and was directed southward toward the Mediterranean Sea. The transition between these two periods occurred during the Oligocene, which corresponds to both the underfilled/overfilled transition and the early overfilled period. In this study, we provide new fieldwork observations, seismic and well data interpretations, biostratigraphic analyses and a literature synthesis to reconstruct the palaeogeographic and source-to-sink evolution of the WAFB from Priabonian to Aquitanian. The aim is to discuss this reorganisation of sediment routing in relation to the evolution of the Alpine orogenic wedge, as well as the structural inheritance and the suite of geodynamic events that affected southeastern France during the mid-Cenozoic. We divided the WAFB sedimentary formations into four depositional sequences (S1 to S4). During the deposition of the first two sequences (Priabonian to early late Rupelian; ~37.4–28.8 Ma), the WAFB routing system was influenced by the end of the Pyrenean-Provençal orogeny, the European Cenozoic Rifting System (controlling the Rhône Valley s.l.) and the Alpine orogenic wedge (controlling the Alpine foredeep). The very first connection between the Alpine domain and the Rhône Valley is established at ~30 Ma, during the late Rupelian (S2 highstand), controlled by E-W inherited Pyrenean-Provençal structures implying a ‘broken foreland’. In the meanwhile, from the Dévoluy Basin and northward, the orogenic wedge controlled a classical, although thin, foreland basin characterized by a northward sediment routing connected to the Northern Alpine Foreland Basin. Most of the S3 sequence (Latest Rupelian to middle Chattian; ~28.8–23.25 Ma) corresponds to a decrease of clastic Alpine inputs throughout SE France caused by a reorganisation of the drainage network related with the exhumation of the southern External Crystalline Massifs. S3 highstand and S4 sequence (late Chattian to Aquitanian; from ~23.25 Ma) correspond to the establishment of a longitudinal sediment routing system in the Rhône Valley, with material flowing southwards toward the Gulf of Lion, and supplied by the Palaeo-Isère to the north and potentially by the Palaeo-Durance to the south. This final stage in the reorganisation of the drainage network is clearly associated with the post-rift phase of the Gulf of Lion, which facilitated the opening of a new sink and the ultimate southward migration of the sedimentary area.

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## 1. Introduction

Peripheral foreland basins develop on both sides of orogens during their formation (Dickinson, 1974; Beaumont, 1981). The foreland sedimentary record results from the interplay between surface processes (deposition, erosion and sediment transfer) and tectonic-subsidence processes (flexural loading, sediment accumulation, foreland deformation) (e.g., Dickinson and Suczek, 1979). The erosional engines deliver sediments that are evacuated from the orogen and transported to the foreland basin by the sediment routing system mainly composed of the river drainage network (Allen, 2008). The routing system and the associated sediment production are influenced by multiple processes, including tectonics (rock exhumation especially at source, competition between drainage mobility and rock-uplift rates, tectonic subsidence) (e.g., Densmore et al., 2007; DeCelles, 2012; Armitage et al., 2015) and climato-eustasy (sea-level variations, precipitation rates acting on erosion and transport capacity), both can induce base level changes (e.g., Colomera et al., 2017; Willett et al., 2018; Peris Cabré et al., 2023;

Vaucher et al., 2023). The duration of variations in these mechanisms associated with the response times of routing systems induce a succession of transient and steady-state stages. Two end-members of routing systems characteristics can be defined during each stage of the usual classification of foreland basins (e.g., Covey, 1986; DeCelles and Giles, 1996; Sinclair, 1997; Ford et al., 2006; Babault et al., 2012; Allen, 2017; Catuneanu, 2019): (1) During the earliest phase of orogenic growth, loading of the distal part of a (frequently thinned) margin results in highly subsiding foredeep, filled by low volumes of sediments issued from the first emerging relief. This results in underfilled basins with a longitudinal routing system established along the flexural basin, parallel to the orogenic front. (2) During the collision, erosional products from an emerged orogen are filling up the foredeep, most of the material is trapped in the foredeep which becomes an overfilled basin. The routing system during the late stages of orogenic development (and moreover during post orogenic times) is formed of large transverse rivers draining the entire orogen and merging into a main, large longitudinal system still flowing in the foredeep. The transition between these two stages



Fig. 1. Geological map with superimposed shaded relief of the western Alps and the WAFB adapted from the geological map of Bureau de recherches géologiques et minières (1/1000000; <https://infoterre.brgm.fr/>).

implies a considerable reorganisation of the routing system, which timing and forcing are usually not very well constrained.

The Northern Alpine Foreland Basin (NAFB) (Fig. 1) is a well-documented example of the evolution of a foreland basin: it is a well-defined and very large (700 km × 150 km) basin along the pro-side of the Alpine orogen. In response to active and continuous lithospheric flexure, up to 5 km thickness of sediment accumulated between the Eocene and Late Miocene (Matter et al., 1980; Kuhlemann and Kempf,

2002; Ford and Lickorish, 2004; Berger et al., 2005). During both the underfilled and overfilled phases, most of the Alpine sedimentary material was transported in a single longitudinal drainage system located in the foredeep and running northeastward, parallel to the orogen front. During the overfilled phase, large transverse drainage systems develop and carried sediments from the Alpine orogenic wedge to the foredeep (Büchi and Schlanke, 1977; Platt and Keller, 1992; Sinclair, 1997; Jin et al., 1995; Kuhlemann and Kempf, 2002; Schlunegger et al., 1997a,

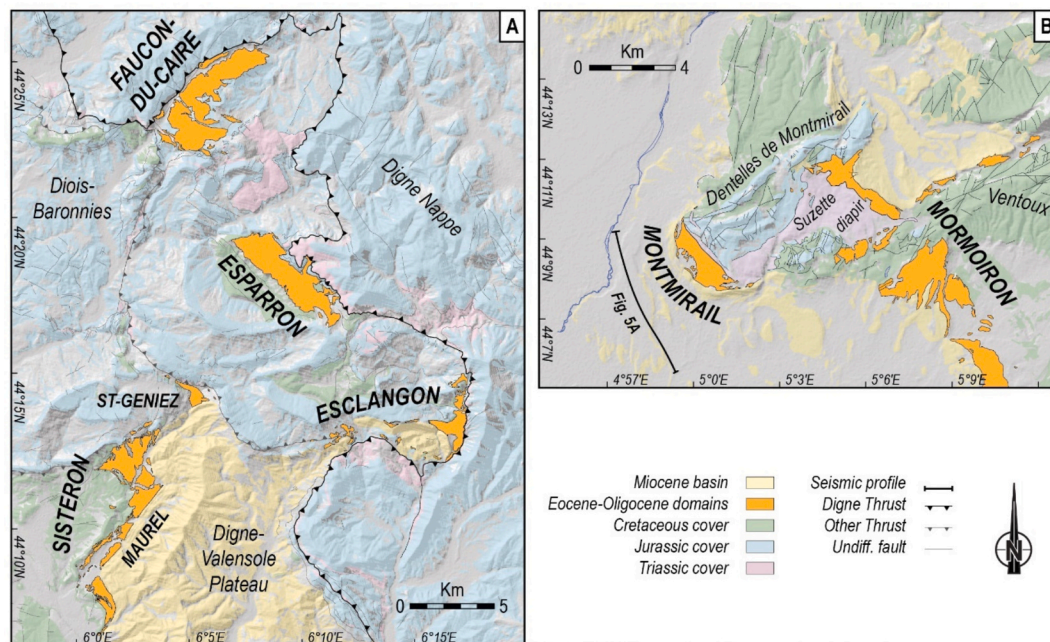


Fig. 2. Zoomed-in geological maps of (A) the Faucon-du-Caire, Esclangon-Esparron and Sisteron domains, and (B) the Montmirail Domain and the northern part of the Mormoiron Domain. Modified from Rousset et al. (1983), Haccard et al. (1989), Gidon et al. (1991), Gigot et al. (in prep), Monier et al. (1991) and Monjuvent et al. (1991).

1997b; Berger et al., 2005; Kuhlemann, 2007; Sharman et al., 2018).

The southwestern equivalent of the NAFB, the Western Alpine Foreland Basin (WAFB) (Figs. 1 and 2) is displaying a congruent although slightly different evolution. During the underfilled stage (Eocene, i.e. Flysch Basin), the configuration of the basin and routing system are the same as for the NAFB. A dominant sediment routing system with northward palaeoflow is located in the foredeep at the front of the proto-Alpine wedge. Sediments were derived from short transverse drainage systems issued from peripheral relief of the upper plate and Provençal reliefs (Ravenne et al., 1987; Sinclair, 1997; Joseph et al., 2005). During the Oligocene, the initiation of the collision (around 32 Ma; e.g., Niggli, 1986; Handy et al., 2010; Bellahsen et al., 2014; Rosenberg et al., 2021) and the creation of the first major Alpine relief through the exhumation of the Internal Alpine nappes (e.g., Jourdan et al., 2012; Fauquette et al., 2015) caused the export of massive sedimentary material to the WAFB. The basin evolved from an underfilled to an overfilled basin (Ford et al., 1999; Ford and Lickorish, 2004; Ford et al., 2006). During the late overfilled stage (Miocene, i.e. Molasse Basin), the main depocenter migrated into the Rhône Valley, shifting the sedimentary depocenter several tens of kilometres to the west of the Alpine front. Palaeogeographic maps (Sissingh, 2001; Kalifi et al., 2022), reconstruction of incised valley geometries (Besson et al., 2005), position of lowstands in the Gulf of Lion (Gorini et al., 1993; Jolivet et al., 2015; Rubino et al., 2015) and volume budgets (Kuhlemann, 2000; Leroux, 2012; Leroux et al., 2017) indicate that the Rhône Valley formed a major longitudinal routing system with a southward transfer of sediments into the Gulf of Lion. The connection with the orogen was ensured through two main transverse drainages that have been active at least since Burdigalian-Langhian times (Perriaux et al., 1984; Montenat et al., 2000; Besson et al., 2005; Kalifi et al., 2022), pre-forming the current routing system of the Western Alps made up of the Isère and Durance rivers (Fig. 1).

The timing and the cause of the routing system migration and the settlement of this particular drainage network are poorly constrained. According to Kalifi et al. (2022), the displacement of the routing system toward the Rhône Valley occurred brutally during the late Burdigalian

(17.8–16.2 Ma) and can be related to the exhumation of the External Crystalline Massifs. This interpretation seems hardly consistent with the occurrence of Alpine material in the middle Rhône Valley Oligocene deposits (Casagrande et al., 1989), which suggests that the routing system was already active at that time. However, this early orogenic evolution of the Alpine routing system during Oligocene times appears unconstrained and poorly known in terms of timing and forcing (location of subsidence, orogenic process, climate). Unlike Eocene and Miocene systems, the Oligocene routing system shows a heterogeneous sediment distribution. They form a mosaic of small domains preserved today between the current wedge-top of the fold and thrust belt named here the Frontal Basin, and the Rhône Valley named here the Rhône Valley Basin. This suggests either poor preservation, a highly fragmented foreland, and/or complex repartitioning of subsidence. Until now very few studies have attempted documenting the palaeogeographic evolution of these domains or their possible connections in terms of sediment routing, whether between the Frontal Basin (i.e. Held, 2011; Grosjean et al., 2017), between the Rhône Valley Basin (i.e. Semmani et al., 2023), or at the scale of the entire WAFB (e.g., Cavelier et al., 1984; Sissingh, 2001). The infilling of these two groups of domains and their associated routing system are to be characterized in order to recombine the evolution of the entire WAFB routing system during early orogeny as well as the dominant processes controlling the routing system evolution. In this study, we investigated the reorganisation of the palaeogeography and routing system of the WAFB during its underfilled/overfilled transition that we are targeting between the Priabonian and the Aquitanian. Our approach consists of an exhaustive tectono-sedimentary synthesis of the WAFB domains during this period starting from the existing data published in the literature completed with new data: fieldwork, biostratigraphic analyses and reinterpretation of subsurface data. This synthesis allowed us to propose new interpretations for each WAFB domain with a synthetic sedimentary log including source and drainage data. Based on the correlation of all the domains, we produced detailed restored source-to-sink and palaeogeographic/palinspastic maps compiling palaeocurrent data, restoration of the orogenic wedge, as well as source and drainage divide positions.

## 2. Geological setting

### 2.1. General geodynamics and structure of the Alpine orogen

The European Alps are part of a broad peri-Mediterranean geodynamic framework, reflecting the complex intricate interplay between the African and Eurasian plates that generated a mosaic of orogenic belts, including the Alps, Pyrenees, Betic–Rif–Maghrebides, Apennines, Carpathians, and Dinarides (e.g., [Guerrera et al., 2019](#); [Martín-Martín et al., 2020](#)). These chains display a common architecture with an internal domain and rather early flysch units and an external domain formed by a folded Mesozoic–Cenozoic sedimentary cover associated to foreland basins, recording substantial sedimentary infill from the Paleogene to the Pliocene.

The Alps result from the subduction of the Valais domain and the Ligurian Ocean between Cretaceous and Eocene times, and from the following collision of the European and Adriatic plates starting during Early Oligocene time (e.g., [Froitzheim et al., 1996](#); [Stampfli et al., 2002](#)). The orogenic wedge built on two main exhumation phases corresponding to the rise of the internal (~Oligocene) and external (Miocene–Pliocene) Alps, domains separated by a major fault zone, the Penninic Thrust ([Schmid et al., 2004](#)) ([Fig. 1](#)). The External Alps, west of the Penninic Thrust, correspond to the proximal margin of the European plate shortened during the collision, i.e., during Oligocene–Miocene times. They include the External Crystalline Massifs (from north to south: Aar, Aiguilles Rouges, Mont Blanc, Belledonne, Grandes Rousses, Pelvoux and Argentera massifs; [Fig. 1](#)) mainly made of Hercynian crystalline basement, Palaeozoic sedimentary rocks and a thick Mesozoic sedimentary cover of marine origin, both recording the rifting phase and the infilling of the passive margin during Jurassic and Cretaceous times ([Baudrimont, 1977](#); [Lemoine et al., 1986](#)). The External Alps were imprinted by collision processes; the External Crystalline Massifs and their cover recorded Alpine greenschist metamorphism due to their burial below the Internal Alpine nappes during the Oligocene (e.g., [Rolland et al., 2003](#); [Challandes et al., 2008](#); [Sanchez et al., 2011](#); [Bellahsen et al., 2012, 2014](#); [Cenki-Tok et al., 2014](#); [Bellanger et al., 2015](#)). The Internal Alps, east of the Penninic Thrust, form a stack of highly deformed units of different origins. The Briançonnais and Internal Crystalline Massifs (from north to south: Monte-Rosa, Grand Paradiso and Dora-Maira massifs; [Fig. 1](#)) consist of pre-Alpine crystalline basement and Upper Palaeozoic to Mesozoic sedimentary cover that originate from the Briançonnais continental block and its distal passive margins ([Schmid et al., 2004](#)). On the other hand, the Valais and Piemont-Liguria units are the remnants of the former oceanic or hyper-extended domains and their deep marine Mesozoic sediments ([Schmid et al., 2004](#); [Agard, 2021](#)). These four units were subducted below the Adriatic Plate, along with its distal margin, i.e., the Austro-Alpine units ([Compagnoni, 1977](#); [Ballèvre et al., 1986](#); [Malusà et al., 2005](#); [Manzotti et al., 2014](#)). They experienced various grades of metamorphism, ranging from greenschist facies in the external Briançonnais rocks, to blueschist facies in the Briançonnais and Schistes Lustrés, to eclogite facies in most of the Internal Crystalline Massifs, and ultra-high pressure facies in the Dora Maira massif (e.g., [Chopin, 1984](#); [Agard et al., 2001](#); [Bousquet et al., 2002](#); [Ganne et al., 2007](#); [Lanari et al., 2012](#); [Groppo et al., 2019](#); [Agard, 2021](#); [Bonnet et al., 2022](#)).

### 2.2. Development of the Alpine Foreland Basin

In the Dauphinois domain (i.e., proximal Europe), the flexural basin started to develop at around 49 Ma ([Ford and Lickorish, 2004](#)) and formed a marine subsiding trench 100–120 km wide ([Ford and Lickorish, 2004](#)), running parallel to the Alpine proto-relief and gradually migrating westwards and northwards (15 mm/year; [Ford et al., 1999](#)). As the orogen did not yet form high topographic relief, sediment input was low and typical of an underfilled foreland basin ([Covey, 1986](#); [Allen et al., 1991](#); [Sinclair, 1997](#)). Eocene sedimentation corresponded to the

classic “Nummulitic trilogy” ([Boussac, 1912](#); [Bodelle, 1971](#); [Campredon, 1972](#)) composed of a carbonate platform (Calcaires nummulitiques) on the western margin and passing basinward to deep marine marls (Marnes bleues) or turbidite deposits (Grès d’Annot, du Champsaur, des Aiguilles d’Arves, de Taveyenne, du Val d’Illiez, Morcles; [Fig. 1](#)). Clastic material originated mainly from the southerly-located Corso-Sardinian and Maures-Esterel massifs and from the upper plate of the developing Alpine orogenic wedge ([Jean et al., 1985](#); [Joseph and Lomas, 2004](#)).

At the same time, the Provence and southern Rhône Valley endured the last phase of N-S compression related to the Pyrenean-Provençal orogeny, which lasted until the Priabonian ([Séranne et al., 2021](#)). This geodynamic event originated from the collision between the Iberian and European plates ([Arthaud and Séguret, 1981](#); [Tempier, 1987](#)) and formed mainly E-W compressive structures (Alpilles, Luberon, Ventoux-Lure massifs; [Fig. 1](#)) throughout southeastern France, from the Provence to the Dévoluy massif ([Balansa et al., 2022](#) and references therein).

From the Late Eocene to Oligocene, the Rhône Valley experienced extension as part of the European Cenozoic Rift System ([Bergerat, 1987](#); [Bergerat et al., 1990](#); [Ziegler, 1992](#); [Dèzes et al., 2004](#)) causing the opening of subsiding trenches controlled by N-S oriented normal faults or NE-SW strike-slip faults. These depocenters form the Rhône Valley Basin that group from north to south: the Bas-Dauphiné, Montmirail, Mormoiron, Durance and Provence domains ([Figs. 1 and 2b](#)). Sedimentation is supposed to be exclusively continental, with evaporitic lake deposits predominating (e.g., [Triat and Truc, 1972](#); [Cavelier et al., 1984](#); [Dumas, 1986](#); [Lesueur, 1991](#); [Semmani, 2022](#)). However, marine incursions have often been suggested but were never documented in detail ([Touraine, 1976](#); [Nury, 1987](#); [Dumas, 1988](#); [Lesueur, 1991](#); [Châteauneuf and Nury, 1995](#); [Châteauneuf et al., 2019](#); [Semmani et al., 2023](#)).

At the Eocene/Oligocene transition, the onset of collision and the activity of the Penninic thrust led to the gradual exhumation of the Internal Alpine units (e.g., [Hurford et al., 1991](#); [Gebauer et al., 1997](#); [Fügenschuh and Schmid, 2003](#); [Malusà et al., 2005](#)) and the formation of high topography (~1900 m of mean elevation) of the orogenic wedge ([Fauquette et al., 2015](#); [Jourdan et al., 2013](#)). The Alpine foreland basin, still migrating north and west, became overfilled because of the increase in sedimentary input ([Ford et al., 1999](#); [Ford and Lickorish, 2004](#); [Ford et al., 2006](#)). This Molasse Basin is well documented in the NAFB where it is summarised in two large transgression/regression cycles between the Oligocene and the Late Miocene (Cycle I: Lower Marine Molasse / Lower Freshwater Molasse; Cycle II: Upper Marine Molasse / Upper Freshwater Molasse; [Matter et al., 1980](#)). However, as mentioned above, it is much less constrained in the WAFB and has never been discussed on a large scale. During the Oligocene, the foredeep was seemingly fragmented in several depocenters that form the Frontal Basin, which group from north to south: the Savoie, Dévoluy, Faucon-du-Caire, Esclançon-Esparron, Sisteron and Barrême domains ([Figs. 1 and 2a](#)). Sedimentation consisted mainly of fluvio-lacustrine deposits (e.g., [de Graciansky et al., 1982](#); [Haccard et al., 1989](#); [Grosjean et al., 2017](#)). The propagation of the first Alpine nappes and thrusts affected the development of the domains, leading to *syn*-tectonic sedimentation, as in the Savoie Domain under the Salève thrust rise ([Allen and Bass, 1993](#); [Girault et al., 2022](#)). The best-known example of an Oligocene thrust is probably the Digne Nappe, considered to be the front of the Alpine deformation between the Dévoluy and the Castellane Arc ([Célini et al., 2022](#)), and whose timing of activation is still debated today, between Early ([Lickorish and Ford, 1998](#); [Ford et al., 1999](#)) to Late ([Gidon, 1997](#); [Célini et al., 2022](#)) Oligocene. All Frontal Basin domains are preserved in the footwall of the Digne Thrust except for the Barrême Domain.

Between the Early Oligocene and the Middle Miocene, southern France experienced the opening of the Gulf of Lion, which represents the northern part of the Liguro-Provençal Basin and developed as a back-arc basin related to the counterclockwise rotation of the Corso-Sardinian block (e.g., [Bois, 1993](#); [Séranne, 1999](#)). This opening occurred in two major stages ([Bache et al., 2010](#); [Jolivet et al., 2015](#)): (1) an Oligocene-Aquitainian rifting phase (~32–24 Ma) with a first rifting stage

(Oligocene) characterized by the development of small, thin, and discontinuous basins along the margin and by a poorly supplied central rift, and a transitional period (Aquitainian) related to a the continental break up and a more continuous deposition; (2) a post-rift Burdigalian–Langhian phase (~24–15 Ma), corresponding to the rapid rotation of the Corso-Sardinian block and the emplacement of oceanic crust (e.g., Gattaceca et al., 2007; Jolivet et al., 2015). The latter phase is marked by high sedimentation rates (up to 8 km accumulated between the Burdigalian and the present; Gorini, 1993; Guennoc et al., 2000; Jolivet et al., 2015), reflecting deepening, significant post-rift subsidence and substantial sediment input derived from the surrounding distributaries among all the former Rhône fluvial system.

Following Oligocene continental sedimentation, the Miocene Marine Molasse Cycle is preserved in the Rhône Valley and extends to the southeast as far as the Digne-Valensole Basin. Its well-documented palaeogeography consists in a narrow seaway that connected the Gulf of Lion and the NAFB (Rubino et al., 1990; Kalifi et al., 2022). It recorded a mixed carbonate-clastic marine sedimentation alternating with fluvial deposits until its complete shift to continental conditions at the end of the Tortonian (Lesueur et al., 1990; Rubino et al., 1990; Crumeyrolle et al., 1991; Besson et al., 2005; Reynaud and James, 2012; James et al., 2014; Kalifi et al., 2020). This Miocene Basin developed during the exhumation of the External Alps and has been affected by the spread of fold and thrust belts related to crustal shortening (External Crystalline Massifs exhumation), well expressed in the Bas-Dauphiné and Savoie domains to the north, which have been subject to continuous *syn*-tectonic sedimentation throughout the Miocene (e.g., Beck et al., 1998; Kalifi et al., 2021; Bilau et al., 2023).

### 3. Material and methods

#### 3.1. Fieldwork sedimentology

Most of the WAFB domains are well outcropped. For those with limited descriptions in existing literature, we carried out sedimentological and stratigraphic logging in the field. Sedimentary facies are identified based on the sedimentary features, geometries and fossil content, and are grouped in sedimentary facies associations indicative of depositional environments. From the spatial and vertical evolution of these sedimentary facies, we identified several sedimentary sequences. We used the Embry (1993, 1995) model that defines depositional sequences as one Transgressive System Tract (TST) topped by a Maximum Flooding Surface (MFS) and one regressive system tract topped by a sequence boundary (SB) which registers the most proximal deposits or exposure. Forced regression and lowstand deposits are rarely preserved or difficult to identify in the Eocene-Oligocene sedimentation of the WAFB. Thus, regressive system tracts mainly correspond to Highstand System Tracts (HST), and we will use this term hereafter. Palaeocurrents measurements (e.g., channel base erosion, dune direction, flute-casts) were also carried out in order to reconstruct the drainage system. In total, we logged more than 1800 m of sedimentary sections in the Bas-Dauphiné, Dévoluy, Faucon-du-Caire, Montmirail, Mormoiron and Sisteron domains.

#### 3.2. Interpretation of subsurface data

The Cenozoic formations of the Bas-Dauphiné, Camargue and Durance domains can be described in the subsurface based on wells and seismic lines (BRGM Database). From well data, we used survey reports, cores and well logs (Gamma Ray (GR) and Spontaneous Potential (SP)) to identify depositional environments and sequences. We carried out well correlation transects to determine the geometry of sedimentary deposits at the basin scale. In total, we studied 37 wells (see location on Fig. S1). From seismic data, we constrained 1) the architecture of the domains and 2) the key surfaces and major sequences based on seismic stratigraphy (erosional truncature, onlap, downlap and toplap

geometries as defined in Mitchum Jr et al. (1977)).

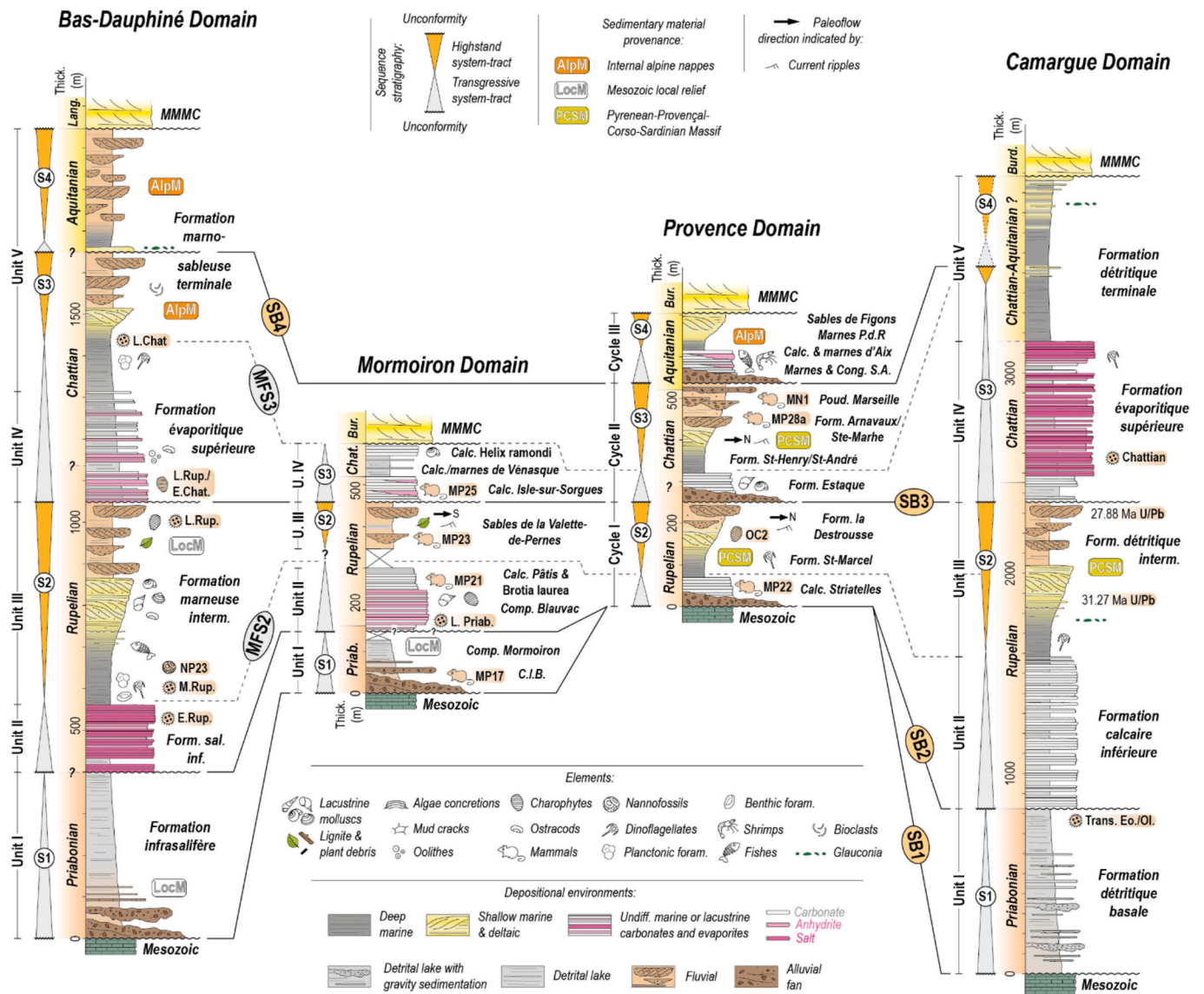
#### 3.3. Biostratigraphy

This study includes a dating section in order to temporally constrain our stratigraphic synthesis of the WAFB. First, we did a review of biostratigraphic assemblages described in the studied domains. There are many biostratigraphic studies on the Priabonian, Oligocene and Aquitanian domains of the WAFB, but most of them pre-date the 2000s. Since then, the biozones and the range of species have evolved and, as a result, the ages of the formations given in the literature need to be updated. We therefore listed all the species (planktonic and benthic foraminifera, calcareous nannofossils, mammals, charophytes) described in the literature and checked their range (first occurrence, last occurrence) in the sedimentary record based on the most recent biostratigraphic studies. The range of more than 160 species described in the WAFB literature was analysed. The results of this review have led us to revise the age of many sedimentary formations in the Barrême, Bas-Dauphiné, Dévoluy, Durance, Provence, Savoie and Sisteron domains. Second, we performed pollen and calcareous nannofossil content analyses to provide new age constraints on undated formations (from Bas-Dauphiné, Dévoluy, Montmirail and Mormoiron domains), either from outcropping terrains or borehole cuttings (GVA1 and SL2 wells; Bas-Dauphiné Domain; see location on Fig. S1). The analysis and determination of pollen was performed using the concept of climatostratigraphy to link flora contents to specific climatic periods within a pre-established chronostratigraphic frame. The analysis and determination of calcareous nannofossil species was performed referring to Martini (1971) and Gradstein et al. (2020). Twenty-four samples were analysed including 14 sterile samples and 10 samples with sufficient material for dating. The location and results of dating, as well as the review of the biostratigraphic data from literature, are available in supplementary data (Figs. S2 and S3; Table S1).

#### 3.4. Palaeogeographic and source to sink restitution

Based on the tectono-sedimentary synthesis from the literature and new data, we produced a series of synthetic logs for each WAFB domain. These logs detail the evolution of the depositional environments, the time framework, the palaeoflow directions of the sedimentary system, the origin of sedimentary inputs and a sequential division. We then produced detailed palaeogeographic and source to sink maps of the entire WAFB for each half-sequence and maximum flooding stages. Our chronostratigraphic framework is based on that of Gradstein et al. (2020). We use the terms early, middle and late to divide the Rupelian (early: 33.9–31.7 Ma; middle: 31.7–29.5 Ma; late: 29.5–27.3 Ma) and Chattian (early: 27.3–25.8 Ma; middle: 25.8–24.5 Ma; late: 24.5–23.03 Ma) periods. We used the works of Bellahsen et al. (2014) and Bellahsen et al. (in prep) to restore the Alpine domain and more specifically the migration of the Penninic Thrust between the Priabonian and the Aquitanian. To reconstruct the source to sink system, we used the palaeoflow directions and the nature of the clastic fraction recorded in the sedimentary formations to highlight the different drainage systems, their origin and outlets, and the location of the drainage divide. Over the studied period, the WAFB received sedimentary input from many source areas. For the sake of simplicity, we regroup them into three main assemblages of distinct origin:

- Internal Alpine Nappes assemblage: From ~32 Ma, the Internal Alps gradually exhumed (e.g., Hurford et al., 1991; Gebauer et al., 1997; Fügenschuh and Schmid, 2003; Malusà et al., 2005). As a result, the subducted and exhumed internal units start to feed the foreland basin with igneous or metamorphic rocks of various grade. In this assemblage, we therefore group all sedimentary input coming from the erosion of the Internal Alpine nappes. According to many authors, in sedimentary deposits, this source can be attested by the presence of



**Fig. 3.** Synthetic logs and stratigraphic correlations between Rhône Valley Basin (Bas-Dauphiné, Mormalion, Provence and Camargue domains). Synthetic logs are adapted from: Bas-Dauphiné: [Dumas \(1986, 1987, 1988\)](#) and this work. Mormalion: [Triat and Truc \(1972, 1974\)](#) and this work. Provence: [Nury \(1987\)](#). Camargue: [Valette \(1991\)](#), [Semmani \(2022\)](#) and this work. Abbreviations: Calc. = Calcaires; C.I.B.=Conglomérats de Crillon-le-Brave; Comp. Blauvac = Complexe argileux blanc de Blauvac; Comp. Mormalion = Complexe argileux vert de Mormalion; Form. = Formation; Form. Sal. Inf. = Formation salifère inférieure; Interm. = intermédiaire; MMMC = Miocene Molasse Marine Cycle; P.d.R. = Puy-du-Roy; Poud. = Poudingues; S.A. = Saint-Anne.

radiolarite, basalt, diabase, gabbro, various greenschists/blueschists and serpentinite pebbles, indicating the recycling of an oceanic crust and its cover, i.e. the Piemont-Liguria ([Fig. 1](#)) ([Chauveau and Lemoine, 1961](#); [Gidon et al., 1971](#); [Ivaldi, 1989](#); [Evans and Mange-Rajetzky, 1991](#); [Meckel et al., 1996](#); [Meckel, 1997](#)). Granite and quartzite are also common and indicate the erosion of the Internal Crystalline Massifs or the Briançonnais units ([Fig. 1](#)) ([Ivaldi, 1989](#); [Meckel, 1997](#)). Concerning the heavy minerals fraction, [Latreille \(1969\)](#) considers the “epidote-garnet-glaucophane” assemblage to be typical of the Internal Alpine nappes, with additional species from metamorphism (lawsonite, pyroxene, jadeite, olivine, etc.).

- Pyrenean-Provençal-Corso-Sardinian massif assemblage: Since its collision with the European block, the eastern part of the Iberian plate constitutes a source for the southern WAFB. In this assemblage, we group sedimentary input coming from the Corso-Sardinian block (at least until its separation from the Provence as a consequence of the Gulf of Lion rifting, ~32–24 Ma; [Séranne, 1999](#); [Jolivet et al., 2015](#)) and from the Maures-Esterel massif ([Fig. 1](#)), which is part of

the same geodynamic unit and mainly consist of acid igneous rocks and volcano-clastic sediments. The petrological signature of this second assemblage in sedimentary deposits corresponds to the presence of quartz/quartzite, quartzophyllade, microgranite and rhyolite pebbles, and a heavy mineral fraction dominated by staurolite, apatite, tourmaline, andalusite/kyanite, zircon and hornblende ([Jean, 1985](#); [Guieu et al., 1987](#); [Evans and Mange-Rajetzky, 1991](#)).

- Local assemblage: The most important sedimentary source in the entire WAFB remains the Mesozoic cover of the External Alps. Indeed, all the studied domains are mainly surrounded by carbonate massifs of Triassic to Cretaceous ages, and their erosion often provides the bulk of sedimentary material through very short and local drainage systems. In this assemblage, we therefore group all External Alps Mesozoic cover inputs.

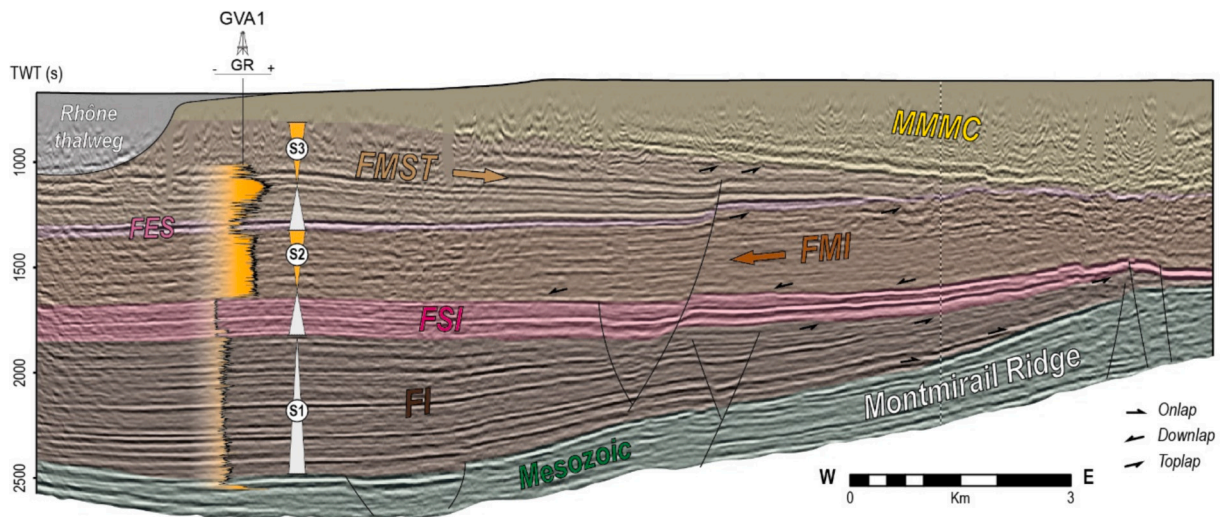


Fig. 4. Seismic interpretation of the Bas-Dauphiné 82SE01 profile. See location on Figs. 1 and S1. Abbreviations: FI=Formation infrasalifère; FSI=Formation salifère inférieure; FMI=Formation marneuse intermédiaire; FES=Formation évaporitique supérieure; FMST = Formation marno-sableuse terminale; MMMC = Miocene Molasse Marine Cycle.

#### 4. Results – sedimentary evolution of the WAFB domains

In this chapter, we describe the chronostratigraphic evolution of each domain within the WAFB, synthesizing evidence from both our own results and previously published studies. This reconstruction is supported by a synthetic sedimentary log and an associated sequence-stratigraphic framework.

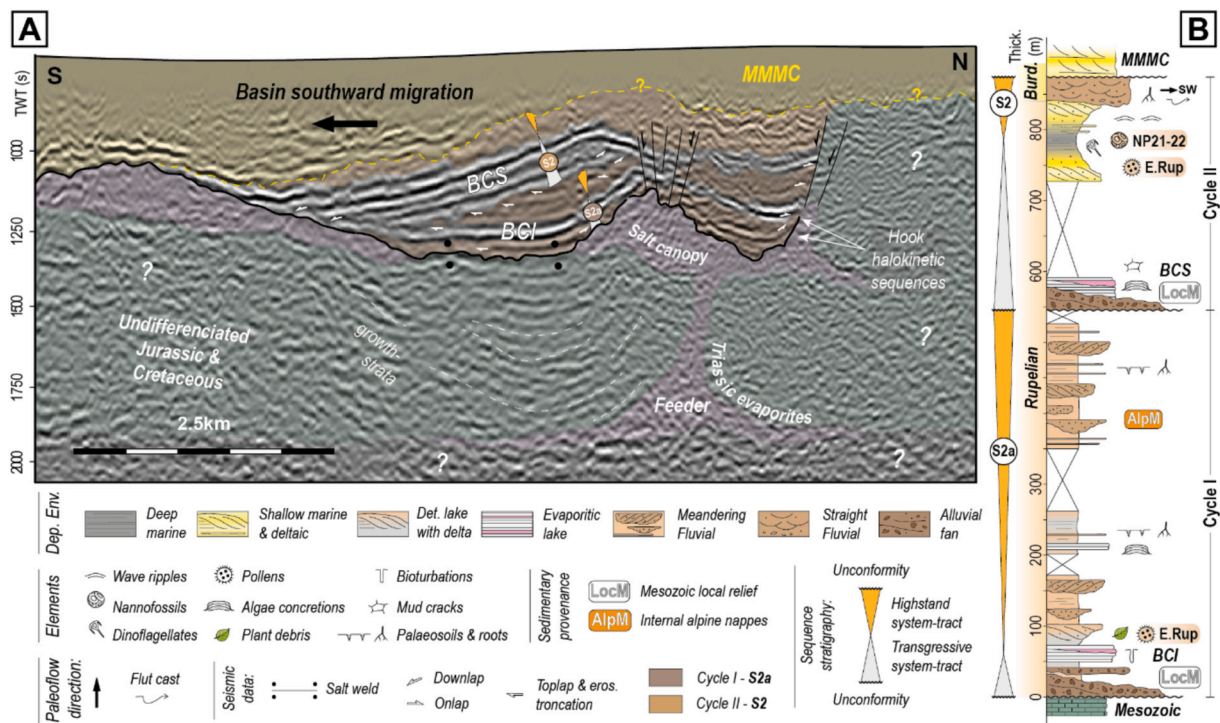
##### 4.1. Rhône Valley Basin

###### 4.1.1. Bas-Dauphiné Domain

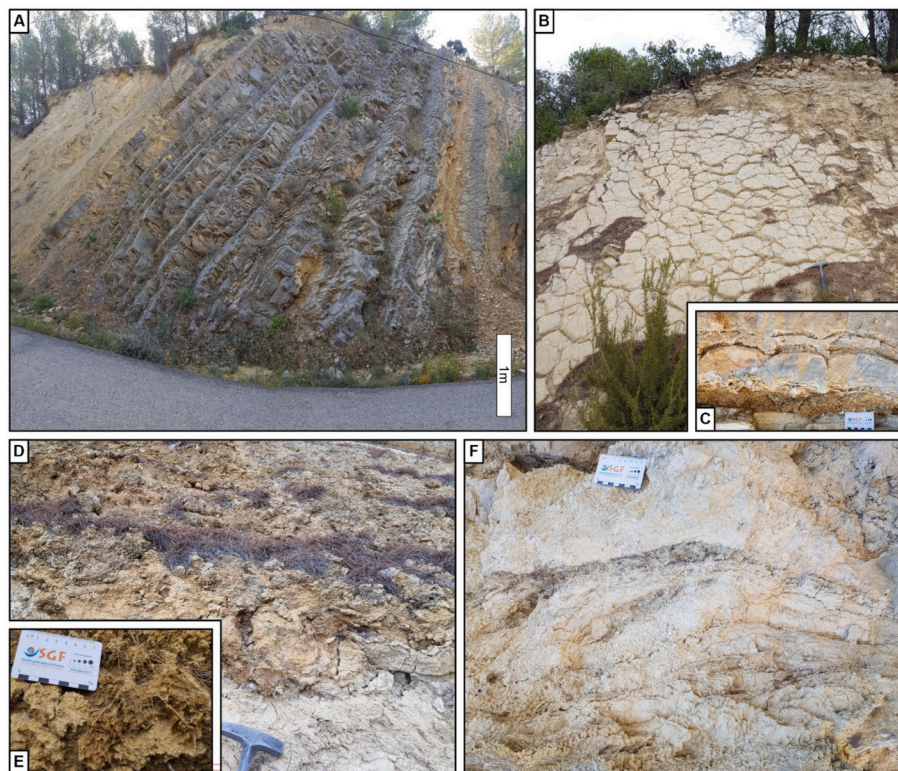
The Bas-Dauphiné Domain is the most northern and largest of all Rhône Valley Basin. It covers a 4000 km<sup>2</sup> area located between the crystalline formations of the Massif Central to the west and the Mesozoic sedimentary rocks of the Sub-Alpine chains (Vercors, Chartreuse) to the east (Fig. 1). It contains several hundred meters of Paleogene deposits (2200 m maximum in the Valence well (GVA1; Figs. S1 and S4) classically related to *syn*-tectonic sedimentation in response to the European Cenozoic Rift System event (Bergerat, 1987; Bergerat et al., 1990). The Miocene Marine Molasse Cycle deposits mainly outcrop and compose most of the landforms of the Bas-Dauphiné Domain (except in the Crest and Monjoyer areas; Figs. 1 and S1), whereas Paleogene deposits are mainly known through subsurface data. In addition to literature data, we use new results from the reinterpretation of seismic and well data as well as fieldwork and biostratigraphic analyses to describe the formations of the Bas-Dauphiné Domain.

According to the nomenclature established by Gudefin (1977), Goy and Busson (1980) and Dumas (1986), these formations are classically divided into five predominantly clastic or carbonate-evaporitic main composition. The first one, named the Formation infrasalifère (up to 1000 m thick; Fig. 3), consists of various clastic lithologies alternating conglomerates, sand and marls. New well log interpretation reveals a fining upward sequence passing from alluvial fan to clastic lake deposits (Figs. S4 and S5). The Formation infrasalifère is therefore regarded as a retrograding sequence. New Seismic interpretation shows growth-strata related to active faults in the western part of the Domain as well as eastward to southward onlap geometries on two Mesozoic topographic highs that we name the Montmirail and Réauville ridges (Figs. 4, S1, S4 and S6). Seismic interpretation shows that the latter forms a 50 km wide E-W anticline that separates the Bas-Dauphiné Domain from the middle Rhône Valley to the south (Fig. S6). The sedimentary material of the Formation infrasalifère is only of local origin, coming either from the

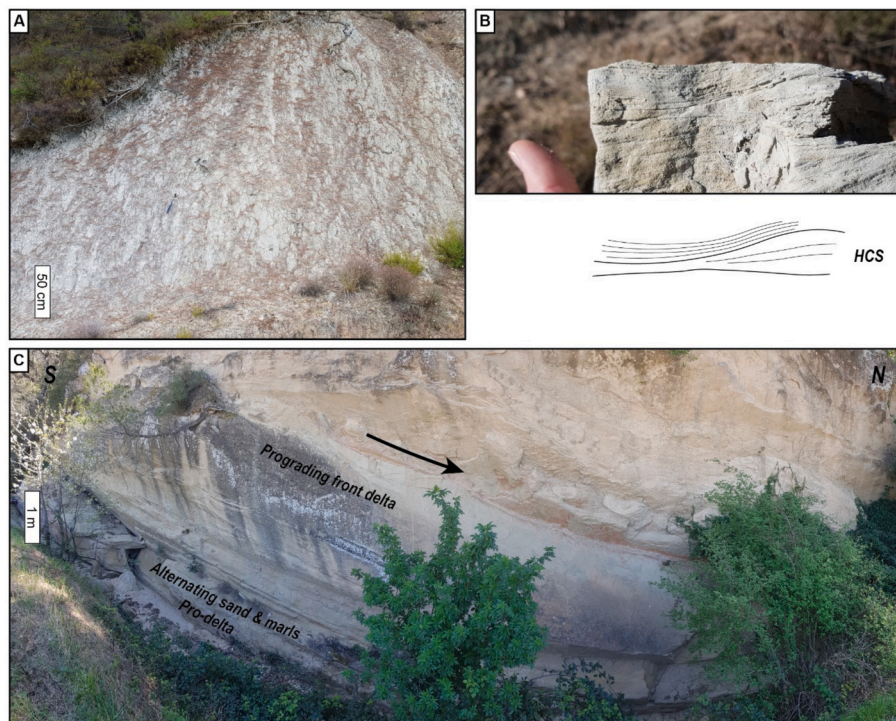
dismantling of the Massif Central in the western part of the Domain, or from the Mesozoic cover in the eastern part (Latreille, 1969; Dumas, 1986, 1987, 1988). The age of the Formation infrasalifère remains unknown due to the lack of biostratigraphic constraints and is classically attributed to the Priabonian (Dumas, 1986, 1987, 1988). The second unit, the Formation salifère inférieure (Fig. 3), corresponds mainly to carbonate-evaporitic deposits and shows strong variations in thickness throughout the Bas-Dauphiné Domain (up to 700 m thick; Fig. S4). It has been widely documented by Dumas (1988), who described a typical playa-lake (according to Eugster and Hardie, 1978) facies distribution: salt and anhydrite deposits in the central part of the Domain, limestones in the outer part, and alluvial fan deposits on the border. The palaeoenvironment is mainly lacustrine but could have been subject to marine incursion (Dumas, 1988; Busson et al., 1992; Fontes et al., 1996). Here we bring new marine evidence in the upper part of the formation based on the occurrence of dinoflagellate cyst and mangrove pollen assemblages (Figs. 3 and S2). Seismic interpretation reveals that the Formation salifère inférieure is strongly unconformable on the underlying Formation infrasalifère and shows growth-strata (Fig. 4). In the Crest and Monjoyer areas, some few outcrops reveal lacustrine limestone beds (Fig. S7). New pollen analyses in the upper part of this second unit reveal typical assemblages of cold and dry climate related to the early Rupelian (Figs. 3 and S2; Table S1). The third unit is named the Formation marneuse intermédiaire (up to 600 m thick; Fig. 3) and is mainly composed of clastic sediments. It consists of fine deposits in the western part of the Domain and alternating sand, conglomerate and marl in the eastern part showing a coarsening up trend (Figs. S4 and S5). New biostratigraphic analyses reveal dinoflagellate cyst and calcareous nannofossil assemblages in the fine deposits indicating an open marine environment (Figs. 3, S3a, S4 and S5). Based on well data analyses, we interpret the sand, conglomerate and marls alternation as the proximal part of the system composed of fluvio-deltaic deposits as suggested by the occurrence of continental species (Hydrobia, snail; Gaudant and Guillevin, 1979; Cavalier et al., 1984; Dumas, 1988; Fontes et al., 1996). Mainly known from subsurface data, our fieldwork in the Crest and Monjoyer areas also documents some outcrops of red marls of alluvial plain and conglomeratic channels that belong to the fluvial part of the Formation marneuse intermédiaire (Fig. S7). Interpretation of seismic data shows downlap geometries (deltaic clinofolds). The whole Formation marneuse intermédiaire is therefore interpreted from east to west as a prograding sequence (Fig. 4). As for Unit I, the sedimentary material is only of local origin (Latreille, 1969). New biostratigraphic



**Fig. 5.** (A) Seismic interpretation of the 82SE4B profile, located west of the Montmirail Domain. See location on Fig. 2b. (B) Synthetic log of the Montmirail Domain. Abbreviations: BCI=Barre calcaire inférieure; BCS=Barre calcaire supérieure; MMMC = Miocene Molasse Marine Cycle.



**Fig. 6.** Montmirail Domain carbonate-evaporitic lake facies with evidence of emersion. (A) Stacked limestone beds. (B) Mud cracks on the top surface of A. (C) detailed section view of B. (D, E) travertine beds. (F) Domed structure showing an alternation of thin carbonate and organic matter-rich marl typical of algae concretions (stromatolite).



**Fig. 7.** Montmirail Domain marine and deltaic facies. (A) Offshore grey to yellow marls. (B) Hummocky cross stratification with interpretation. (C) Northward deltaic progradation showing basal sand/marl alternations of pro-delta and top sandy facies with clinofolds of the delta front. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

analyses of pollen and calcareous nannofossil (NP23; **Figs. 3, S2 and S3a; Table S1**) assemblages indicate that the Formation marneuse intermédiaire could cover the middle and late parts of the Rupelian. The fourth unit, the Formation évaporitique supérieure (**Fig. 3**), is unconformable on the previous one as indicated by seismic reinterpretation (**Fig. 4**) and marks the return to a dominant playa-lake sedimentation in the entire Domain. As Unit II, it shows strong thickness variations (up to 500 m thick; **Fig. S4**) and the same facies distribution: salt and anhydrite deposits in the central part of the Domain and limestones in the outer part. However, the absence of marine criteria would indicate that the Formation évaporitique supérieure deposited in an evaporitic lake (**Dumas, 1986, 1988**). Our pollen analysis was unsuccessful (**Table S1**), so this fourth formation remains poorly dated and is roughly associated to the Rupelian/Chattian transition (**Triat and Truc, 1974**). The fifth and last unit of the Bas-Dauphiné Domain, named the Formation marno-sableuse terminale (up to 1000 m thick; **Fig. 3**), is transitional with the previous one. New well data interpretation indicates the same facies as Unit III: fine deposits of marine environment (with dinoflagellate cyst assemblages) and fluvio-deltaic sand-marl alternations organised in two marine-continental Cycles (**Figs. 3, S4 and S5**). Unlike Unit III, the proximal-distal profile is distributed from east to west. **Latreille's (1969)** petrological work showed the predominance of epidote-glaucophane-garnet minerals in the Formation marno-sableuse terminale, which is interpreted as Internal Alpine Nappes inputs (**Figs. 3 and S5**). This constitutes the first occurrence of Alpine material in the Upper Rhône Valley. New pollen analyses indicate that the Formation marno-sableuse terminale is of late Chattian age and could even extend to the Aquitanian (**Figs. 3, S2, S4 and S5; Table S1**).

We distinguish four depositional sequences for the formations of the Bas-Dauphiné Domain (**Fig. 3**). The Formation infrasalifère forms the first one and corresponds to a transgressive half-sequence (alluvial fan to lacustrine environments). Despite the lack of biostratigraphic data, we consider it to be of Priabonian age based on the age of the overlying Unit II. We interpret the strong unconformity and the brutal change of facies between Unit I and II as a sequence boundary. The second depositional

sequence therefore corresponds to Unit II (lacustrine (TST) to marine (MFS)) as the TST and Unit III (marine to fluvio-deltaic) as the HST. It ranges from early to late Rupelian. We interpret the strong unconformity and the brutal change of facies between Unit III and IV as a sequence boundary. The third depositional sequence therefore corresponds to Unit IV (lacustrine) as the TST and the first marine-continental Cycle of Unit V (marine (MFS) to fluvio-deltaic) as the HST. It ranges from latest Rupelian to late Chattian. The fourth depositional sequence corresponds to the second marine-continental Cycle of the Formation marno-sableuse terminale (marine (TST and MFS) to fluvial (HST)). Based on the age of the underlying unit, we consider that this last depositional sequence is of Aquitanian age.

#### 4.1.2. Montmirail Domain

The Montmirail Domain forms a small (4 km long) NW-SE arc-shaped band of Paleogene formations located north of the Mormoiron Domain (**Figs. 1 and 2b**). It borders the Dentelles de Montmirail Mesozoic massif that constitutes the western extension of the Ventoux and Baronnies massifs into the middle Rhône Valley (**Figs. 1 and 2b**). The Montmirail Domain development is classically related to the European Cenozoic Rift System event. According to previous studies, Paleogene *syn*-tectonic sedimentation took place in the NE-SW axis graben of Nîmes-Pujaut-Théziers, bordered respectively to the north and the south by the Nîmes and Pujaut faults (**Goguel, 1938, 1963; Brasseur, 1962; Flandrin, 1966; Emre, 1977; Casagrande, 1985; Casagrande et al., 1989**). The basin was later deformed and uplifted by salt movement of the surrounding Suzette Triassic diapir in response to Neogene Alpine tectonic loading (**de Laparent, 1940; Brasseur, 1962; Fahy, 1965; Emre, 1977; Casagrande, 1985; Casagrande et al., 1989**). The Montmirail Domain has been poorly documented. Here we bring new tectono-sedimentary descriptions of the Paleogene formations based on extensive fieldwork, biostratigraphy analyses and seismic reinterpretation (S2SE4B profile; **Fig. 5a**).

We divide these formations into two sedimentary cycles (Cycles I and II (**Fig. 5b**)) showing facies and depositional environment similarities. They both begin with local-sourced conglomeratic channels and red

marls of alluvial fan that progressively retrograde to stacked limestones and gypsum of evaporitic lake (Figs. 5b, 6a and S8). Carbonate beds show algae concretions and mud cracks indicating containment or drying up of the system (Fig. 6). These carbonate-evaporitic deposits form two distinct ridgelines in the Montmirail Domain landscapes and

have a particular signature (strong amplitude) on seismic profiles (Fig. 5a). We name them the Barre calcaire inférieure (Cycle I) and the Barre calcaire supérieure (Cycle II) (Figs. 5 and S8). Evaporitic sedimentation evolved to grey marl deposits of a calm depositional environment that we consider of lacustrine origin for Cycle I. However, the

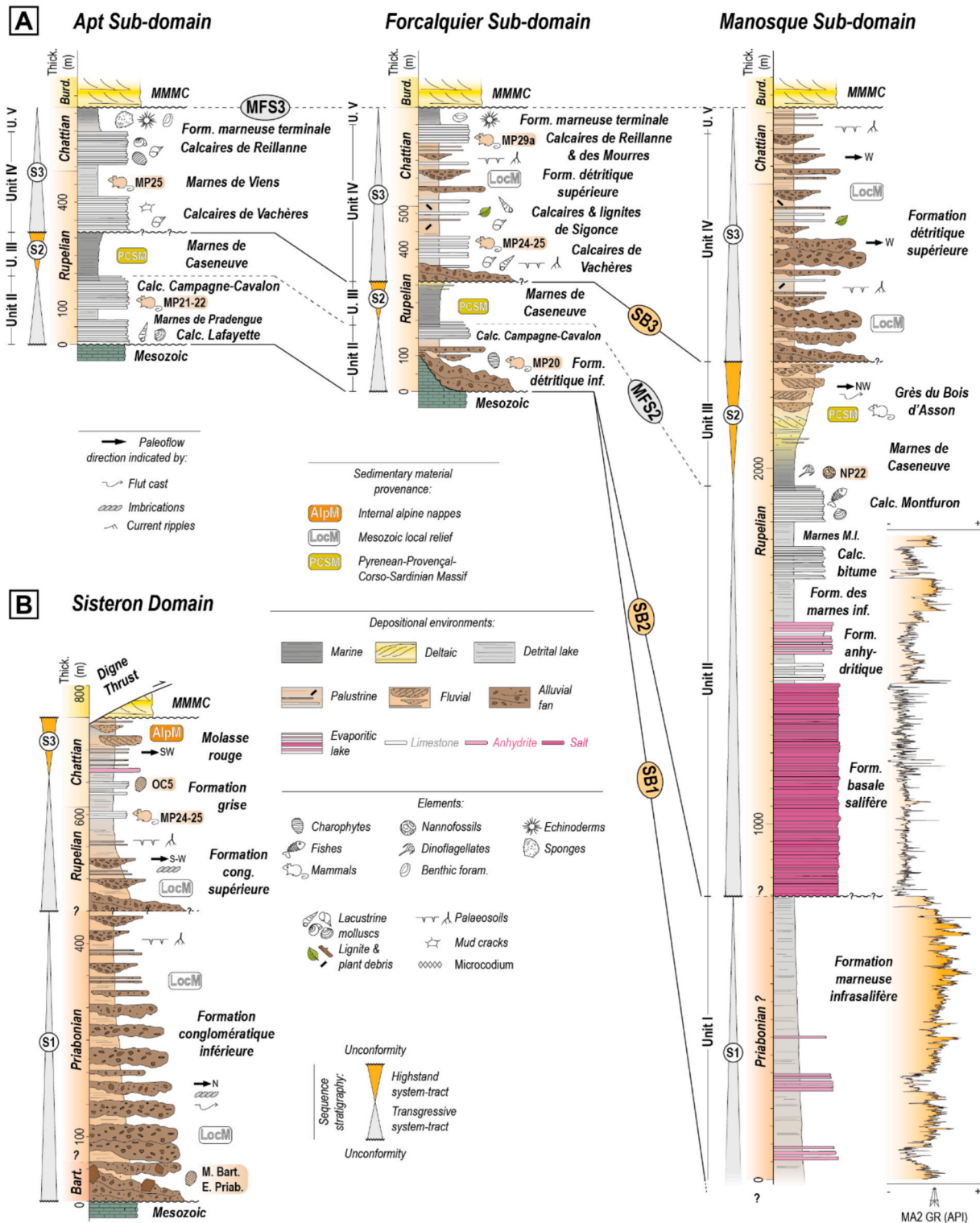


Fig. 8. (A) Synthetic logs and stratigraphic correlations between the Durance sub-domains (Apt, Forcalquier and Manosque sub-domains) adapted from Lesueur (1991) and this work. (B) Synthetic log of the Sisteron Domain is adapted from Gigot et al. (in prep), Hamon (2022) and this work. Abbreviations: Calc. = Calcaires; Calc. bitume = Calcaires à indices de bitume; Cong. = conglomératique; Form. = Formation; inf. = inférieure; Marnes M.I. = Marnes et gypses de la Mort d’Imbert; MMMC = Miocene Molasse Marine Cycle.

occurrence of intercalated sandy beds with wavy structures as well as dinoflagellate cyst and calcareous nannofossil assemblages in the fine deposits of Cycle II clearly demonstrate marine conditions (Figs. 5b, 7a, b and S8). In both cycles, fine sediments progressively evolved into fluvio-deltaic formations (Figs. 7c and S8) that form prograding sequences. Fluvial deposits in Cycle I consist of alternating meandering channels, multicolour palaeosoils, red marls of alluvial plain and occasional carbonate beds of ephemeral lakes with algae concretions (Figs. 5b, S8, S9a, S9b and S9c). Fluvial deposits in Cycle II consist of massive amalgamated sandy bodies with both channels and flute casts migrating in the same direction (SW), indicating a straight river type system (Figs. 5b, S8, S9d and S9e). Mainly sourced by the local Mesozoic terrains (Brasseur, 1962; Casagrande, Emre, 1977; Emre and Truc, 1978; Casagrande, 1985; Casagrande et al., 1989), the fluvial formations of the Montmirail Domain also contain allochthonous material coming from the Internal Alpine Nappes (Casagrande et al., 1989; Huet, 2024). Seismic line shows that between cycle I and II, then between Cycle II and Burdigalian Miocene Marine Molasse Cycle, two clear erosional surfaces can be distinguished (Fig. 5a). In order to propose a chronostratigraphic framework that does not exist in the literature, we performed biostratigraphic analyses in the lacustrine (Cycle I) and marine (Cycle II) grey marls. They show typical pollen grains ascribed to plants of cold and dry climate and calcareous nannofossil assemblages of NP21–22 biozones, both indicating an early Rupelian age (Figs. 5b, S2, S3a and S8; Table S1).

The two sedimentary cycles of the Rupelian Montmirail Domain form two depositional sequences (Figs. 5 and S8). The TSTs record the transition from alluvial fan to lacustrine (Cycle I) or marine (Cycle II) environments, in which we place the MFS. Fluvio-deltaic formations correspond to the HSTs.

Seismic interpretation reveals also that the Montmirail Domain can be divided into two depocenters separated by a salt diapir that constitutes an underground extension of the easterly-located Suzette Triassic diapir (Fig. 5a). Inclined reflectors and halokinetic sequences highlight that the diapir was active during the basin development, which caused a progressive migration of the southern depocenter toward the south until Miocene time. The Triassic feeder is located below Jurassic and Cretaceous formations that show growth-strata, indicating salt movement long before Cenozoic sedimentation (Fig. 5a). The salt layer below the Rupelian Montmirail Domain is therefore a canopy that is very thin or even absent under the southern depocenter (Fig. 5a). Except for a few minor normal faults, the seismic data do not reveal any major extensive faults at the edges of the basin that might indicate a graben geometry typical of European Cenozoic Rift System basins. The Rupelian basin of Montmirail could correspond to a salt related mini-basin with local subsidence, but with distant connexion as demonstrated by the presence of Alpine derived sediments in its fluvial formations.

#### 4.1.3. Mormoiron Domain

The Mormoiron Domain is located between the Montmirail Domain to the north and the Vaucluse plateau to the south (Figs. 1 and 2b). It forms a 20 km long arc-shaped band of Eocene-Oligocene formations that borders the western slope of the Ventoux-Lure massif. Here we use mainly literature data and a few results based on new fieldwork and palynological analyses to describe the Paleogene formations of the Mormoiron Domain.

In this study, we divide them into four units (Unit I to IV) of mainly clastic or carbonate-evaporitic composition. Unit I consists mainly of lacustrine green clay and lignite named the Complexe argileux vert de Mormoiron (up to 100 m thick; Fig. 3) (Triat and Truc, 1972). To the north of the Domain, fine deposits evolve laterally to conglomerates of alluvial fan named the Conglomérats de Crillon-le-Brave (Figs. 3 and S10) (Triat and Truc, 1972). Sedimentary material is only of local origin and contains massive clasts (up to 2 m large) of Mesozoic carbonates coming directly from the near Ventoux massif (Fig. S3b) (Triat and Truc, 1972). Our fieldwork revealed a global fining upward trend within this

first unit, indicating retrogradation (Figs. 3 and S10). Mammal assemblages found at the base of the Complexe argileux vert de Mormoiron date the beginning of sedimentation to the early Priabonian (Fig. 3) (*Palaeotherium medium euzetense*, *Plagiolophus annectens* (Montenat, 1968; Triat and Truc, 1972), MP17 (Schmidt-Kittler et al., 1987; Perales-Gogenola et al., 2022)). The transition between Units I and II is not documented and has not been found in the field. The second unit of the Mormoiron Domain is of carbonate-evaporitic composition and groups three formations of a playa-lake environment (Fig. 3) (Triat and Truc, 1972; Triat and Truc, 1974) (from the base to the top): (1) the Complexe argileux blanc de Blauvac (40 m thick), which consists of gypsum, dolomites and white clay alternations; (2) the Calcaires du Pâtis and (3) the Calcaires à *Brotia laurea* (50 m thick in total), which both consist of stacked carbonate beds. New pollen analyses date the base of the Complexe argileux blanc de Blauvac to the very late Priabonian (Figs. 3 and S2; Table S1). The base of the Calcaires à *Brotia laurea* is from early Rupelian (MP21) based on mammal assemblages (Triat et al., 1971; Maridet et al., 2013) (Fig. 3). The transition between Units II and III is not documented and has not been found in the field. The third unit is named the Sables de la Valette-de-Pernes (or Sables et argiles d'Unang) (up to 80 m thick; Fig. 3). It consists of clastic sediments, marls alternating with palaeosoils, and sandy meandering channels of fluvial environment showing a global southward palaeoflow direction (Triat and Truc, 1972; Costeur et al., 2019). Mammal assemblages give a middle Rupelian (MP23) age to Unit III (Fig. 3) (Murs mammal deposit; Costeur et al., 2019). Unit IV overlies unconformably the underlying Sables de la Valette-de-Pernes (Triat and Truc, 1974). Just like Unit II, it is of carbonate-evaporitic composition and groups three formations of lacustrine origin (Fig. 3) (Triat and Truc, 1974) (from the base to the top): (1) the Calcaires et gypses de l'Isle-sur-Sorgue (up to 80 m thick), which consists of alternating fossil-rich carbonate and gypsum beds; (2) the Calcaires et marnes de Vénasque (up to 40 m thick), which consists of alternating green marls and white limestones; (3) the Calcaires à *Helix ramondi* (up to 30 m thick), which consists of massive stacked white limestones. The occurrence of *Cadurcotherium nouleti* in the Calcaires et gypses de l'Isle-sur-Sorgue (Roman and Joleaud, 1909; Roman, 1912) constrains the age of this formation to the very late Rupelian (MP25) (Crusafont Pairó and Aguirre, 1973) (Fig. 3). The upper part of Unit IV is undated but considered to be of Chattian age based on the occurrence of *Helix ramondi* (Rey, 1966, 1968). The Burdigalian deposits of the Miocene Marine Molasse Cycle unconformably succeed Unit IV in its upper part.

We divide the Paleogene formations of the Mormoiron Domain into three depositional sequences (Fig. 3). Unit I forms the first one and corresponds to a transgressive hemicycle (alluvial fan to lacustrine environments) during the Priabonian. We consider the transition between Units I and II as a sequence boundary, based on the comparison with the Bas-Dauphiné Domain stratigraphy. Therefore, the second depositional sequence corresponds to Unit II as the TST (lacustrine) and Unit III as the HST (fluvial) and ranges from uppermost Priabonian to middle-late Rupelian. Unit IV forms the third depositional sequence and corresponds to a transgressive hemicycle (lacustrine) that ranges from latest Rupelian to undifferentiated Chattian.

#### 4.1.4. Durance Domain

The Durance Domain is located in an intermediate position east of the Rhône Valley and only a few kilometres west of the Sisteron Domain (Fig. 1). Because of palaeogeographic and stratigraphical similarities, we place it into the Rhône Valley Basin group. The Durance Domain forms a triangular shaped area of approximately 600 km<sup>2</sup> delimited to the north and the west by the Ventoux-Lure massif and to the south and east by the Durance alluvial plain (Fig. 1). It actually corresponds to three connected sub-domains forming a series of SSW-NNE synclines and anticlines: The Apt, Forcalquier and Manosque sub-domains, to the west, the northeast and the southeast, respectively (Fig. 1). Paleogene sedimentation has been associated with the Pyrenean-Provençal orogeny

and the European Cenozoic Rift System event, which led to the formation of *E-W* compressive structures (Luberon massif) in the central area (Guyonnet-Benaize et al., 2015) and a large *NNE-SSW* subsiding trench (3000 m deep) in the eastern part of the Domain controlled by the Durance Fault (Lesueur, 1991). The Durance Domain offers good outcrop quality and subsurface data acquired for gas storage (Manosque storage site). Here we mainly use published data, but also additional results coming from the reinterpretation of subsurface data to describe the Durance Domain.

We divide the Paleogene formations into five main units (Unit I to V) of clastic or carbonate-evaporitic composition (Fig. 8a). At the base, Unit I, named the Formation marneuse infrasalifère, is restricted to the Manosque area and has been identified only in the Manosque 2 (Ma2) well (Destombes, 1962). It consists of 800 m thick dark marls with anhydritic intercalations (Fig. 8a). The age and depositional environment of this first unit have never been documented but probably consists of lacustrine deposits. The progressive increase in gamma ray values (MA2 well; Fig. 8a) between the base and the top of Unit I suggests a fining-up trend in sedimentation that we interpret as retrogradation. Unit II mainly consists of salt, anhydrite and limestones of lacustrine origin with marl intercalations (Destombes, 1962; Triat and Truc, 1974; Freydet and Truc, 1975; Freydet and Plaziat, 1982; Lesueur, 1991) (Fig. 8a). It is present throughout the entire Durance Domain but shows significant facies/thickness variations. In the Forcalquier sub-domain. The Unit II starts with retrograding alluvial fan deposits (Formation détritique inférieure; 0 to 100 m thick) of early Rupelian age (MP20; Gigot et al., 1982; Fig. 8a), which fill local depressions in an irregular palaeotopography (Lesueur, 1991). This formation constitutes the only significant clastic input of Unit II. In the Manosque sub-domain, where sedimentation has been associated to the hanging wall motion of the Middle Durance Fault, it begins with 900 m of evaporitic and marly deposits (Formation basale salifère, Formation anhydritique, Formation des marnes inférieures; Destombes, 1962) known only through well data. Well log interpretation (gamma ray signal of the MA2 well; Fig. 8a) shows that they overlie Unit I with a sharp contact. These formations are then covered by carbonate and marly deposits named the Calcaires à indices de bitume (90 m thick) and Marnes et gypses de la Mort d'Imbert (50 m thick) in the Manosque sub-domain, and Calcaires de Lafayette (100 m thick) and Marnes de Pradengue (25 m thick) in the Apt sub-domain (Lesueur, 1991) (Fig. 8a). Unit II ends with the Calcaires de Montfuron (Manosque area; 100 m thick) and Calcaires de Campagne-Cavalon (Forcalquier and Apt areas; 20 to 100 m thick) dated of early Rupelian age (Céreste mammal deposit; MP21–22; Schmidt-Kittler and Storch, 1985) (Fig. 8a). Unit III is composed of clastic sediments and is present in the whole Durance Domain (Fig. 8a). Transitional with Unit II, it consists mainly of dark marls (Marnes de Caseneuve; 100 m thick in Apt and Forcalquier areas) of an open marine environment indicated by dinoflagellate cysts and calcareous nannofossil assemblages recently dated to the early Rupelian (NP22; Suc et al., 2025) (Fig. 8a). In the Manosque sub-domain, fine grained deposits evolve laterally and vertically to sandy and conglomeratic sediments (Formation du Bois d'Asson; up to 350 m thick) which record a fluvio-deltaic progradation toward the northwest, fed by the Pyrenean-Provençal-Corso-Sardinian massif (Lesueur, 1991) (Fig. 8a). Unit IV is of mixed carbonate-evaporitic/clastic sedimentation and outcrops very well throughout the entire Durance Domain. It shows a great sedimentary facies diversity distributed on an *E-W* proximal-distal playa-lake depositional profile that has been extensively documented by Lesueur (1991) (Fig. 8a). Proximal alluvial fan (Formation détritique supérieure; up to 650 m thick) and palustrine deposits (Calcaires de Vachères (proximal part), Calcaires et lignites de Sigonce; up to 450 m thick) are restricted in the Manosque sub-domain and the eastern part of the Forcalquier area (Fig. 8a). Sedimentary thicknesses and facies suggest some control of the Middle Durance Fault. The detrital systems are composed of Mesozoic clasts from the Valensole High (Lesueur, 1991). These deposits evolve westward (Apt sub-domain and western part of the Forcalquier area) to

lacustrine marls and limestones that constitutes the distal part of the system (Calcaires de Vachères, Marnes de Viens, Calcaires de Reillanne, Calcaires des Mourres; up to 300 m thick) (Fig. 8a). Unit IV ranges from late Rupelian to late Chattian age (Fig. 8a). From base to top, biostratigraphic constraints are: La Blache and Saint-Martin-de-Castillon mammal deposits in the Calcaires de Vachères (MP24–25; Huguency, 1994; Gigot et al., 1982; Maridet et al., 2010), Sigonce mammal deposits in the Calcaires et lignites de Sigonce (MP24–25; Ducreux et al., 1985; Escarguel et al., 1997), Aubenas-les-Alpes mammal deposits in the Marnes de Viens (MP25; Ménouret et al., 2015; Maridet et al., 2019) and La Ferme Pavoux mammal deposits in the Calcaires des Mourres (considered to be into MP30 biozone by Ménouret (2014), replaced within MP29 biozone after biostratigraphic revision (Bolliger, 2013; Freudenthal and Martín-Suárez, 2016)) (Figs. 8a and S3b).

Unit V, that we name the Formation marneuse terminale, is in continuation with Unit IV and displays a partly similar lithology but contains also evidence of a marine incursion (Lesueur, 1991). Alluvial fan bodies are still present in the Manosque area on the edge of the Domain, but in the Apt and Forcalquier sub-domains, lacustrine deposits progressively evolved upward to clays and limestones of lagoon environments containing marine fauna (echinoderms, sponges and bryozoans; Lesueur, 1991) (Fig. 8a). Unit V is relatively thin (50 m thick maximum) and still undated, but its continuation with the underlying Calcaires de Reillanne et Calcaires des Mourres may indicate a possible late Chattian age (MP29). The late Burdigalian transgressive sandstones of the Miocene Marine Molasse Cycle finally overlie this last unit of the Durance Domain (Lesueur et al., 1990; Rubino et al., 1990).

We divide the Paleogene formations of the Durance Domain into three depositional sequences (Fig. 8a). The Formation marneuse infrasalifère of Unit I is the first one and corresponds to a transgressive hemicycle. We consider it to be of Late Eocene age based on the age of the overlying Unit II. Despite the absence of seismic data to confirm it, we interpret the brutal shift of well log signal and lithology between Units I and II as a sequence boundary based on similarities with the Bas-Dauphiné Domain. The second depositional sequence therefore corresponds to Unit II (alluvial fan to lacustrine; TST) and III (marine (MFS) to fluvio-deltaic progradation; HST) and ranges from early to early late Rupelian. The third depositional sequence corresponds to a transgressive hemicycle that includes Unit IV (alluvial fan to lacustrine deposits; TST) and V (shallow marine environment; MFS) and ranges from latest Rupelian to late Chattian.

#### 4.1.5. Provence Domain

The Provence Domain is located 30 km east of the Rhône Valley and is surrounded by the Mesozoic massifs of the Provence (Fig. 1). The Domain can be divided into two sub-domains separated by the *E-W* Etoile Thrust (Fig. 1): (1) the Aix-en-Provence sub-domain to the north that forms a *N-S* 20 km long syncline, and (2) the Marseille-Aubagne sub-domain to the south, which corresponds to the current location of the Marseille urban area (Fig. 1). No fieldwork was done in this Domain, and all following descriptions are coming from available literature data.

We can summarise the Provence Paleogene formations in three superimposed sedimentary cycles (cycles I to III) showing the same lithological and depositional environment patterns (Fig. 3). The first two are only preserved in the Marseille-Aubagne sub-domain, whereas the last one is restricted to the Aix-en-Provence area. Each cycle begins with carbonate and gypsum deposits (up to 50 m thick) of evaporitic lake or lagoon associated with lignite and local sourced breccia deposits of palustrine environments on the margin (cycle I: Calcaires à Striatelles; cycle II: Formation de l'Estaque; cycle III: Marnes et Conglomérats de Saint-Anne, Calcaires et Marnes des stations d'essence, Calcaires et Marnes à gypse d'Aix) (Nury, 1977, 1987; Gaudant et al., 2018). Then comes clastic sedimentation always starting with grey to red marls of a marine environment with thin sandy beds (cycle I: Formation de St-Marcel; cycle II: Formation de St-Henry/St-André; cycle III: Marnes du Puy-du-Roy), followed by sandy to conglomeratic fluvial and alluvial fan

deposits (cycle I: Formation de la Destrousse; cycle II: Formation des Arnavaux/Ste-Marthe, Poudingues de Marseille; cycle III: Sables de Fignons) (Nury, 1977, 1987). The clastic parts of each cycle are therefore included in prograding sequences. The transition from one to another cycle is unclear because of the lack of descriptions in the literature. Apart from the basal breccia containing local Mesozoic clasts, the detrital units of cycle I and II are sourced from the southerly-located Pyrenean-Provençal-Corso-Sardinian massif and shows northward palaeoflow directions (Arlhac et al., 1973; Nury, 1977, 1987; Guieu et al., 1987). On the contrary, heavy mineral studies on the Sables de Fignons of cycle III indicate an Internal Alpine Nappes provenance and attest to the first arrival of Alpine material in the Provence Domain (Nury, 1987). The revision of biostratigraphic assemblages dates Cycle I from early (MP22; Figs. 3 and S3b) to early late Rupelian (OC2; Figs. 3 and S3b). Cycle II is supposed to initiate during the late Rupelian (Nury, 1987) without solid biostratigraphic data to confirm it. The top of cycle II, the Formation des Arnavaux/Ste-Marthe is classically considered to be of early to middle Chattian age (Thaler, 1966; Hugueney and Truc, 1976; Vianey-Liaud, 1979; Riveline, 1986; Nury, 1987). Here we move it to the late Chattian (MP28) based of revision of biostratigraphic data (Figs. 3 and S3b). The Poudingues de Marseille, placed in the late Chattian by Nury (1987), is also moved to the early Aquitanian (MN1; Figs. 3 and S3b). Cycle II therefore extends between latest Rupelian and late Chattian/early Aquitanian. Cycle III is poorly dated. The occurrence of *Rhodanomya schlosseri* in the Marnes et Conglomérats de Saint-Anne (Nury, 1987) would however place its base in the early Aquitanian (MN1; Brinkkötter and Martin, 2015). Additional formations that are not part of the three cycles are preserved along the Mediterranean shoreline in la Nerthe/Carry-le-Rouet area, 15 km west of Provence Domain. It consists at their base of fluvial sandstones and conglomerates, which progressively evolve to shallow marine deposits alternating marls, bioclastic limestones and reef constructions (Andreieff et al., 1972; Maurel-Ferrandini, 1976; Nury, 1987; Oudet et al., 2010). These formations are dated to the late Chattian-early Aquitanian interval (*Sphenolithus delphix* (Oudet et al., 2010); i.e. 23.35–23.02 Ma based on palaeomagnetic calibration (Fabbrini et al., 2019)) and could therefore be a lateral equivalent of the Marnes et Conglomérats de Saint-Anne. Shallow marine environment will persist in the Provence Domain until Tortonian as part of the Miocene Marine Molasse Cycle (Rubino et al., 2015).

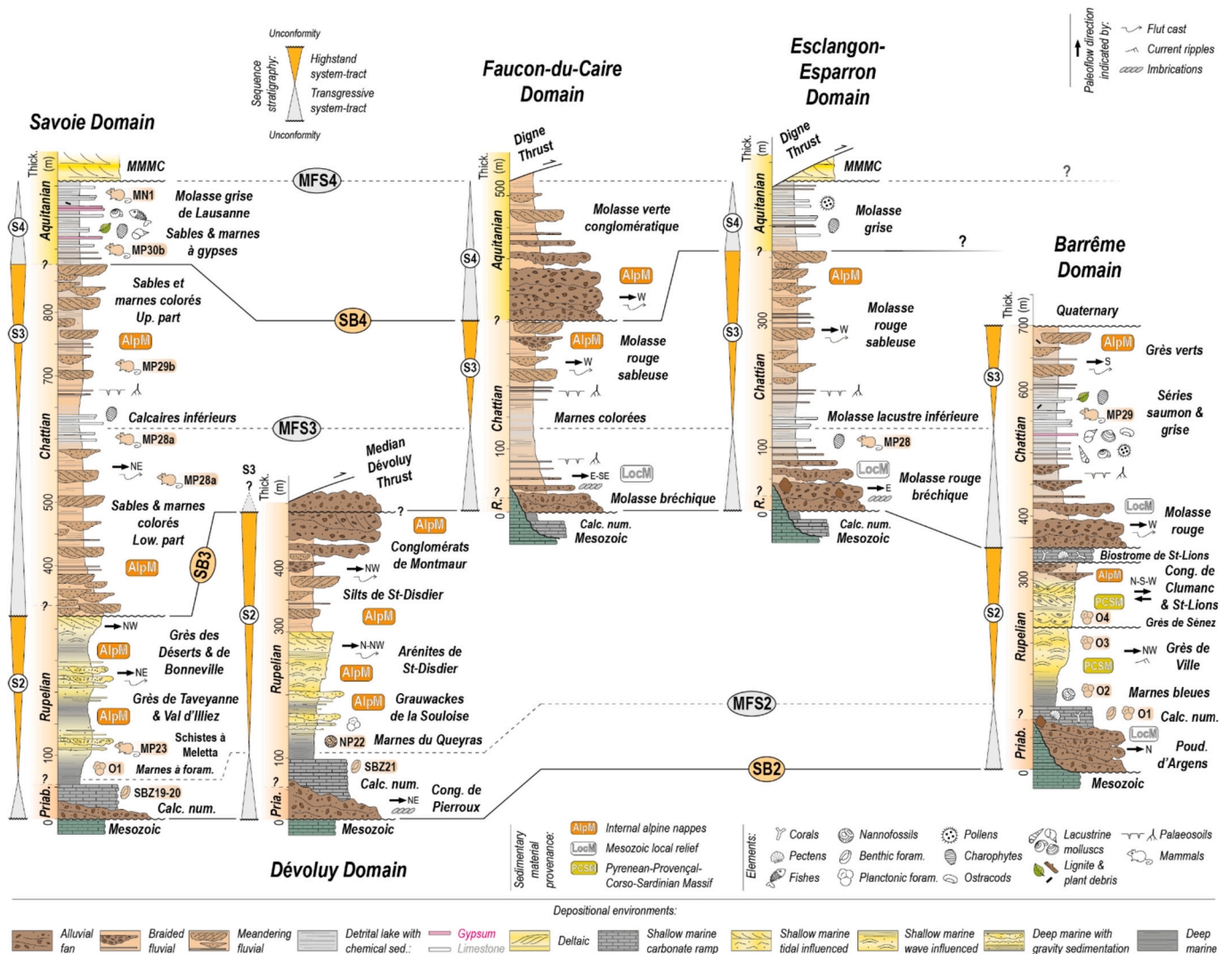
The three cycles that compose the Paleogene formations of the Provence Domain are here interpreted as depositional sequences as they record the transition from lacustrine to marine (TST) and finally fluvio-deltaic (HST) environments (Fig. 3).

#### 4.1.6. Camargue Domain

The Camargue Domain is the southernmost of all Rhône Valley domains (Fig. 1). Covering an area of approximately 1500 km<sup>2</sup>, it is also one of the largest domains after the Bas-Dauphiné. It is bounded by the Garrigues region to the north and the west, the Alpilles massif to the east, and extends under the Mediterranean Sea to the south (Fig. 1). The structure of the Camargue Domain has been related to two geodynamic events during the Cenozoic: (1) the Upper Cretaceous to Eocene Pyrenean-Provençal orogeny, which led to the formation of ~ E-W folds and thrusts and sinistral strike-slip, and (2) the Oligocene to Early Miocene opening of the Gulf of Lion (Arthaud and Séguret, 1981; Valette and Benedicto, 1995; Benedicto et al., 1996; Mauffret and Gorini, 1996; Séranne et al., 2021). These events induced the motion of the normal/strike-slip Nîmes Fault, which controlled the deposition of a thick sedimentary series along a NE-SW axis. The Camargue Domain is the thickest domain of the entire Rhône Valley with a maximum of 4000 m of Paleogene sediments identified in the Pierrefeu 1 well (PRF1; Fig. S11). Most of the Cenozoic formations are buried below Quaternary alluvial deposits of the Rhône River and are therefore documented only through subsurface data acquired during petroleum exploration and gas storage campaigns. All boreholes are concentrated in the central part of

the Camargue Domain. Consequently, the other parts remain poorly documented. Here we use literature data and new results based on the reinterpretation of well data to describe the Paleogene formations of the Camargue Domain.

According to Gorini (1993), Valette (1991) and Semmani (2022), these formations are classically divided into three units (Série grise, Série rouge and Série calcaire-salifère; Fig. S11), each comprising several sub-units that are not always identified or described in the literature, causing confusion. Based on the reinterpretation of well data, we choose to adopt our own nomenclature and divide the formations of the Camargue Domain into five units (Unit I to V) of clastic or carbonate-evaporitic composition (Figs. 3 and S11). We name Unit I the Formation détritique basale (Fig. 3). It reaches 800 m thick above the Mesozoic basement in the PRF1 well and consists of clastic deposits alternating dominant grey marl, sand and conglomeratic beds interpreted by Semmani (2022) as turbidites deposited in a deep lake environment. New well log interpretations indicate that Unit I has a distinct gamma ray signal showing high values and organised in a fining up sequence that we interpret as retrogradation (Fig. S11). Sedimentary material is only of local origin and contains Mesozoic carbonate clasts from nearby sources (Semmani, 2022). Unit I as a whole is poorly dated. Pollen assemblages found at its very top indicate an age of the Priabonian-Rupelian transition (Semmani et al., 2023) (Figs. 3 and S11). We name Unit II the Formation calcaire inférieure (up to 600 m thick; Fig. 3). New results show that its transition with the underlying Unit I corresponds to a brutal shift in both well log and lithological data, related to a rapid decrease of terrigenous supply. Unit II consists mainly of stacked carbonates beds with few marl intercalations that display low and constant gamma ray values (Figs. 3 and S11). Poorly constrained, the depositional environment is supposed to be lacustrine. The Formation calcaire inférieure is completely undated. In its upper part, it is progressively dominated by clastic deposits of alternating marl, sand and conglomerate, with sometimes isolated limestone and anhydrite beds. Semmani (2022) and Semmani et al. (2023) provide detailed sedimentological descriptions of these deposits based on core data analyses of the Gallician 9 well (Fig. S11). In this study, we associate these clastic deposits with the third unit of our nomenclature that we name the Formation détritique intermédiaire (up to 250 m; Fig. 3). The Formation détritique intermédiaire records a progressive coarsening up trend, from marls of a calm environment to fluvio-deltaic formations, indicating a progradation. The occurrence of dinoflagellate cysts and glauconitic sands in the marls indicate a marine depositional environment for unit III (Semmani et al., 2023). In the marine delta, sedimentary material contains quartz, silicate and metamorphic minerals coming from the dismantling of the Pyrenean-Provençal-Corso-Sardinian massif (Semmani, 2022). Unit III is well dated by U/Pb analyses of calcite veins: the deltaic and fluvial parts are respectively of middle (31.27 ± 1.90 Ma) and late (27.88 ± 1.71 Ma) Rupelian age (Semmani et al., 2023) (Figs. 3 and S11). We name Unit IV the Formation évaporitique supérieure (Fig. 3). It overlies unconformably the Formation détritique intermédiaire (Valette and Benedicto, 1995; Benedicto et al., 1996). The Formation évaporitique supérieure shows strong thickness variations (300 to 900 m thick; Fig. S11) and consists of alternating limestone, dolomite, anhydrite and salt deposits with occasional marl and sand beds. Although dominated by continental fauna (gastropods in Vauvert 1 well report), Semmani et al. (2023) has recently documented dinoflagellate cyst assemblages in the basal part of the unit, which would indicate a marine environment. Based on pollen records, Unit IV was assigned a Chattian age (Semmani et al., 2023) (Figs. 3 and S11). We name the fifth and last formation of the Camargue Domain the Formation détritique terminale (Fig. 3). It is transitional with the Formation évaporitique supérieure. The Formation détritique terminale consists of two superimposed cycles showing clayey sediments in their basal parts and grey clay with glauconitic sand at their tops. The occurrence of glauconite and dinoflagellate cysts would indicate a marine environment (Semmani et al., 2023). New well log reinterpretations show that



**Fig. 9.** Synthetic logs and stratigraphic correlations between Frontal Basin (Savoie, Dévoluy, Faucon-du-Caire, Esclangon-Esparron and Barrême domains). Synthetic logs are adapted from: Savoie: Charollais et al. (2007) and Girault (2020). Dévoluy: Meckel et al. (1996), Meckel (1997) and this work. Faucon-du-Caire: Tremblay (2002) and this work. Esclangon-Esparron: Bialkowski (2002), Held (2011) and this work. Barrême: Evans and Elliott (1999) and Callec (2001). Abbreviations: Calc. num. = Calcaires nummulitiques; Cong. = Conglomerats; foram. = foraminifères; low. = lower; MMMC = Mioocene Marine Molasse Cycle; Poud. = Poudingues; up. = upper.

sandy and clayey levels are respectively matching with coarsening and fining upward trends of the gamma ray signal (Fig. S11), probably related to shallowing and deepening of the system. The Formation détritique terminale is still completely undated and could be of late Chattian to Aquitanian age (Figs. 3 and S11). Unit V is unconformably overlain by the Burdigalian deposits of the Mioocene Marine Molasse Cycle.

We divide the Paleogene formations of the Camargue Domain into four depositional sequences (Figs. 3 and S11). Unit I is the first one and corresponds to a transgressive hemicycle of probably Priabonian age. Despite the absence of seismic data to confirm it, we interpret the brutal shift of well log signal and lithology between Units I and II as a sequence boundary based on similarities with the Bas-Dauphiné Domain. The second depositional sequence therefore corresponds to Unit II (lacustrine) as the TST, and Unit III (marine (MFS) to fluvial environment) as the HST. It ranges from early to late Rupelian. The unconformity between Unit III and IV is interpreted as a sequence boundary. The third depositional sequence therefore corresponds to Unit IV (marine) and the first cycle of Unit V (deep (MFS) to shallow marine). The second cycle of Unit V forms the fourth depositional sequence.

## 4.2. Frontal Basin

### 4.2.1. Savoie Domain

The Savoie Domain is the northernmost and largest of all frontal molasse domains of this study. It forms a 3000 km<sup>2</sup> triangular shaped area that is separated from the Bas-Dauphiné Domain by the Jura folds to the south and extends into the NAFB to the north (Fig. 1). The Savoie Domain has been largely studied and all following descriptions come from literature data.

Sedimentation started to the east of the domain with conglomerates of fan delta and fossil-rich (coral, algae, bryozoan, bivalve and echinoderm) bioclastic limestones of a shallow marine environment named the Calcaires nummulitiques (up to 50 m thick; Fig. 9). The Calcaires nummulitiques developed above a very clear lithophagous perforation surface of Upper Cretaceous sedimentary rocks (Charollais and Wellhauser, 1962; Pairis, 1975; Charollais et al., 1988). Revision of biostratigraphic information results now in a middle to late Priabonian age (SBZ19–20; Figs. 9 and S3c) of this first formation, based on benthic foraminifera assemblages (Charollais et al., 1988). Shallow marine deposits progressively evolved to fine-grained sediments of a deep marine

environment named the Marnes à foraminifères (lower part) and the Schistes à Meletta (upper part) formations (up to 70 m thick; Fig. 9) (Charollais et al., 1988). Charollais et al. (1980) estimated a water depth ranging between 400 and 800 m based on ostracods assemblages. These fine-grained deposits are classically considered to be of early Rupelian age (Charollais et al., 1980; Detraz et al., 1986; Charollais et al., 1988; Doudoux et al., 1992). Here we restrict the Marnes à foraminifères to the early Rupelian (O1) and extend the Schistes à Meletta to the middle Rupelian (MP23), based on an exhaustive revision of biostratigraphic data (Figs. 9 and S3c). The Schistes à Meletta are progressively enriched upward with turbidites. They form massive conglomeratic and sandy formations named the Grès de Taveyenne and Grès de Val d'Illicz (up to 460 m thick; Fig. 9) that are part of the Helvetic nappes and extend northwards to the Central Alps (Charollais et al., 1988; Doudoux et al., 1992). These formations, which are unfortunately poorly dated (early to middle Rupelian age presumed by Charollais et al., 1988), mark the first significant siliciclastic inputs into the Savoie Domain. They show north-eastward palaeoflow directions (Martini, 1968; Charollais et al., 1988) and are mainly sourced by the Internal Alpine Nappes (Martini, 1968; Sawatzki, 1974; Lateltin, 1988; Ruffini et al., 1995). Turbidite deposits progressively evolved to coarsening upward sand and marl alternations with wavy and hummocky cross stratification structures. These sedimentary bodies form the Grès des Déserts and Grès de Bonneville (a few dozen meters thick; Fig. 9) and are interpreted as north-westward prograding deltaic sediments deposited under wave influences (Girault, 2020). The sedimentary material contains strong proportions of epidote and garnet in their heavy mineral assemblages (Latreille, 1969), which could be related to an Internal Alpine Nappes provenance. Due to the lack of biostratigraphic data, these deltaic formations remain undated but are considered to be of late Rupelian age (Girault, 2020). The Grès des Déserts and Grès de Bonneville are unconformably overlain by the Sables et marnes colorés (up to several hundred meters thick; Charollais et al., 2007; Girault, 2020) (Fig. 9), which start the transition to continental sedimentation in the Savoie Domain. This continental unit consists mainly of red marls of alluvial plain with palaeosoils and meandering fluvial channels. These deposits outcrop mainly in the western part of the domain (South Geneva and Bornes Plateau areas). They form two parts (lower and upper parts; Fig. 9) separated by a 70 m thick level of lacustrine sediments showing grey marls and carbonate beds named the Calcaires inférieurs (Rigassi, 1957; Charollais et al., 2007; Girault, 2020). Fluvial channels show northeastern palaeoflow directions and contain sedimentary material coming from the Internal Alpine Nappes (Latreille, 1969; Mange-Rajetzky and Oberhaensli, 1982, 1986; Girault, 2020). According to our biostratigraphic revision, the Calcaires inférieurs and the lower part of the Sables et marnes colorés have an early late Chattian age (MP28a; Figs. 9 and S3c). The upper part of the Sables et marnes colorés extends to the end of the Chattian based on magnetostratigraphic work (Burbank et al., 1992). The Oligocene sedimentation of the Savoie Domain ends with the Sables et marnes à gypses and the Molasse grise de Lausanne (up to 300 m thick; Fig. 9). These formations consist of dominant grey marls of lacustrine and palustrine environments, alternating with carbonate beds and gypsum beds of playa-lake (Kissling, 1974; Charollais et al., 1988; Girault, 2020). Biostratigraphic revision dates these last formations to the end of the Chattian (MP30b) and to the beginning of the Aquitanian (MN1) (Figs. 9 and S3c). The Sables et marnes à gypses and the Molasse grise de Lausanne are unconformably overlain by the shallow marine deposits of late Aquitanian age that belong to the Miocene Marine Molasse Cycle (Beck et al., 1998; Kalifi et al., 2022).

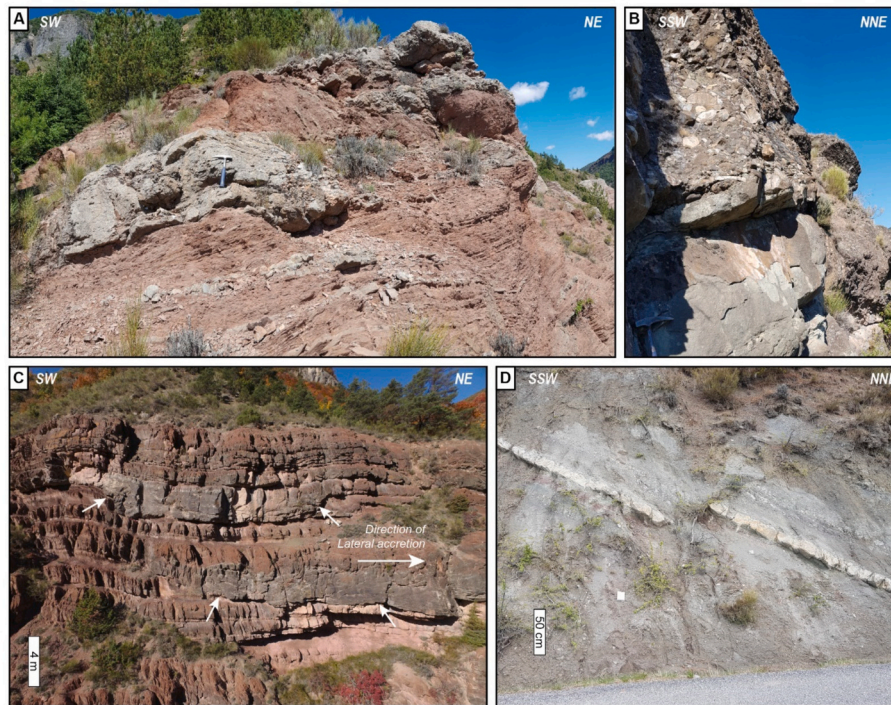
We divide the Paleogene formations of the Savoie Domain into three depositional sequences (Fig. 9). The first one groups the Calcaires nummulitiques, the Marnes à foraminifères (MFS) and the Schistes à Meletta as the TST (shallow to deep marine), and the Grès de Taveyenne, the Grès de Val d'Illicz, the Grès des Déserts and the Grès de Bonneville as the HST (deep marine to deltaic). That sequence ranges from the middle-late Priabonian to the late Rupelian. The second depositional

sequence groups the Sables et marnes colorés and the Calcaires inférieurs (MFS) (fluvial – lacustrine – fluvial). Despite the lack of biostratigraphic data at its base, it probably covers the entire Chattian. The third depositional sequence corresponds to a transgressive hemicycle that includes the Sables et marnes à gypses and the Molasse grise de Lausanne (lacustrine). It ranges from very late Chattian to early Aquitanian age.

#### 4.2.2. Dévoluy Domain

The Dévoluy Domain is located at the junction of the Sub-Alpine chains (Vercors and Trièves Valley) to the west, the Pelvoux crystalline massif to the north and the east, and the Diois-Baronnies chains to the south (Fig. 1). It forms a syncline surrounded by high-relief Mesozoic sedimentary rocks. Paleogene deposits constitute a 30 km long N-S band between St-Disdier to the north and Châtillon-le-Désert to the south (Fig. S12). The Dévoluy Domain has a polyphase tectonic history that comprises three main events: (1) E-W/NNW-SSE folding during Cretaceous and Eocene times related to the Pyrenean-Provençal orogeny, (2) Late Eocene and Oligocene W-E extensive stress with *syn*-tectonic sedimentation and (3) Miocene-Pliocene last phase of folding related to Alpine compression (Glangeaud and d'Albissin, 1958; Mercier, 1958; Debelmas and Lemoine, 1970; Gidon et al., 1971; Gidon and Pairis, 1976; Gidon et al., 1980; Fabre et al., 1986; Meckel et al., 1996; Meckel, 1997). The N-S Median Dévoluy Thrust that results from the westward inversion of Eocene-Oligocene normal faults during the Neogene cuts the whole domain in its central part (Fig. S12) (Gidon and Pairis, 1976; Gidon et al., 1980; Fabre et al., 1986; Meckel et al., 1996; Meckel, 1997). It corresponds to the northern extension of the Digne Thrust but records only a small displacement of a few tens of meters (Gidon and Pairis, 1976; Gidon et al., 1980; Fabre et al., 1986). The Dévoluy Domain is of particular interest for the comprehension of the WAFB development because it is one of only two domains (with the Barrême Domain) to record the transition from Flysch to Molasse Basin. In addition to literature data, we use new results from fieldwork (focused on the Souloise Group) to describe the Paleogene formations of the Dévoluy Domain.

According to Meckel (1997), these formations form two main groups. The first one, the carbonate-dominated Queyras Group, is mainly preserved in the north-eastern part of the domain and groups three formations. It starts with the (1) Conglomérats de Pierroux (30–150 m thick; Fig. 9) which consist of alluvial fan to fan delta deposits with local sedimentary material (Mesozoic carbonates clasts) flowing to the northeast (Meckel et al., 1996; Meckel, 1997). Then follows the transgressive (2) Calcaires nummulitiques (20–100 m thick; Fig. 9) deposited on a shallow carbonate ramp, and finally the (3) Marnes du Queyras (20–50 m thick; Fig. 9) deposited in a deep offshore environment (Meckel et al., 1996; Meckel, 1997). The Queyras Group was classically considered of late Priabonian in age based on faunal assemblages (Fabre et al., 1986; Pairis, 1988; Meckel, 1997), but our biostratigraphic revision places the Queyras Group within the early Rupelian (SBZ21 (Calcaires nummulitiques); NP22 (Marnes du Queyras)) (Figs. 9 and S3a). The second group is named the Souloise Group and is composed of four dominantly siliciclastic formations. It starts with the Grauwackes de la Souloise (70 m thick; Fig. 9) that consists of sandy turbidites restricted to the northwestern part of the domain and showing a coarsening upward trend above the deep offshore Marnes du Queyras (Meckel et al., 1996; Meckel, 1997). The Grauwackes de la Souloise progressively evolved to sandy bodies named the Arénites de St-Disdier (100 m thick; Fig. 9). New fieldwork reveals that the Arénites de St-Disdier consist of shallow marine deposits showing a regressive trend from lower shoreface to median estuary in a mixed wave-tidal depositional environment (Figs. S13a and S13b). The following Silts de St-Disdier (up to 100 m thick; Fig. 9) mark the transition to a continental sedimentation in the whole domain. This formation is either transitional with the underlying shallow marine deposits in the northern part of the domain or directly unconformable on the Mesozoic basement to the south (Meckel et al., 1996; Meckel, 1997; Grosjean et al., 2017). New fieldwork revealed that



**Fig. 10.** Continental sedimentary facies of the Faucon-du-Caire and Sisteron domains. (A) The Molasse bréchiqque (Faucon-du-Caire Domain) showing dominant red marls of alluvial plain and isolated conglomeratic channels with local sedimentary material. (B) The Molasse verte conglomératique (Faucon-du-Caire Domain) forming amalgamated channels and eroding the underlying sandy deposits of the Molasse rouge sableuse. (C) Drone-shot panorama of the Molasse rouge sableuse (Faucon-du-Caire Domain) showing plurimetric meandering channels (more or less amalgamated) alternating with red marls and crevasse splays of alluvial plain. Note the significant erosion at the base of some channels (white arrow). (D) The Formation grise (St-Geniez sub-domain) showing lacustrine grey marls with an isolated carbonate bed. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

it shows lateral facies variations on an N-S transect: meandering channels and alluvial plain to the north (St-Disdier area), meandering and braided channels in the central part (Montmaur area) and pure braided channels in the south (Châtillon-le-Désert area) (Figs. S12, S13c, S13d and S14). In both central and southern parts of the domain, the Silts de St-Disdier deposits progressively evolved to massive conglomeratic bodies named the Conglomérats de Montmaur (up to 200 m thick) (Fig. 9), alternating debris flows and clast-supported channels of prograding alluvial fans (Figs. S13c and S13d). The Conglomérats de Montmaur are thrust in their upper part by the Digne Thrust. Although previous works described various palaeodirections (Meckel, 1997; Grosjean et al., 2017), our new results clearly indicate that the entire Souloise Group flowed WNW-N based on trough cross-bedding and flute cast directions (Fig. S12 and S13d). Sedimentary material contains local Mesozoic clasts and allochthonous components coming from the Internal Alpine Nappes (Gidon, 1971; Ivaldi, 1989; Meckel et al., 1996; Meckel, 1997; Cordey et al., 2012; Jourdan et al., 2013). The Souloise Group is poorly dated. The Silts de St-Disdier were initially considered to be of Chattian age based on the presence of charophyte species (*Sphaerochara hirmeri*, *Rhabdochara kraeuseli*), which nowadays belong to biozones OC2-MC3 of Rupelian to Langhian age (Mojon et al., 2018). New palynological analyses were attempted in this study but the selected samples proved to be barren after analysis (Table S1).

The formations of the Dévoluy Domain form one depositional sequence (Fig. 9). The TST and HST correspond respectively to the Queyras (alluvial fan to deep marine environments) and Souloise groups (deep marine to alluvial fan environments). We place the MFS within the Marnes du Queyras.

#### 4.2.3. Faucon-du-Caire Domain

The Faucon-du-Caire Domain forms a SW-NE oriented band of Cenozoic formations located 25 km south of the Dévoluy Domain

(Figs. 1 and 2a). Partially covered up by the Digne Thrust in the north, it has been deformed by both salt and Alpine tectonics during Cenozoic times (e.g., Célini et al., 2021; Hamon, 2022). The Paleogene formations of the Faucon-du-Caire Domain have been poorly documented and consequently we use literature data but mainly new fieldwork to describe these formations.

Sedimentation started with the Molasse bréchiqque (50 m thick; Figs. 9, 10a and S15a), which overlies unconformably either Mesozoic terrains or scattered pockets of Eocene Calcaires nummulitiques. The Molasse bréchiqque consists of brecciated bars of alluvial fan deposits showing south-east to east palaeoflow directions (Hamon, 2022). Sedimentary material is only of local origin (Mesozoic carbonate clasts; Gidon et al., 1991; Hamon, 2022). New fieldwork shows that alluvial fan deposits retrograde to colourful marls and palaeosoils of alluvial plain environments named the Marnes colorées (up to 200 m thick; Figs. 9 and S15a). These first two formations are onlapping to the east on the Mesozoic terrains, which confirms the retrograding trend of the system (Fig. S15b). Fine-grained deposits evolved to a fluvial formation named the Molasse rouge sableuse (100 m thick; Figs. 9, 10b, c and S15c). It consists of isolated sandy bodies at its base and alternating sandy to conglomeratic channels at its top, indicating a progradation with western palaeoflow based on flute casts direction. Sedimentary material contains radiolarite and green schist components from the Internal Alpine Nappes. Tremblay (2002) has documented the upper part of the Faucon-du-Caire Domain formations. It consists of the Molasse verte conglomératique (up to 200 m thick; Figs. 9 and 10b), which rests unconformably either on the underlying Oligocene units or the Calcaires nummulitiques (Gidon et al., 1991). It begins with massive braided conglomeratic channels that evolved to meandering channels and red marls of alluvial plain environments that display hook halokinetic sequences relative to the active salt-tectonics of the Triassic Astoin diapir during deposition of Cenozoic sediments (Célini et al., 2021). The Digne

Thrust covers this last formation. No biostratigraphic data for constraining the stratigraphic age of the Cenozoic Faucon-du-Caire formations are available. It is the only domain of the entire WAFB that remains completely undated.

We divide the Faucon-du-Caire Domain formations into two depositional sequences (Fig. 9). The first one groups the Molasse bréchiq ue and the Marnes color ees as the TST (alluvial fan to alluvial plain environments including the MFS) and the Molasse rouge sableuse as the HST (fluvial environment). The Molasse verte conglom eratique forms the second depositional sequence and corresponds to a transgressive hemicycle.

#### 4.2.4. Esclangon-Esparron Domain

The Esclangon and Esparron domains are only ten kilometres apart and both are located at the front of the Digne Thrust (Figs. 1 and 2a). Their Cenozoic formations have about the same sedimentological and stratigraphical characteristics and are therefore described together in this study. The Esparron Cenozoic series forms a 6 km long and 300 m thick NW-SE striking vertical cliff, whereas the Esclangon formations are preserved in a small area of triangular shape underlying either the Aquitanian marine molasse or the Digne Thrust (Cou eff e and Maridet, 2003) (Fig. 2a). The Esclangon and Esparron domains have been heavily documented, and the following descriptions are coming from numerous previous studies.

Sedimentation started with the Molasse rouge br eche, which overlies unconformably either the Mesozoic formations or the Calcaires nummulitiques (Haccard et al., 1989) (Fig. 9). It consists mainly of conglomerates of alluvial fan deposits showing gravity facies (mud and debris flows) at the base of the system, and channelized deposits with alluvial plain deposits at the top, indicating retrogradation (Haccard et al., 1989; Tremblay, 2002; Hamon, 2022). Sedimentary material is only of local origin and consists of Mesozoic carbonate clasts that can reach a few meters in width (Haccard et al., 1989). The whole system flows to the east, filling local depressions in an irregular palaeotopography, which explains the strong heterogeneity of the formation in terms of thickness variations (0 to 120 m) (Rousset et al., 1983; Haccard et al., 1989; Tremblay, 2002; Hamon, 2022). Alluvial fan deposits progressively evolved to lacustrine carbonate beds and red to grey marls that compose the Molasse lacustre inf erieure (20 m thick; Fig. 9). The overlying Molasse rouge sableuse constitutes the main part of the Esclangon and Esparron domains (200 to 350 m thick; Fig. 9). Transitional with the underlying lacustrine deposits, it consists of fluvial facies showing red marls of alluvial plain deposits with palaeosoils and meandering channels showing a westward palaeoflow direction (Rousset et al., 1983; Haccard et al., 1989; Bialkowski, 2002; Tremblay, 2002). Sedimentary material is both local and allochthonous: quartzite, serpentinite, radiolarite and green schist clasts indicate a source located in the Internal Alpine Nappes (Rousset et al., 1983; Haccard et al., 1989; Tremblay, 2002). Although Tremblay (2002) described a retrogradational trend for the Molasse rouge sableuse, Bialkowski (2002) and Held (2011) proposed a more complete description of both the Esparron and Esclangon domains, clearly indicating a prograding fluvial formation with dominant alluvial plain deposits at the base of the section and amalgamated meandering channels at the top. The Molasse rouge sableuse gradually evolved in its upper part to grey marls and carbonates beds of lacustrine environment that compose the Molasse grise (30 to 100 m thick; Fig. 9). The Esparron and Esclangon sections remain poorly dated. Tremblay (2002) described charophytes assemblages in the Molasse rouge br eche ranging from late Rupelian to early Chattian (MP23 to MP26 equivalent), but no details on the species assemblages are provided. Concerning the Esclangon area, Bialkowski assigns an age ranging between Late Eocene and middle Rupelian to the entire section basing on  $\delta^{13}\text{C}$  record in palaeosoils. The most reliable dating elements are coming from the discovery of several mammal remains in the Molasse lacustre inf erieure (*Mesaceraotherium gaimersheimense* (Antoine et al., 2022), MP28–30 (Tissier et al., 2020)) and in the middle part of

the Molasse rouge sableuse (*Issiodoromys limognensis* (Tortosa, 2005), MP28 (Bolliger, 2013)), which constrain the age of these formations to the late Chattian (Fig. 9).

We divide the Paleogene formations of the Esclangon-Esparron Domain into two depositional sequences. The first one comprises the Molasse rouge br eche and the Molasse lacustre inf erieure as the TST (Alluvial fan to lacustrine (MFS)) and the Molasse rouge sableuse as the HST (fluvial). The Molasse grise forms the second one and corresponds to a transgressive hemicycle.

#### 4.2.5. Sisteron Domain

The Sisteron Domain is located between the Durance and the Esclangon-Esparron domains (Figs. 1 and 2a). It forms a 8 km long NNE-SSW oriented and 1 km wide band of Paleogene formations that can be divided into two sub-domains: (1) the Maurel sub-domain to the south, which constitutes the basement of the Miocene-Pliocene Digne-Valensole Plateau, and (2) the St-Geniez sub-domain to the north, which forms a small cliff trapped below the Digne Thrust (Fig. 2a). Here we use previous studies and new fieldwork (mainly focused on the Maurel sub-domain) to document this domain.

Sedimentation begins with the Formation conglom eratique inf erieure (up to 900 m thick, Fig. 8b) which is restricted to the Maurel sub-domain. It consists of conglomeratic bodies alternating with red to brown marls. New fieldwork shows that gravity deposits (mud and debris flow) and massive clasts dominate in the basal part of the formation, whereas clast-dominated channels and fine-grained sediments are more important at the top (Figs. S16 and S17a). The whole formation is therefore a retrograding alluvial fan with northward dominant palaeoflow directions and pure locally derived sedimentary material (Upper Cretaceous clasts). Gigot et al. (in prep) noticed that conglomeratic bodies change toward the north to grey lacustrine marls that could correspond to the distal part of the system. The whole Formation conglom eratique inf erieure is poorly dated. Charophytes assemblages found at its base would however date the beginning of sedimentation between middle Bartonian and early Priabonian times (*Chara friteli*, *Harrisichara lineata*; Mart ın-Closas et al., 1999; Gigot et al., in prep; Mojon et al., 2018) (Fig. 8b). The second formation, named the Formation conglom eratique sup erieure (Fig. 8b), shows significant thickness variations and is present in both the Maurel (100 m thick) and St-Geniez sub-domains (10 m thick). We did not find its base in the field, but Tremblay (2002) described it as unconformably overlying the Formation conglom eratique inf erieure. The Formation conglom eratique sup erieure consists dominantly of orange marls with palaeosoils of alluvial plain environments and small isolated fluvial channels composed of clast-supported conglomerates showing south to west palaeoflow directions. Sedimentary material derived from local Mesozoic rocks. New fieldwork shows that the conglomeratic channels are restricted to the lower part of the Formation conglom eratique sup erieure, whereas alluvial plain deposits and numerous palaeosoils are dominant in the upper part, indicating a retrogradation of the system (Figs. 8b, S16, S17b and S17c). Fluvial deposits then progressively evolved to lacustrine grey marls and carbonates or gypsum beds belonging to the Formation grise (Figs. 8 and 10d). This formation shows thickness variations between the Maurel (50 m thick) and St-Geniez (110 m thick) sub-domains. According to our revised biostratigraphy, established from the work of Gigot et al. (in prep), a late Rupelian age (MP24–25; Figs. 8b and S3b) is proposed to the lower part of the Formation grise based on mammal assemblages, and a late Chattian age (OC5; Figs. 8b and S3b) to its upper part based on charophytes assemblages. The last deposits of the Sisteron Domain consist of multicolour marls with calcareous palaeosoils and meandering sandy channels belonging to the Molasse rouge (up to 100 m thick; Fig. 8b). Documented by Gigot et al. (in prep) in the Maurel sub-domain, we did not find this last formation during our fieldwork. In the St-Geniez sub-domain, the Molasse rouge is transitional with the underlying lacustrine deposits and is finally covered up by the Digne Thrust. Sandy fluvial facies show southwest palaeoflow directions

and contain numerous serpentine grains (up to 50 % of the detrital fraction) coming from the Internal Alpine Nappes (Tremblay, 2002; Gigot et al., in prep; Hamon, 2022). This last formation is poorly dated.

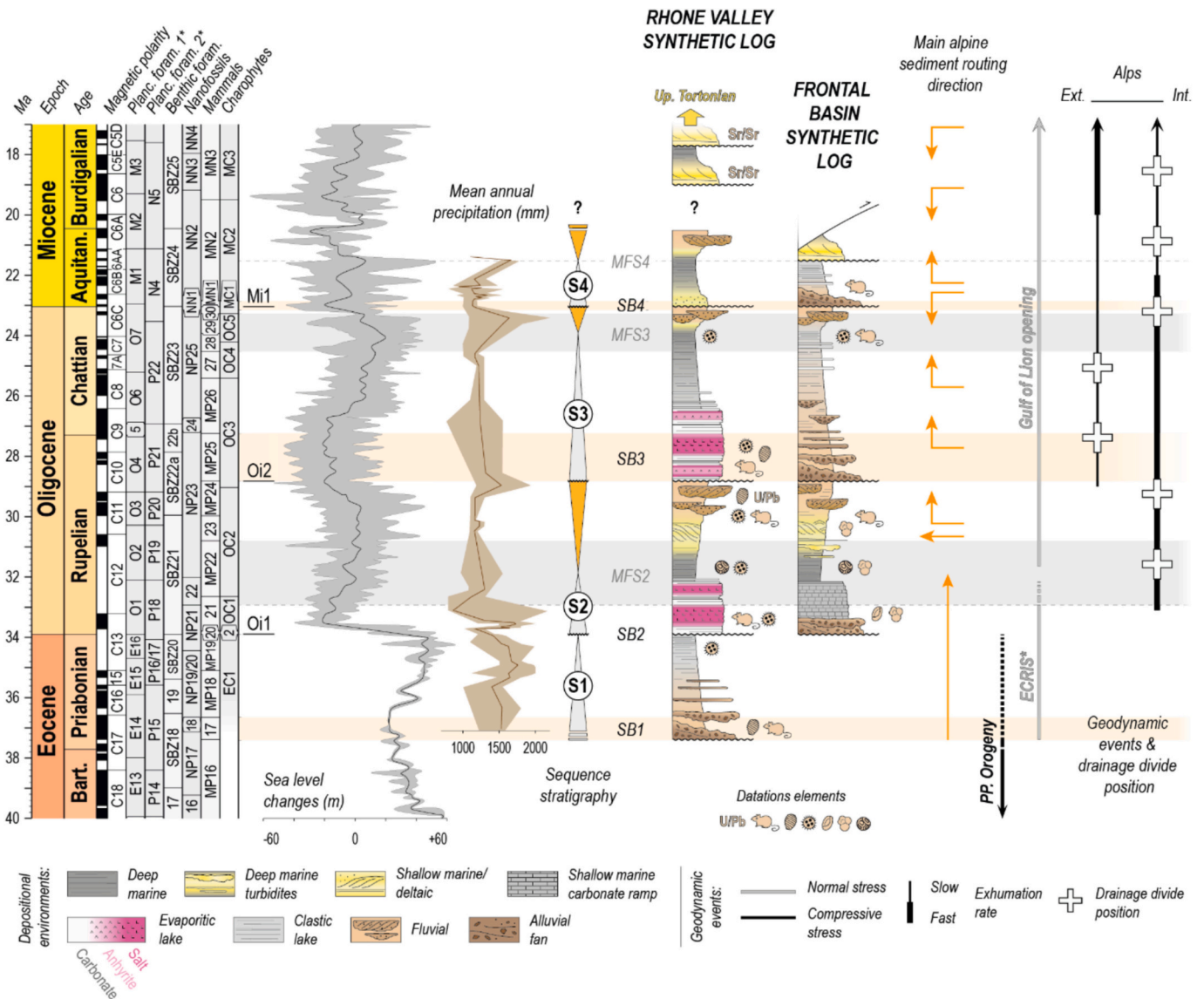
We divide the Paleogene formations of the Sisteron Domain into two depositional sequences (Fig. 8b). The Formation conglomeratique inférieure forms the first sequence and corresponds to a transgressive hemicycle (retrograding alluvial fan deposits). Its base is dated from middle Bartonian to early Priabonian, but the age of the whole depositional sequence is not known. The second depositional sequence groups the Formation conglomeratique supérieure and the Formation grise (fluvial to lacustrine (MFS)) as the TST, and the Molasse rouge (fluvial) as the HST. It ranges from late Rupelian to late Chattian in age.

#### 4.2.6. Barrême Domain

The Barrême Domain is the southernmost of the Frontal Basin. It is located within the Sub-Alpine chains of the Castellane Arc, 20 km south of the Esclançon-Esparron areas and 12 km east of the Miocene-Pliocene Digne-Valensole Plateau (Fig. 1). The Cenozoic formations of the Barrême Domain are the only one of all frontal molasse domains to be preserved above the Digne Thrust, as a transported basin, and form a narrow (3 km wide at most) 20 km long NNW-SSW oriented syncline. Paleogene sedimentation has been associated with relatively continuous Alpine tectonics, which led to the formation of growth folds and syntectonic unconformities because of progressive transport on the Digne Thrust (de Graciansky et al., 1982; Artoni and Meckel, 1998; Evans and Elliott, 1999; Callec, 2001). All following descriptions are a synthesis of previous work.

Sedimentation started with clast-supported conglomeratic beds, named the Poudingues d'Argens (locally up to 600 m thick, but highly variable; Fig. 9), which overlies unconformably Mesozoic formations above a weathering horizon (de Graciansky et al., 1982). The Poudingues d'Argens consists of alluvial fan deposits restricted in the central part of the syncline and filling local depressions in an irregular palaeotopography with northward progradation directions (de Graciansky et al., 1982; Evans and Elliott, 1999; Callec, 2001; Grosjean et al., 2017). Sedimentary material consists of local Mesozoic carbonates clasts (de Graciansky et al., 1982; Evans and Elliott, 1999). Because of a lack of biostratigraphic data, the Poudingues d'Argens are still undated. The alluvial fan deposits are vertically and laterally overlain by the Calcaires nummulitiques (10 m thick; Fig. 9). They correspond to carbonate ramp deposits showing a typical fauna of a shallow marine environment (Bodelle, 1971; de Graciansky et al., 1982; Evans and Elliott, 1999; Callec, 2001). The Calcaires nummulitiques were until now considered to be of Priabonian age (Bodelle, 1971; de Graciansky et al., 1982; Callec, 2001), but our last biostratigraphic revision move them to early Rupelian (O1; Figs. 9 and S3d). The carbonate facies progressively evolved into the Marnes bleues (up to 500 m thick; Fig. 9) of deep-offshore environment, following the transgressive trend initiated by the previous units (de Graciansky et al., 1982; Evans and Elliott, 1999; Callec, 2001). According to Callec, 2001, the Marnes bleues extended from the early to middle Rupelian (P18-P19/O1-O2 zones), but we restrict it to the lower part of the P19/O2 zones of the middle Rupelian (Figs. 9 and S3d). The Marnes bleues were progressively enriched upward with coarse deposits that constitute the Grès de Ville (55 m thick; Fig. 9). This formation shows numerous turbidites, hummocky cross stratifications and wave ripples that characterize a shallow marine platform varying between offshore transition and shoreface environment (Evans and Elliott, 1999; Callec, 2001). The sandy structures show a constant N-NW palaeoflow direction (Evans and Elliott, 1999; Callec, 2001) and sedimentary material indicates two main sources: (1) the southerly-located Pyrenean-Provençal-Corso-Sardinian massif and (2) andesitic volcanoclastic debris coming from an unknown location because of the absence of preserved Cenozoic volcanic formations in SE France (Evans and Mange-Rajetzky, 1991). The Grès de Ville were supposed to range from P19/O2 to P20/O3 biozones but following our biostratigraphic revision we restrict them to P20/O3 biozones of middle

to late Rupelian age (Figs. 9 and S3e). The Grès de Ville are unconformably overlain by conglomerate and sandstone units that are classically subdivided into three members of lateral equivalents, the Conglomérats de Clumanc, the Conglomérats de St-Lions and the Grès de Sénéz. The Conglomérats de St-Lions and Clumanc are located in the northern and central parts of the Barrême Domain and consist of conglomeratic formations of fan-delta deposits showing lateral facies variations between prodelta, delta front and delta plain environments (Callec, 2001; Grosjean et al., 2017) (Fig. 9). Intercalated marine sandstone bodies comprise wave and tidal structures, which attest of a mixed-influenced shallow marine environment (Callec, 2001). The flow direction is mainly westward (Bodelle, 1971; Evans and Elliott, 1999; Callec, 2001) but a gradual southbound reorientation occurs in the upper part of the Conglomérats de St-Lions (Evans and Elliott, 1999; Callec, 2001). In its upper part, the Conglomérats de St-Lions comprises an internal unconformity that separates underlying deltaic conglomeratic deposits from a shallow marine coral biostrome (Biostrome de St-Lions; Fig. 9). To the south of the domain, conglomeratic formations evolve laterally to the Grès de Sénéz (Fig. 9). They consist of coarsening upward sandy deposits prograding to the NW and showing typical storm wave structures of shoreface environments (Evans and Elliott, 1999; Callec, 2001). Although the Grès de Sénéz were mainly sourced by the Pyrenean-Provençal-Corso-Sardinian massif (Evans and Mange-Rajetzky, 1991), petrological (Chauveau and Lemoine, 1961; de Graciansky et al., 1971, 1982; Evans and Mange-Rajetzky, 1991; Cordey et al., 2012; Schwartz et al., 2012; Jourdan et al., 2012), geochronological (Morag et al., 2008) and thermochronological (Bernet et al., 2009; Jourdan et al., 2013) analyses attest for Internal Alpine Nappes inputs in the Conglomérats de Clumanc and Conglomérats de St-Lions. This constitutes the first Alpine provenance in the Barrême sedimentation. These formations were supposed to be of middle to late Rupelian age based on faunal assemblages (P20/O3; Callec, 2001). Our biostratigraphic revision restricts all three members to the O4 Zone of late Rupelian age (Figs. 9 and S3e). The overlying formation is named the Molasse rouge (150 m thick; Fig. 9) and marks the transition to fully continental sedimentation in the Barrême Domain. Callec (2001) described it as transitional with the underlying Conglomérats de St-Lions. In contrast, Gubler (1958), Chauveau and Lemoine (1961), Bodelle (1971), de Graciansky et al. (1982), and Evans and Elliott (1999) asserted that a strong angular discontinuity separates the two formations, which would explain why the Molasse rouge overlies locally either the Poudingues d'Argens, the Calcaires nummulitiques, the Marnes bleues, or the conglomeratic members of St-Lions and Clumanc. The Molasse rouge consists mainly of red marls of alluvial plain deposits with recurrent calcareous palaeosols, isolated micritic limestones of ephemeral lakes, and channel-shaped or sheet-like conglomeratic bodies of braided river (de Graciansky et al., 1982; Evans and Elliott, 1999). These facies characterize a retrograding alluvial fan environment with westward to south-westward palaeoflow directions (de Graciansky et al., 1982; Evans and Elliott, 1999). Sedimentary material was mainly of local origin (Upper Cretaceous clasts) but contains limited allochthonous components either reworked from underlying units (de Graciansky et al., 1982; Evans and Mange-Rajetzky, 1991) or derived from the crystalline basement of the western Alps (Bernet et al., 2009). The Molasse rouge is poorly dated. The overlying alluvial fan deposits retrograded to dominantly multicoloured marls and carbonate rock to gypsum beds of lacustrine and palustrine environments (de Graciansky et al., 1982; Evans and Elliott, 1999) named the Série saumon and Série grise (10 to 180 m thick; Fig. 9). Mammal assemblages constrain a Chattian age to the Série grise (Carbonnel et al., 1972; de Graciansky et al., 1982), but we restrict it to a late Chattian age (MP29) after our biostratigraphic revision (Figs. 9 and S3f), which is consistent with pollen data from Fauquette et al. (2015). The last and youngest formation of the Barrême Domain, the Grès verts (80 m thick; Fig. 9), is transitional with the underlying unit except in the eastern part of the area where its base is a local syn-tectonic unconformity (Evans and



**Fig. 11.** Correlation of synthetic logs of the Rhône Valley Basin and Frontal Basin with their sequential division, climato-eustatic curves, main alpine sediment routing direction and geodynamic events. Magnetic polarity, foraminifera, nannofossil and mammal biozones are from Gradstein et al. (2020). Charophyte biozone are from Mojon et al. (2018). Sea level curve is from Miller et al. (2020). Mean annual precipitation curve is from Suc et al. (2025). Tectonic features are from Séranne (1999), Séranne et al. (2021), Bellahsen et al. (in prep).

Elliott, 1999). It consists of fluvial deposits alternating red marls with palaeosoils of alluvial plain and sandy meandering channels showing dominant southward palaeoflow direction (de Graciansky et al., 1982; Evans and Elliott, 1999; Held, 2011; Grosjean et al., 2017). Petrological analyses attest of an Internal Alpine Nappes provenance for the sedimentary components (de Graciansky et al., 1982; Evans and Mange-Rajetzky, 1991; Jourdan et al., 2012; Schwartz et al., 2012). There is no robust data concerning the age of this last formation, which is classically attributed to late Chattian/early Aquitanian.

We divide the Paleogene formations of the Barrême Domain into two depositional sequences (Fig. 9). The first sequence groups the Poudingues d’Argens, the Calcaires nummulitiques and the Marnes bleues (which include MFS) as the TST (alluvial fan to deep marine environments), and the Grès de Ville, the Grès de Sénéze, the Conglomérats de Clumanc and the Conglomérats de St-Lions as the HST (offshore transition to fan-delta environments). This sequence was deposited over the entire Rupelian. The second depositional sequence groups the Molasse rouge, the Série saumon and the Série grise (MFS) as the TST (alluvial fan to lacustrine), and the Grès verts as the HST (fluvial), deposited

during the entire Chattian.

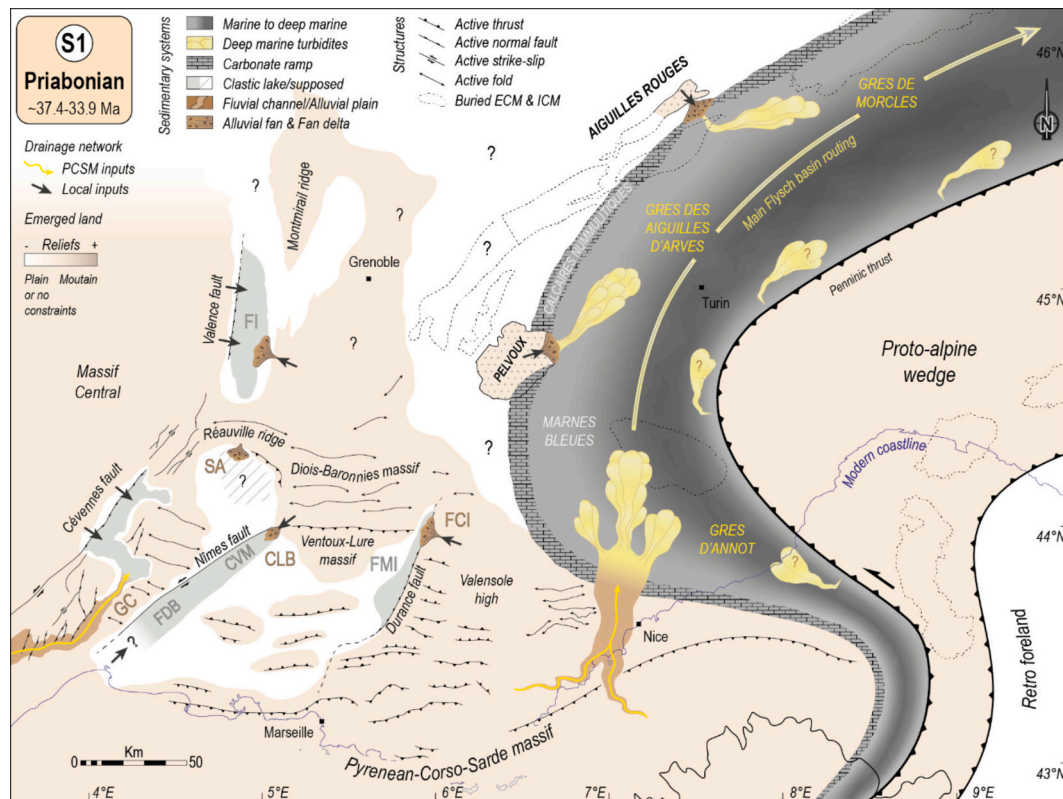
## 5. Interpretation

### 5.1. Sequence stratigraphy

Based on the chronostratigraphic framework previously established, we propose to group the Priabonian to Aquitanian formations of the WAFB into four main depositional sequences (S1 to S4) integrated into two synthetic logs for the Rhône Valley Basin and the Frontal Basin (Fig. 11).

Sequence S1 (Priabonian; ~37.4–33.9 Ma; Fig. 11) is mainly present in the Rhône Valley Basin, except for the Provence Domain. It groups the first half-sequences of the Bas-Dauphiné, Camargue, Durance and Mormoiron domains, but also the first sequence of the Sisteron Domain. Poorly dated as a whole, deposition of sequence S1 started during biozone MP17 (~37.4–36.65 Ma) and probably continued until the end of the Priabonian.

Sequence S2 (early to early late Rupelian; ~33.9–28.8 Ma; Fig. 11) is



**Fig. 12.** Palaeogeographic and source-to-sink map of the WAFB during Priabonian (S1 sequence). Languedoc, Alès and Nîmes palaeogeography is compiled from Lettéron et al. (2017, 2018, 2022), Séranne et al. (2021) and Semmani et al. (2023). Pyrenean-Provence structures are from Balansa et al. (2022) and references therein. Flysch Basin palaeogeography is compiled from Sinclair (1997) and Joseph et al. (2005). Corso-Sardinian palaeogeography is from Romagny et al. (2020). The above references also apply to Figs. 13 to 19. The reliefs of the Massif Central and the Pyrenean-Provençal massif have not been suggested due to lack of paleotopographic data (also applied to Figs. 13 to 19). Abbreviations: CLB=Conglomerats de Crillon-le-Brave; CVM = Complexe argileux vert de Mormoiron; ECM = External Crystalline Massifs (also applied to Figs. 13 to 19); FCI=Formation marneuse infrasalifère; FDB=Formation détritique basale; FI=Formation infrasalifère; FMI=Formation marneuse infrasalifère; GC = Grès de Célas; ICM = Internal Crystalline Massifs (also applied to Figs. 13 to 19); PSCM = Pyrenean-Provençal-Corso-Sardinian Massif (also applied to Figs. 13 to 19); SA = Sables d'Aleyrac.

present in most domains, except for the Esparron-Esclangon, Faucon-du-Caire and Sisteron domains, where it was most likely eroded. It constitutes the earliest Paleogene deposits in the Barrême, Dévoluy, Savoie and Provence domains. Although the formations of the Dévoluy Domain lack biostratigraphic constraints, we associate them with sequence S2 because of stratigraphic similarities with the Savoie and Barrême domains. Even though most of the transgressive formations mark the early Rupelian, the base of the Complexe argileux blanc de Blauvac of the Mormoiron Domain yielded pollen assemblages indicating the very end of the Priabonian. We therefore assume that S2 started at the Priabonian/Rupelian transition at 33.9 Ma. We restrict the MFS between the base of biozone NP22 and the end of biozone MP22 (~32.9–30.9 Ma), i. e. between the late early and early middle Rupelian. The Montmirail Domain records an additional Rupelian sequence that we name S2a (Figs. 5b and S8). It occurs over a short period within the early Rupelian. Considering its important thickness, we relate its deposition to salt tectonics of the surrounding Suzette diapir that produced local subsidence.

Sequence S3 (latest Rupelian to late Chattian; ~28.8–23 Ma; Fig. 11) is present in all domains, except for the Dévoluy and Montmirail domains where recent erosion is exposing sequence S2. Sequence S3 is partially preserved in the Durance and Mormoiron domains where the HST has been eroded during the Burdigalian transgression. Despite its complete lack of biostratigraphic age control, we associate the first depositional sequence of the Faucon-du-Caire Domain to S3 because of its lithological and stratigraphical similarities with the other surrounding Frontal Basin domains. The base of sequence S3 is very well dated in the Rhône Valley Basin but not in the Frontal Basin where no

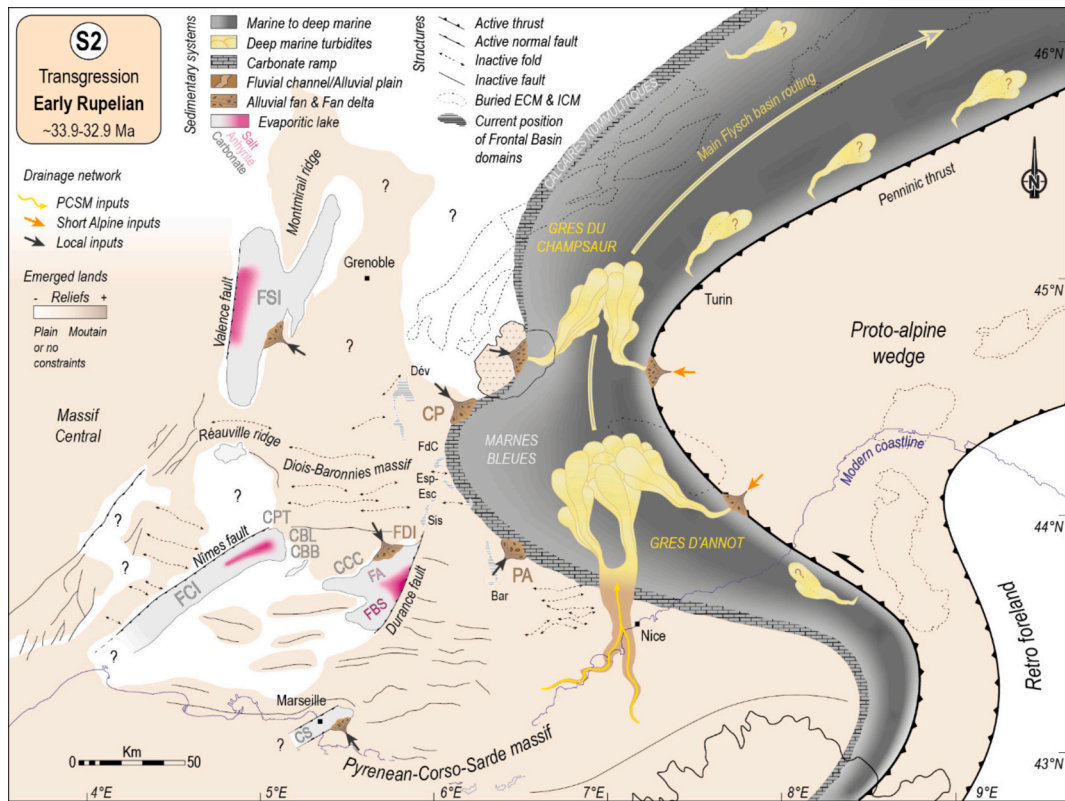
biostratigraphic data are available. In accordance with mammal assemblages of the Durance and Mormoiron domains, we place the base of S3 within biozone MP25 (~28.8–27.2 Ma). The MFS can be positioned between the base of biozone MP28 and the end of biozone MP29 (~24.55–23.25 Ma), during the late Chattian.

Sequence S4 (Aquitanian; ~23–? Ma; Fig. 11) is the least preserved. It was partly eroded by Quaternary glaciations in the Frontal Basin (Barrême, Dévoluy and Sisteron domains) and by the Miocene transgression in the Rhône Valley Basin (Durance, Montmirail and Mormoiron domains). Except in the Savoie Domain, S4 has not provided biostratigraphic data. From the Savoie Domain data, we assume that deposition of S4 started between biozone MP30b and biozone MN1 (~23.15–22.85 Ma), around the Chattian-Aquitanian transition. Although its complete duration is not known, we consider that its MFS is prior to the late Aquitanian transgression of the Miocene Marine Molasse Cycle in the Savoie Domain (~21 Ma; Kalifi et al., 2021, 2022).

## 5.2. Palaeogeographic and source-to-sink evolution of the WAFB between Priabonian and Aquitanian times

### 5.2.1. S1 – Priabonian – ~37.4–33.9 Ma

During the Priabonian (Fig. 12), the arc-shaped Flysch Basin developed along the proto-Alpine chain. Sedimentation took place in a 100–120 km width trench (Ford and Lickorish, 2004) and is classically described as the Nummulitic trilogy (Boussac, 1912; Bodelle, 1971; Campredon, 1972) (Fig. 12). Turbidite deposits developed in a major S–N routing system along the foredeep (Ravenne et al., 1987; Sinclair, 1997) (Fig. 12). The Pyrenean-Provençal-Corso-Sardinian massif is the



**Fig. 13.** Palaeogeographic and source-to-sink map of the WAFB during early Rupelian (S2 sequence - TST). Abbreviations: CBB=Complexe argileux blanc de Blauvac; CBL = Calcaires à Brotia lauraea; CCC=Calcaires de Campagne-Cavalon; CP=Conglomerats de Pierroux; CPT = Calcaires du Pâtis; FA = Formation anhydritique; FBS=Formation basale salifère; FCI=Formation calcaire basale; FDI=Formation détritique inférieure; FSI=Formation salifère inférieure; PA = Poudingues d'Argens.

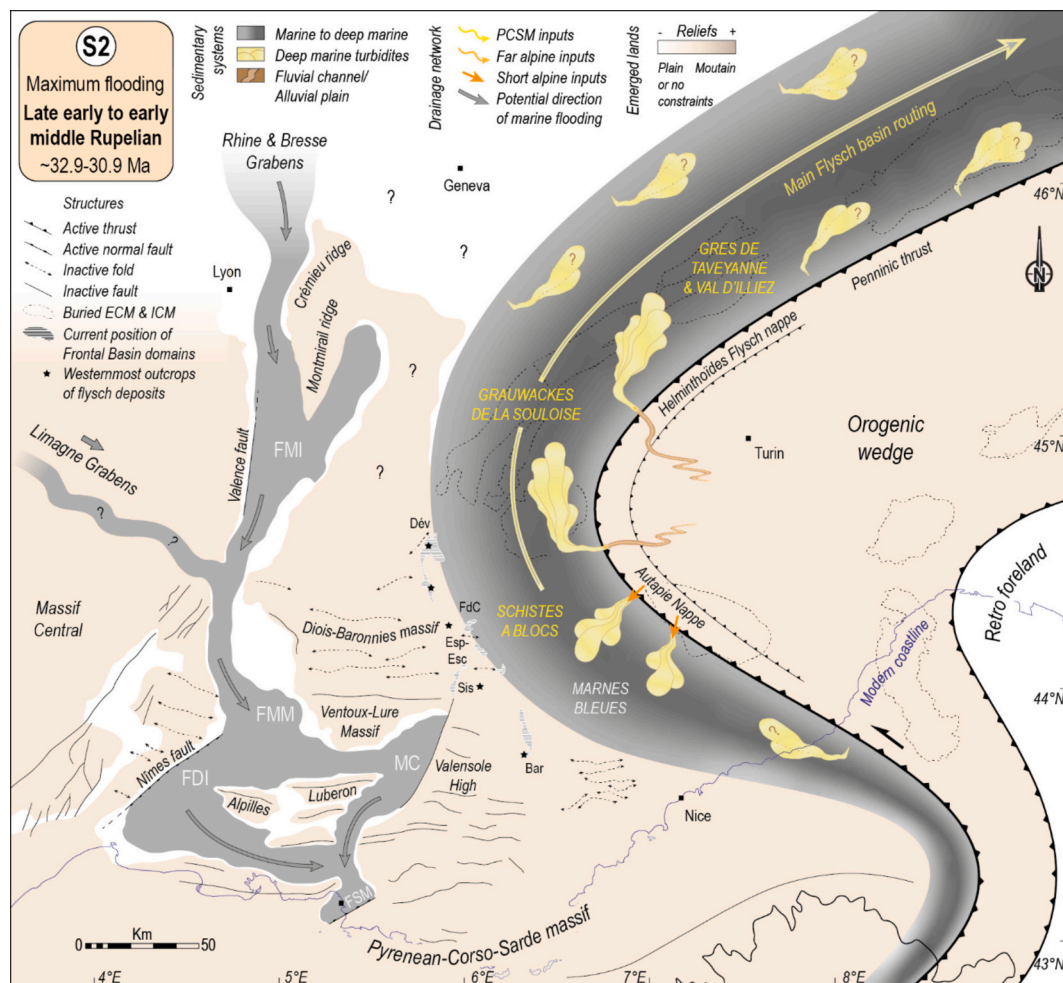
main source for the Grès d'Annot, whereas the Pelvoux and Aiguilles Rouges palaeohighs provided sedimentary material to the Grès des Aiguilles d'Arves and Grès de Morcles, respectively (Jean et al., 1985; Ivaldi, 1989; Mayoraz, 1995; Ford, 1996; Burkhard and Sommaruga, 1998; Joseph and Lomas, 2004).

To the west, in the Rhône Valley, S1 consists of retrograding clastic lake with locally basal alluvial fan deposits. Sedimentation is synchronous with the final stage of the Pyrenean-Provençal orogeny (Séranne et al., 2021) and most of the sedimentary material was derived from the dismantling of the previously uplifted topography of the Mesozoic terrains. The eastern border of the Massif Central, which was exhumed at least since the Lower Cretaceous (e.g., Barbarand et al., 2001), may also have supplied the Bas-Dauphiné Domain with crystalline basement or Mesozoic cover material (Latreille, 1969). Most of the Rhône Valley Basin evolved as separated domains: there is no evidence of connections between the Bas-Dauphiné Domain and the middle part of the Rhône Valley (Valréas Domain) suggested by onlaps of the Formation infrasalifère on the Réauville ridge to the south and the east of the domain. We consequently interpret this Réauville ridge as a Pyrenean-Provençal structure as it folds Upper Cretaceous terrains and predates Priabonian sedimentation. It corresponds to one of the northernmost structures of Pyrenean-Provençal origin in SE France, with the ones preserved in the Pelvoux massif and Dévoluy Domain (Gidon et al., 1971; Gidon and Pairis, 1976; Fabre et al., 1986; Ford, 1996). We suspect that the Priabonian basin in the middle Rhône Valley was larger at this time, due to sedimentary remnants as conglomeratic bodies of alluvial fan deposits retrograding to lacustrine marls in the Monjoyer area (Sables d'Aleyrac; Figs. 1, 12; S1 and S18). To the south, and as suggested by Semmani et al. (2023), both the Mormoiron and Camargue domains could have formed a unique lacustrine system (Fig. 12). Facies distributions suggest that the main depocenter was in the Camargue area. A connection could also have existed between the Maurel and Manosque sub-domains. We

consider that the alluvial fan deposits of the Lower Conglomeratic Formation (Maurel sub-domain) as a proximal equivalent of the lacustrine Formation marneuse infrasalifère (Manosque sub-domain) (Fig. 12). As alluvial fan deposits are absent in the St-Geniez area, the border of the basin was probably located in the Maurel sub-domain. The Durance Domain was probably isolated from the Camargue Domain to the west because of topographic growth caused by Pyrenean-Provençal compression (Ventoux-Lure Palaeohigh, Vaucluse plateau and Luberon and Alpilles thrusts) (e.g., Ford and Stahel, 1995; Bestani et al., 2015, 2016; Zeboudj et al., 2025) (Fig. 12). Mainly present in the Rhône Valley, S1 is absent in the Provence Domain that could have been located on a topographic high during the Priabonian.

### 5.2.2. S2 – transgression – early Rupelian – ~33.9–32.9 Ma

During early Rupelian times, the activity of the Penninic frontal thrust in the Alpine domain caused the westward migration of the Flysch Basin (~30 km) and shallow marine carbonate facies developed in the Dévoluy and Barrême areas (Fig. 13). S2 is absent in the Faucon-du-Caire and Esparron-Esclançon domains, but residual sediments of foraminifera-rich limestones documented below the Digne Thrust and the Cenozoic continental formations (Gidon et al., 1971; Rousset et al., 1983; Gidon et al., 1991) indicate that the early Rupelian Nummulitic transgression reached these areas and was eroded before molasse deposition (S3). Turbidite deposits were mainly sourced from the Pyrenean-Provençal-Corso-Sardinian massif (Grès d'Annot) and the Pelvoux massif (Grès du Champsaur) (Jean et al., 1985; Joseph and Lomas, 2004). However, the occurrence of andesite clasts related to active volcanism in the orogen highlights the very first Alpine inputs in the Frontal Basin (32.5 ± 2.2 Ma - Ar-Ar dating on amphiboles in the Grès du Champsaur (Féraud et al., 1995); 33.7. ± 1.6 Ma, 32.4 ± 1.6 Ma, 30.5 ± 3 Ma - K/Ar dating in the Grès d'Annot (Ginsburg et al., 1980; Baudron and Cavelier, 1982).



**Fig. 14.** Palaeogeographic and source-to-sink map of the WAFB during late early to early middle Rupelian (S2 sequence - MFS). Abbreviations: FDI=Formation détritico-intermédiaire; FMI=Formation marneuse intermédiaire; FSM = Formation de St-Marcel; FMM = Formation marine de Montmirail (Cycle II); MC = Marnes de Caseneuve.

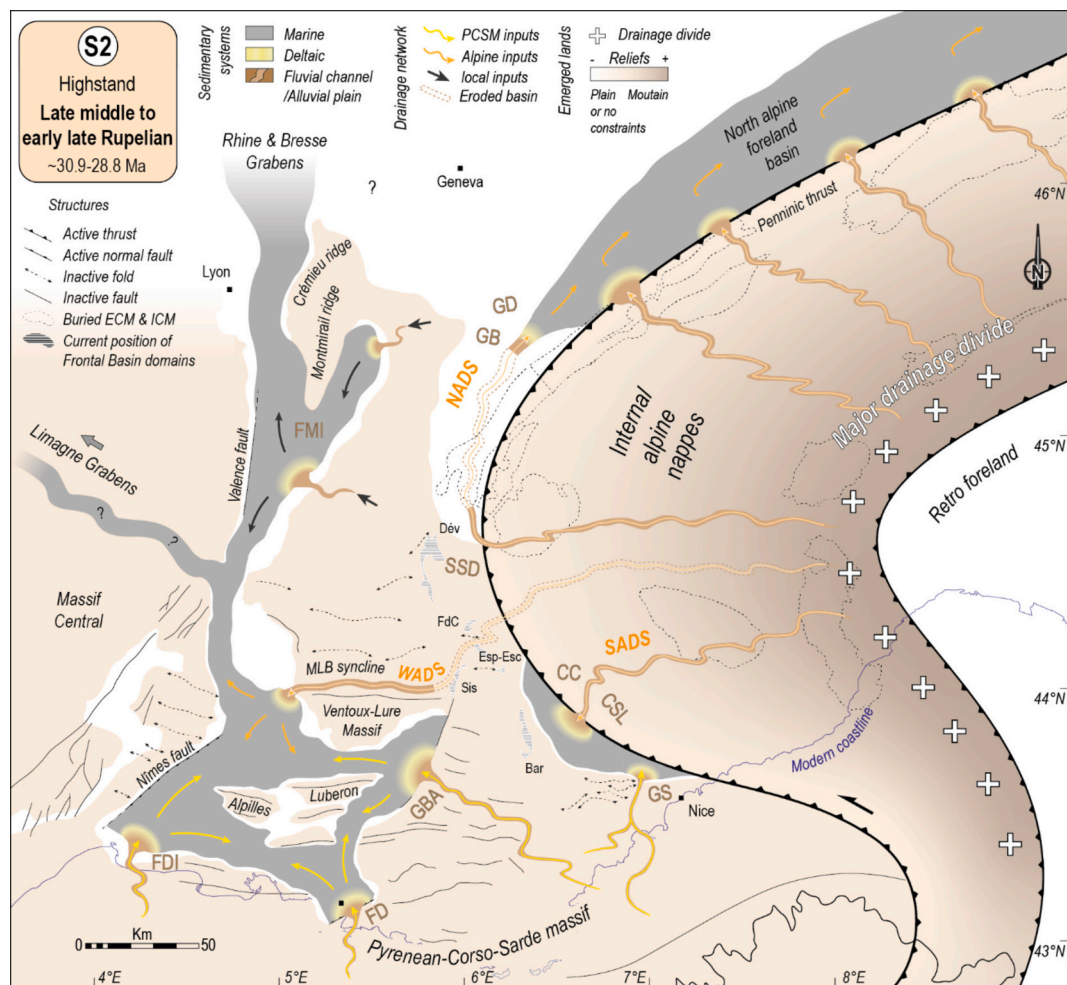
To the west, in the Rhône Valley Basin, the Priabonian/Rupelian transition corresponds to a change in the sedimentation: above a very clear unconformity, clastic deposits abruptly evolved to carbonate-evaporitic deposits. Playa-lake facies developed on a classic proximal-distal depositional profile: halite and anhydrite deposits in the centre, carbonate on the sides and alluvial fan deposits on the margins (Fig. 13). The Bas-Dauphiné and Durance domains facies distribution demonstrate that the depocenters were located near active structures, the Valence and Durance faults respectively. Concerning the Camargue/Mormoiron lake system, the depocenter was in the northeastern area, where massive anhydrite beds are outcropping (Fig. 13). The Provence Domain contained only carbonate deposits in the S2. As for S1, there is no evidence of connections between the domains. In the entire Rhône Valley Basin, the clastic supply was very limited and the few alluvial fans that developed around evaporitic domains were exclusively sourced by the dismantling of Pyrenean-Provençal structures, which are now considered as inactive (Séranne et al., 2021).

### 5.2.3. S2 – maximum flooding – late early to early middle Rupelian – ~32.9–30.9 Ma

During this period, the Flysch Basin reached its maximum extension and foraminifera rich marls of deep marine environment were deposited in the Frontal Basin (Fig. 14). The orogenic wedge provided most of the sedimentary material in a single northward flowing routing system located in the foredeep. To the south, the erosion of the upper part of the subduction complex (Helminthoïdes Flysch terrains of the Autapie

nappe) fed the Schistes à Blocs complex (Kerckhove, 1969, 1975), while to the north, serpentinite and metamorphic fragments attests to an internal Alps source for the Grès de la Souloise and Grès de Taveyanne. Active volcanism provided some material to the basin (31.1 ± 0.4 Ma, 30.8 ± 0.4 Ma, Ar–Ar dating in the Grès de Taveyanne; Féraud et al., 1995).

To the west, a marine transgression took place in the entire Rhône Valley (Fig. 14). In the Bas-Dauphiné Domain, our results and revised biostratigraphic interpretation indicate that the transgression initiated in the upper part of the Formation salifère inférieure (T4 Unit; Dumas, 1988). In all the Rhône Valley Basin, the transition from playa-lake to calcareous nannofossils-rich offshore marls, is quite sharp. We assume that the Diois-Baronnies massif characterized by E-W structures was deformed during the Pyrenean-Provençal shortening and thus formed a topographic barrier since the Priabonian. The connection between the upper and middle Rhône Valley may have been through a narrow seaway that flooded the Réauville Ridge and passed along the eastern border of the Massif Central (Fig. 14). However, the occurrence of wave ripples in the Montmirail Domain involves a basin large enough to generate waves caused by wind fetch. We consider that the transgression was limited to the south by the Pyrenean-Provençal-Corso-Sardinian massif, to the east by the Valensole High, and to the north by the Ventoux-Lure and Diois-Baronnies massifs. The Alpes and Luberon Pyrenean-Provençal folds may also have formed palaeohighs between which transgressions connected the Camargue, Provence and Durance domains.



**Fig. 15.** Palaeogeographic and source-to-sink map of the WAFB during late middle to early late Rupelian (S2 sequence - HST). Abbreviations: CC=Conglomerats de Clumanc; CSL = Conglomerats de St-Lions; FD=Formation de la Destrousse; FDI=Formation détritque intermédiaire; FMI=Formation marneuse intermédiaire; GB = Grès de Bonneville; GBA = Grès du Bois d'Asson; GD = Grès des Déserts; GS = Grès de Sénez; MLB = Montbrun-les-Bains; NADS = Northern Alpine Drainage System; SADS=Southern Alpine Drainage System; SSD=Silts de St-Disdier; WADS=Western Alpine Drainage System.

#### 5.2.4. S2 – highstand – late middle to early late Rupelian – ~30.9–28.8 Ma

Significant palaeogeographic changes occurred between late middle and early late Rupelian times. This period is characterized by massive input of clastic material and the development of fluvio-deltaic systems in all domains (Fig. 15). To the East, the orogenic wedge already possibly reached a mean elevation of about 1900 m (Fauquette et al., 2015). The orogen adopted an asymmetric topography with a steep eastern flank (on the retro-side) and a low-angle western flank (Jourdan et al., 2012, 2013; Fauquette et al., 2015) (Fig. 15). In the Frontal Basin, the remnant Flysch Basin progressively retreated northward into the NAFB, while an isolated sea persisted to the south in the Barrême vicinity. The single northward drainage system that developed during the Eocene finally disappeared and gave way to several continental systems supplied from the Internal Alpine Nappes. Based on palaeoflow directions described earlier, we propose that a short transverse drainage system developed southward (Southern Alpine Drainage System; SADS; Fig. 15) and fed the Barrême Domain while a longitudinal drainage system (Northern Alpine Drainage System; NADS; Fig. 15) flowed northward through the Dévoluy and probably connected to the Savoie Domain.

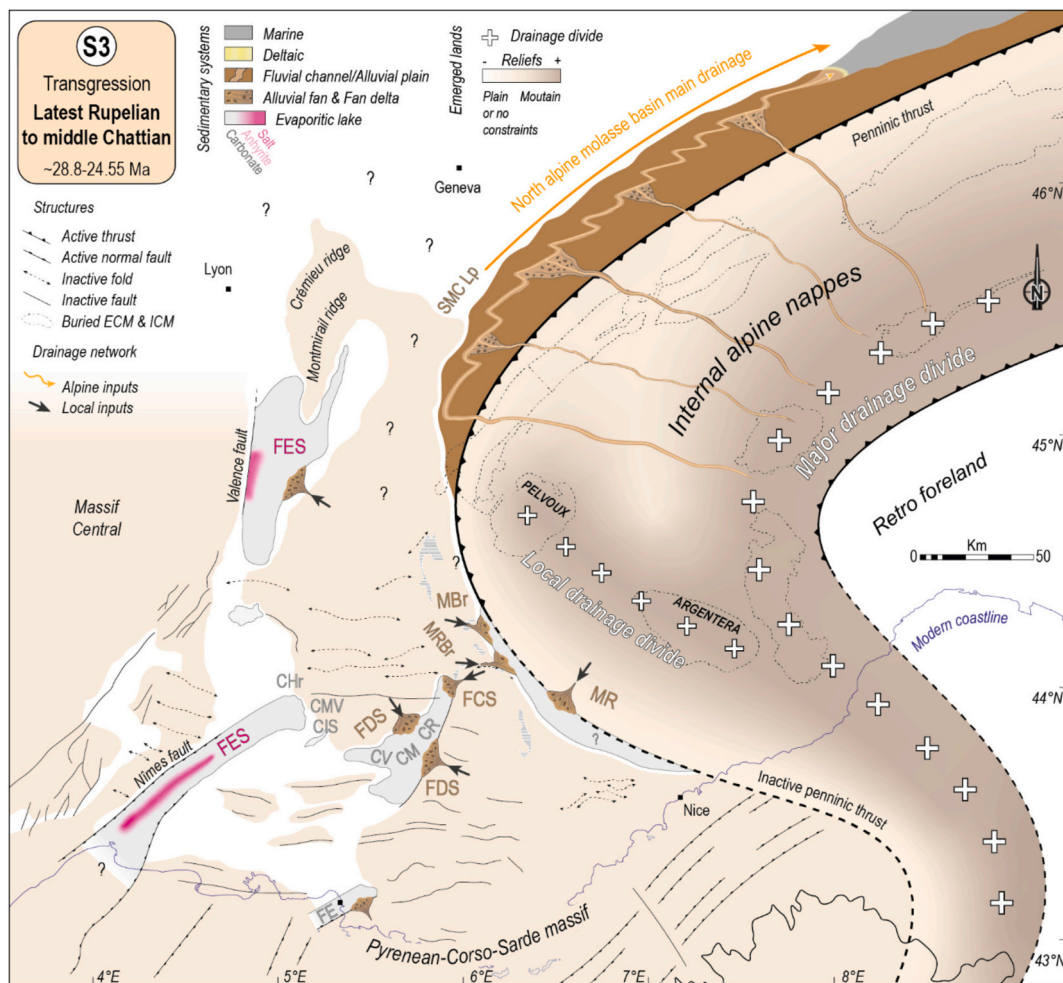
To the west, the Rhône Valley Basin received sedimentary input from various sources. In the Lower Rhône Valley, fluvio-deltaic systems supplied the Camargue, Durance and Provence domains with Pyrenean-Provençal-Corso-Sardinian massif material (Fig. 15). In the Bas-Dauphiné, the absence of Alpine material in the fluvio-deltaic deposits of the Formation marneuse intermédiaire highlights the presence of a

drainage divide between the upper Rhône Valley and the Northern Alpine Drainage System. In contrast, the occurrence of Alpine material in the Montmirail Domain attests for a possible connection between the middle Rhône Valley and the orogenic wedge as early as the middle Rupelian (Western Alpine Drainage System; WADS; Fig. 15). The upstream part of this system may have passed through the Faucon-du-Caire or Esparron-Esclangon domains, but was not subsequently preserved, as S2 is not recorded in these areas.

#### 5.2.5. S3 – transgression – latest Rupelian to middle Chattian – ~28.8–24.55 Ma

At the end of the Rupelian, the Alpine drainage network is composed of one main axial fluvial system that connects the Savoie Domain and the NAFB. It flowed to the northeast and was fed by minor transverse systems sourced from the Internal Alpine Nappes (e.g., Schlunegger et al., 1997a, 1997b; Berger et al., 2005) (Fig. 16). In the other Frontal Basin domains, the Alpine inputs completely stopped. Sedimentation was characterized by local-sourced alluvial fans that possibly retrograded in a single confined basin (Fig. 16). Based on palaeocurrent data recorded in the conglomeratic facies, we interpret the Faucon-du-Caire and Esclangon-Esparron areas as the western border of this basin, and the Barrême area as the eastern one.

In the Rhône Valley Basin, a major unconformity marks the S2/S3 transition. Marine and fluvio-deltaic formations were suddenly replaced by evaporite and carbonate deposits of playa-lake environments



**Fig. 16.** Palaeogeographic and source-to-sink map of the WAFB during latest Rupelian to middle Chattian (S3 sequence - TST). Abbreviations: CHR = Calcaires à Helix ramondi; CIS=Calcaires et gypses de l'Isle-sur-Sorgue; CM = Calcaires des Mournes; CMV=Calcaires et marnes de Vénasque; CR = Calcaires de Reillanne; CV=Calcaires de Vachères; FCS=Formation conglomératique supérieure; FDS=Formation détritique supérieure; FE = Formation de l'Estaque; FES=Formation évaporitique supérieure; MBr = Molasse bréchique; MR = Molasse rouge; MRBr = Molasse rouge bréchique; SMC Lp = Sables et marnes colorés Lower part.

(Fig. 16). The facies distribution is the same as for the early Rupelian (S2 TST). However, in the Camargue/Mormoiron lake system, the depocenter migrated southward, probably because of increasing subsidence related to the opening of the Gulf of Lion.

#### 5.2.6. S3 – maximum Flooding – early late Chattian – ~24.55–23.25 Ma

From the early late Chattian, the Penninic Thrust is considered inactive in the southwestern Alps (Bellahsen et al., 2014) (Fig. 17). The axial fluvial system that established during the latest Rupelian in the NAFB remained active until the late Aquitanian (Berger et al., 2005). In contrast, Alpine derived sediments were still absent in the Frontal Basin and sedimentation occurred in clastic-dominated lake environments that were potentially connected (Faucon-du-Caire/Esparron-Esclanon/Barrême system) (Fig. 17).

To the west, the sea flooded the Rhône Valley for the second time during the Oligocene (Fig. 17). The absence of the characteristic open-sea calcareous nannofossils and the presence of dinoflagellate cysts, echinoderms, sponges, and bryozoans in the fossil record, indicate that the Chattian Sea was probably shallower than its Rupelian equivalent.

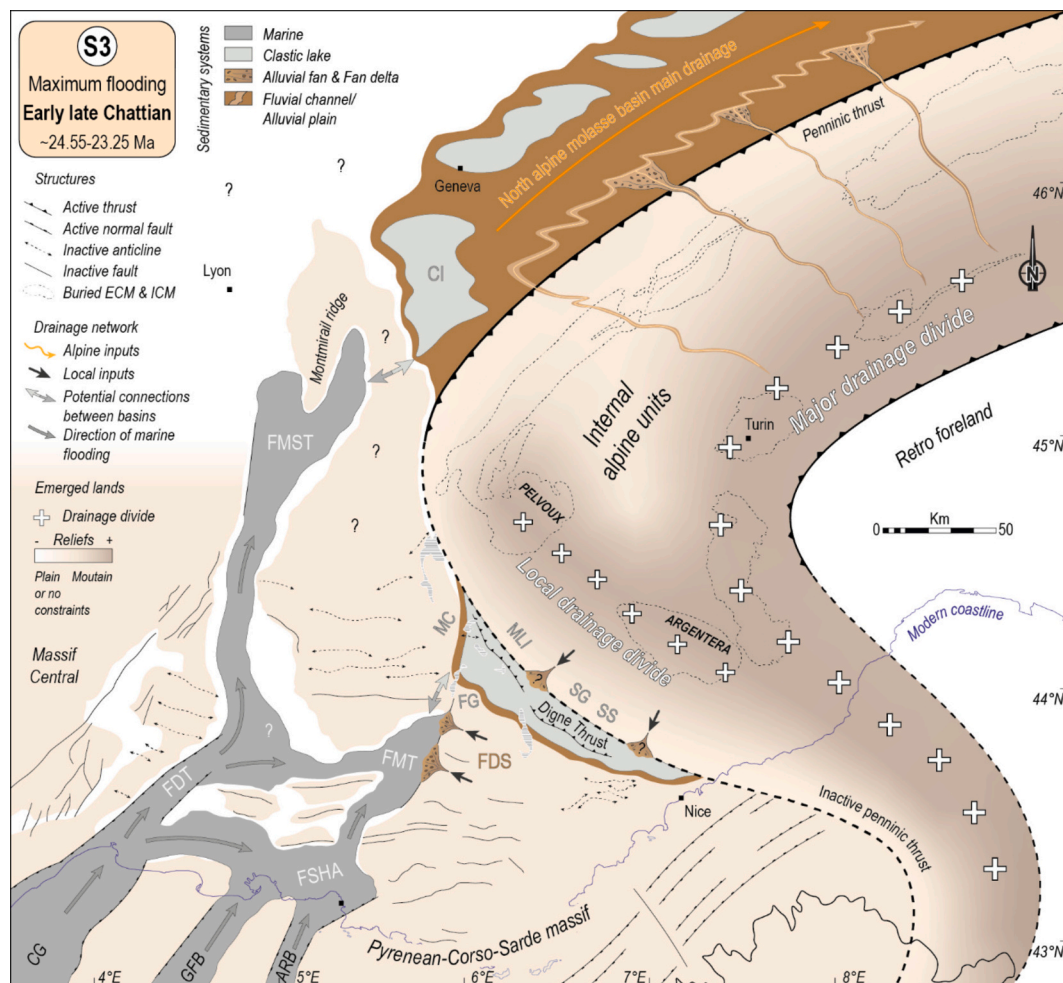
#### 5.2.7. S3 – highstand – latest Chattian - ~23.25–23 Ma

The end of the Chattian is a period of high clastic sediment input in the entire WAFB and most of the domains were sourced from the Internal Alpine Nappes again (Fig. 18). Based on palaeoflow directions described earlier, we propose that the Alpine drainage network

consisted of four main systems. Two equivalent drainage systems of the Southern Alpine Drainage System and Northern Alpine Drainage System were active again, but the upstream part of the Northern Alpine Drainage System was relocated in the Savoie Domain as S3 is not present (or not preserved) in the Dévoluy Domain (Fig. 18). The new equivalent Western Alpine Drainage System flew through the Faucon-du-Caire, Esparron-Esclanon and Sisteron domains (Fig. 18) and probably connected to the Rhône Valley Basin through the Montmirail Domain as during the S2 HST, although, no late Chattian formations were preserved in the middle Rhône Valley at that time. Alpine derived sediments reached for the first time the upper Rhône Valley through the Bas-Dauphiné Domain (Western Alpine Drainage System 2; WADS2; Fig. 18). To the south, the Provence Domain was fed again by the Pyrenean-Provençal-Corso-Sardinian massif.

#### 5.2.8. S4 – transgression – Aquitanian - ~23-? Ma

The S4 sequence palaeogeography is poorly constrained. During the transgression, the sea is still present in the Rhône Valley (Fig. 19) but there is no documentation concerning the nature of sedimentary inputs. In the Frontal Basin, fluvial systems have been replaced by clastic-dominated lakes and alluvial fans fed from the internal Alpine nappes (Fig. 19).



**Fig. 17.** Palaeogeographic and source-to-sink map of the WAFB during early late Chattian (S3 sequence - MFS). The palaeogeography of the Gulf of Lion is from Oudet et al. (2010) and Semmani et al. (2023) (also applied to Figs. 18 and 19). Abbreviations: ARB = Arlesienne Basin; CG = Central Graben; GFB = Grand Faraman Basin; CI = Calcaires inférieurs; FDS = Formation détritique supérieure; FDT = Formation détritique terminale; FG = Formation grise; FMST = Formation marno-sableuse terminale; FMT = Formation marneuse terminale; FSHA = Formation de St-Henry/St-André; MC = Marnes colorées; MLI = Molasse lacustre inférieure; SG = Série grise; SS = Série saumon.

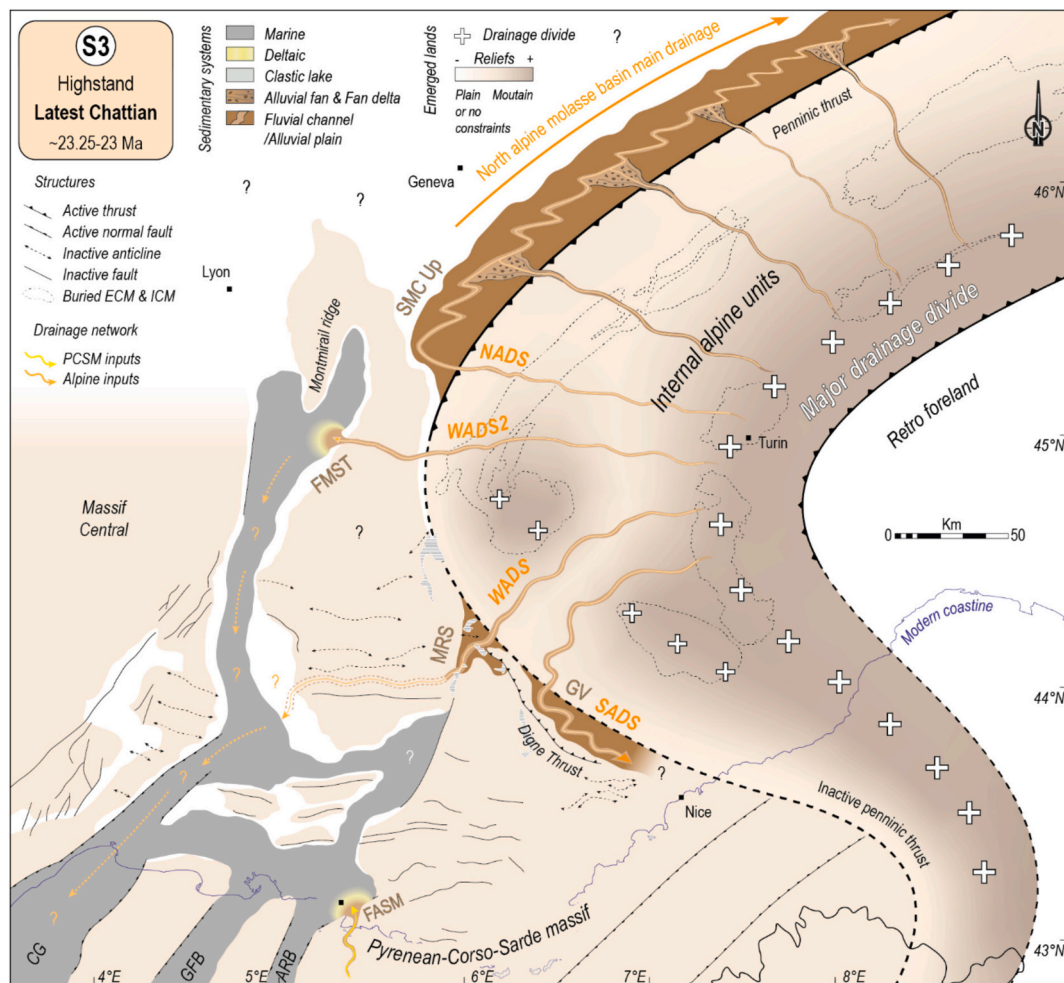
## 6. Discussion

### 6.1. General development of the Alpine drainage network

#### 6.1.1. Priabonian to early Rupelian - underfilled Flysch Basin and European Cenozoic Rift System related sedimentation

Most of this period predates the initiation of the Alpine collision (around 32 Ma; e.g., Niggli, 1986; Bellahsen et al., 2014; Rosenberg et al., 2021 and references therein). The few Alpine sourced sediments resulting from the erosion of the very first early Rupelian Alpine topography were trapped in the marine to deep marine Flysch Basin and are flowing into a single northward longitudinal routing system along the foredeep (Figs. 12, 13, 14 and 20a). The Rhône Valley thus did not receive Alpine material. Sedimentation took place in isolated lakes (i.e. Bas-Dauphiné, Camargue-Mormoiron, Durance domains) which were only fed by short drainage network sourced from local reliefs resulting from the Pyrenean-Provençal orogeny (Figs. 12, 13 and 20a). Subsidence was controlled by active faulting. In the Camargue Domain, the latest stages of the Pyrenean-Provençal orogeny are responsible for the motion of the Nîmes strike-slip Fault (Semmani et al., 2023). The Bas-Dauphiné and Durance domains are classically interpreted as resulting from the hangingwall motion of the Valence and Durance faults related to the European Cenozoic Rift System development (Bergerat, 1987; Bergerat et al., 1990; Lesueur, 1991; Guyonnet-Benaïze et al., 2015).

Based on our results, we here propose to associate the initiation of the European Cenozoic Rift System in the Rhône Valley to the base of S1 (~37.4–36.65 Ma), which corresponds to the very first period of accommodation space creation in the Bas-Dauphiné and Durance domains. It lasted until the S2 TST (~32.9–30.9 Ma) during which syntectonic sedimentation ceased to develop in the Bas-Dauphiné Domain (Fig. 4). In the Durance Domain, similar sedimentary thicknesses in the Apt, Manosque, and Forcalquier sub-domains during the same period (Fig. 8a) suggest that subsidence was no longer influenced by the Durance fault motion, indicating its inactivity. Consequently, we propose that the European Cenozoic Rift System event lasted from at least the base of the Priabonian to the early Rupelian in the Rhône Valley. Even after early Rupelian when extension have ceased, subsidence took place in the Bas-Dauphiné and the Durance domains which were flooded during Rupelian and Chattian transgressions (S2 and S3 MFS) and accommodate the development of a thick late Rupelian-middle Chattian (S3TST) evaporitic sedimentation. Furthermore, the Miocene seaway established at the same place. This probably resulted from subsidence linked to post rift thermal relaxation. This implies that the structural and thermal inheritance of European Cenozoic Rift System played a major role in the evolution of the western Alpine foreland Oligocene-Miocene sedimentation and routing system.

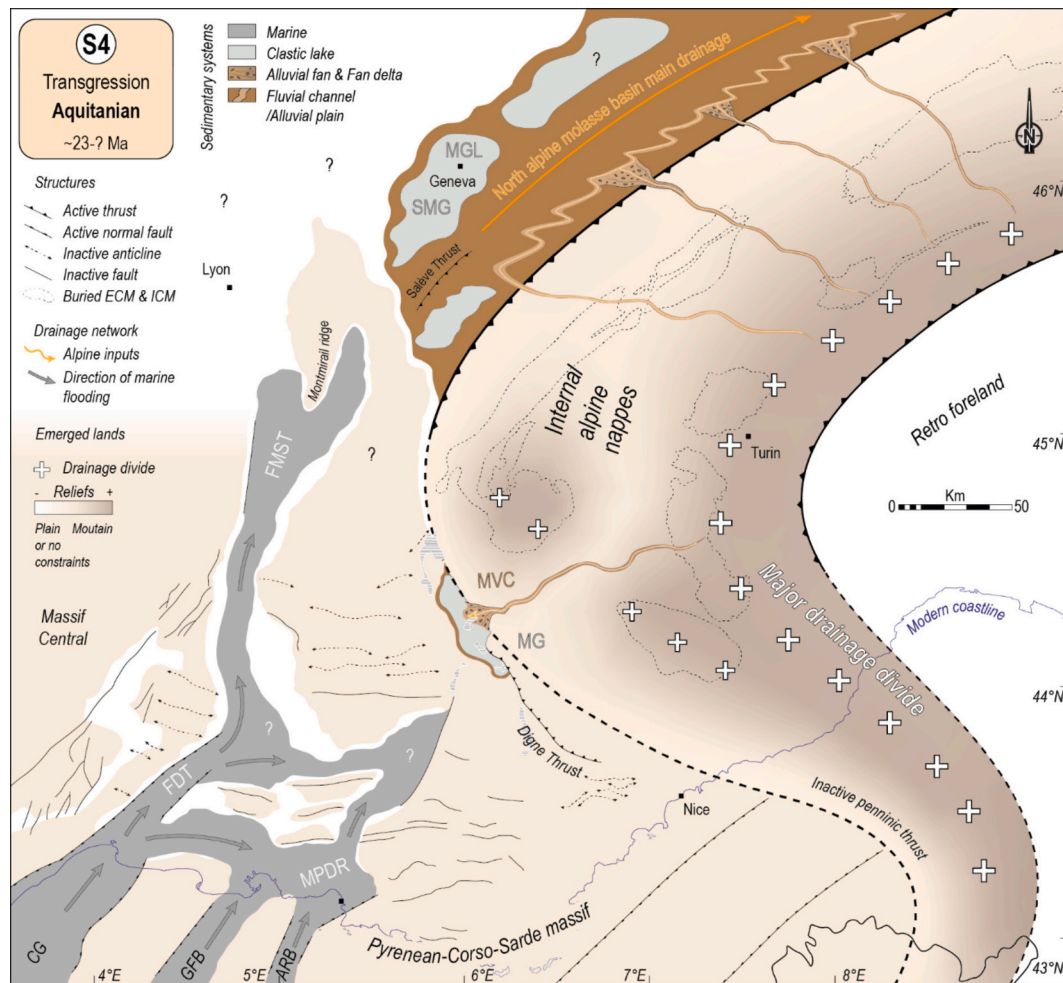


**Fig. 18.** Palaeogeographic and source-to-sink map of the WAFB during late Chattian (S3 sequence - HST). Abbreviations: ARB = Arlesienne Basin; CG = Central Graben; FASM = Formation des Arnavaux/Ste-Marthe; FMST = Formation marno-sableuse terminale; GFB = Grand Faraman Basin; GV = Grès verts; MRS = Molasse rouge sableuse; NADS = Northern Alpine Drainage System; SADS = Southern Alpine Drainage System; WADS = Western Alpine Drainage System; WADS2 = Western Alpine Drainage System 2.

### 6.1.2. Middle Rupelian to early late Rupelian – transition to overfilled Molasse Basin stage and first Alpine drainage network

From ~32 Ma, the internal Alps record strong and fast exhumation (e.g., Malusà et al., 2005; Morag et al., 2008; Bernet et al., 2009; Bernet and Tricart, 2011; Jourdan et al., 2013; Fox et al., 2016) and the refolding of the subduction nappes produced the very first high topography and relief in the orogenic wedge (Jourdan et al., 2012, 2013; Fauquette et al., 2015). The resulting massive export of erosion products caused the rapid infilling of the foredeep (i.e. the Frontal Basin), the retreat of the marine domain and the installation of a continental depositional environment. As a result, the first Alpine drainage network developed. The major drainage divide was in its eastern part and the WAFB rivers drained all the internal Alpine nappes (Fig. 15). Our results suggest that three main Alpine drainage systems (Fig. 15; Northern Alpine Drainage System, Western Alpine Drainage System and Southern Alpine Drainage System) developed in the WAFB during middle Rupelian times (Fig. 20b). They strongly differed in terms of morphology and connections. To the north, the Northern Alpine Drainage System connecting the Dévoluy Domain to the NAFB is mainly longitudinal and takes place in the foredeep (as former Flysch Basin) following the strike of the high orogenic flexure (and the associated subsidence) at the front of the chain during the Early Oligocene (Ford and Lickorish, 2004). To the south, the Western Alpine Drainage System and Southern Alpine Drainage System were isolated from this main drainage system. The Southern Alpine Drainage System was mainly transverse and flowed

directly to the remnant of the foredeep (Barrême Domain) which seems to have trapped most of the sediments, whereas the Western Alpine Drainage System was a long transverse drainage system transporting sediments up to the Rhône Valley Basin. These geometries raise two conclusions: the foredeep was divided into several depocenters and the routing of the Western Alpine Drainage System and Southern Alpine Drainage System was not only influenced by flexural subsidence like the Northern Alpine Drainage System, implying that other mechanisms were affecting the development of the foreland basin. Several parameters influence the different flexural patterns (variations of orogenic loading, variations of elastic thicknesses of lower plate, and deep processes). We know that from the middle Rupelian onwards, flexural subsidence was probably lower in the WAFB than in the NAFB (Ford et al., 1999; Ford and Lickorish, 2004; Ford et al., 2006). The structural inheritance of the lower plate must also be considered, given that the westward oriented propagation of Alpine deformation took place on a pre-structured European continental crust. This inheritance is due to two major geodynamic events from the Dévoluy to the Provence: (1) a Lower Cretaceous extension mainly expressed in SE France by the opening of the Vocontian trough (e.g., Friès, 1987; de Graciansky and Lemoine, 1988), evidenced by thick Lower Cretaceous depositions and where a thermal anomaly has been recorded (Célini et al., 2023). The resulting thinned crust section probably had a low elastic thickness during Alpine compression. (2) The Pyrenean-Provençal orogeny between the Upper Cretaceous and the Eocene, characterized by thick-skinned tectonics in



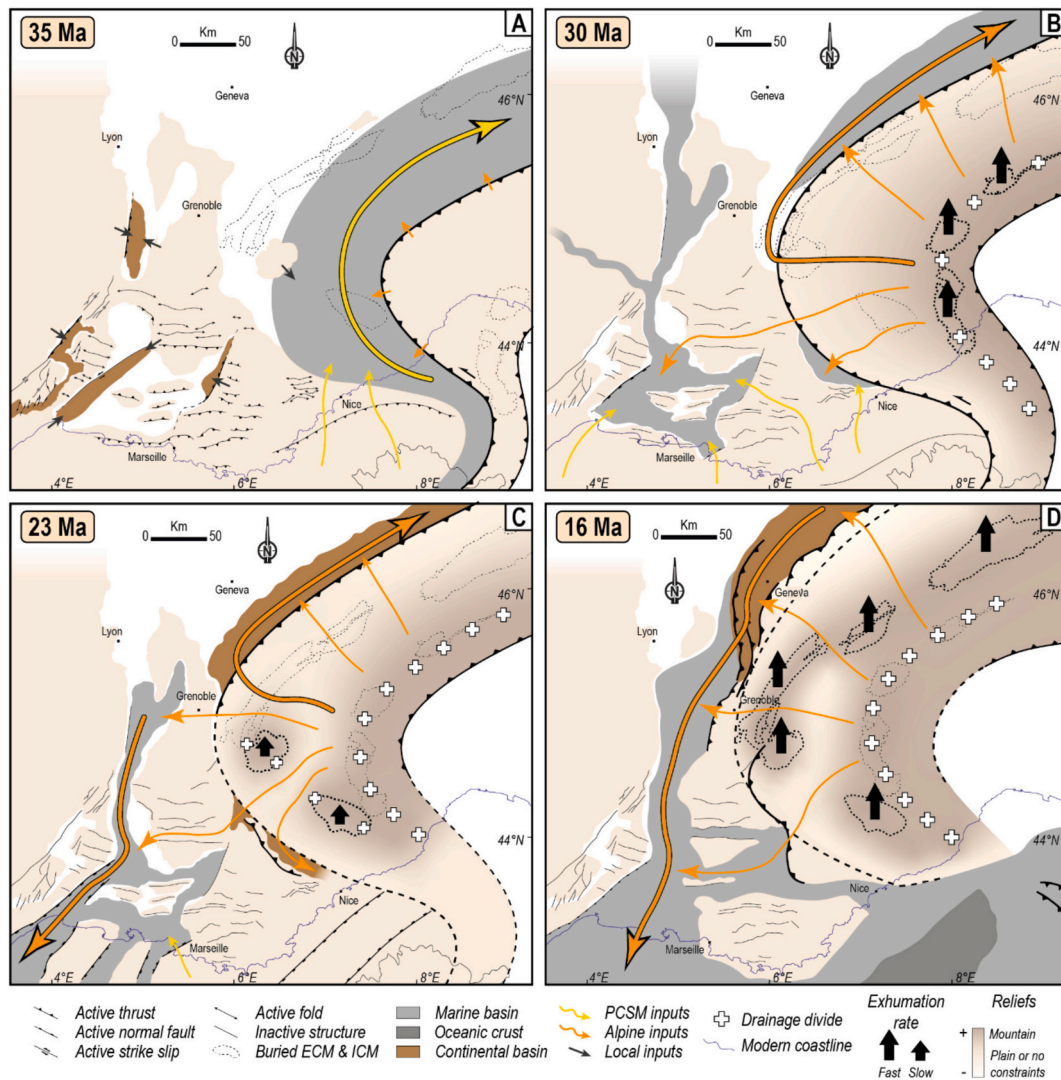
**Fig. 19.** Palaeogeographic and source-to-sink map of the WAFB during Aquitanian (S4 sequence). Abbreviations: ARB = Arlesienne Basin; CG = Central Graben; FDT = Formation détritique terminale; FMST = Formation marno-sableuse terminale; GFB = Grand Faraman Basin; MPDR = Marnes du Puy-du-Roy; MG = Molasse grise; MGL = Molasse grise de Lausanne; MVC = Molasse verte conglomératique; Sables et marnes à gypses.

the Provence and thin-skinned tectonics to the north of the Durance Fault (Bestani et al., 2016). This orogeny is responsible for the formation of E-W trending structures throughout SE France. This compressional event led to changes of thickness, between the Provence and the Devoluy, in the continental crust of the European plate. This process could induce some changes in elastic thickness along a N-S strike for the lower plate during Eocene times. The Pyrenean-Provençal orogeny was associated with strong salt tectonics, well-marked in the Diois-Baronnies and Barles areas from the Lower Jurassic to the Oligocene (e.g., Célini et al., 2021, 2022; Hamon, 2022; Faure et al., 2024), which possibly also impacted the distribution of Oligocene depocenters. Finally, we propose that this inherited structure is mainly responsible for the fragmentation of the foredeep resulting in a broken foreland that influences the development of the Alpine drainage network. In the case of the Southern Alpine Drainage System, the Barrême Domain was bounded by the Castellane Arc to the south and the Valensole High to the west, both of which are considered to have been previously structured by the Pyrenean-Provençal orogeny (e.g., Jourdon et al., 2014; Balansa et al., 2022). We propose that this structuring deformation inhibited the spread propagation of flexural subsidence westwards, and these reliefs prevented the Alpine drainage from spreading outside the Barrême Domain. In the case of the Western Alpine Drainage System, the northern part of the Montmirail Domain was located at the outlet of the Montbrun-les-Bains Valley (Fig. 15), which forms a 70 km long E-W syncline trapped between the Ventoux-Lure and the Diois-Baronnies massifs and was probably deformed during the Pyrenean-Provençal

orogeny (Saillard, 1991). Montenat et al. (2000) demonstrated that a drainage was at least active in this synclinal depression or valley during the Langhian and assimilated it to the Palaeo-Durance. We propose that the Western Alpine Drainage System already flowed through this valley during Rupelian. This seems attested by the occurrence of remains of Rupelian fluvial sediments in the Montbrun-les-Bains syncline containing Alpine material and Eocene planktonic foraminifera (probably reworked from the Flysch Basin deposits) (Saillard, 1991). Rifting inheritance, Pyrenean-Provençal structures and variations of the flexural response to the Alpine orogen may have favoured the connection between the frontal domain (which will alternate between a bypass zone and a deposit zone) and the middle Rhône Valley. We think that the resulting drainage pattern was directed by inherited E-W synclines or depressions, and possibly by salt tectonics.

### 6.1.3. Latest Rupelian to middle Chattian – retreat of the Alpine drainage network

From the latest Rupelian to the middle Chattian, clastic sedimentary material sourced from the Internal Alps was absent in the Rhône Valley Basin and almost completely from the Frontal Basin, except for the Savoie Domain. This indicates strong modifications compared to the middle to early late Rupelian drainage network, and a differentiation of source areas between the northern and southern Frontal Basin domains. This source shift has already been highlighted and discussed at smaller scale in the Barrême Domain, based on thermochronological (Bernet et al., 2009; Jourdon et al., 2013) and petrological (Evans and Mange-



**Fig. 20.** Palaeogeographic and simplified source-to-sink maps illustrating the four major phases of drainage network development in the WAFB. (A) At 35 Ma, the WAFB is divided into two independent sedimentation domains: to the west, the European Cenozoic Rift System in the Rhône Valley, and to the east, the Flysch Basin. (B) At 30 Ma, the Alpine drainage network begins to develop. In the south, the WAFB is fragmented. Sedimentation is either trapped in the Frontal Basin or exported to the Rhône Valley Basin via a system following a pathway inherited from Pyrenean–Provencal compression. In the north, Alpine sedimentary material is exported to a longitudinal drainage system connected to the NAFB, transporting sediments toward the NE. (C) At 23 Ma, North to south longitudinal sediment routing is initiated in the Rhône Valley. Material is exported to the Gulf of Lion, and the connection with the orogenic wedge is ensured through two transverse drainage systems. North of the WAFB and within the NAFB, the sediment-routing configuration remains unchanged. (D) At 16 Ma, sediment routing north of the WAFB and within the NAFB is reversed. All sediments derived from the dismantling of the Western Alps flow through the Rhône Valley and are exported to the Gulf of Lion. The following references apply to Fig. 20d: the position of the Miocene seaway and the Subalpine chains at 16 Ma is from Kalifi et al. (2022); the position of the NAFB is from Berger et al. (2005); the palaeogeography of the Gulf of Lion is from Romagny et al. (2020); the position of the Alpine front, the external and internal crystalline massifs is from Bellahsen et al. (2014), Girault et al. (2022) and Bellahsen et al. (in prep).

Rajetzky, 1991; Schwartz et al., 2012; Jourdan et al., 2012) data and has been related to the thrusting of the Digne Nappe (Jourdan et al., 2012). On a larger scale, at the Rupelian/Chattian transition, the southern External Crystalline Massifs started to record slow exhumation (26–29 Myr; Argentera, Pelvoux and Grandes Rousses massifs; Bigot-Cormier et al., 2006; Van Der Beek et al., 2010; Girault et al., 2022), which could be associated with relief growth and the occurrence of a new local drainage divide on top of the uplifting External Crystalline Massifs (Fig. 16), to the west of the inactive penninic thrust. As suggested by Mark et al. (2016), we propose that this event caused the trapping of internal Alpine material inside the orogenic wedge and the shrinkage of the Southern Alpine Drainage System and Western Alpine Drainage System. To the north, the northern External Crystalline Massifs recorded a later onset of exhumation (~18–22 Myr; Aar, Aiguilles Rouges, Mont Blanc, Northern Belledonne massifs; Boutoux et al., 2016; Berger et al.,

2020; Girault et al., 2022) and no major thrust initiated before the late Chattian. Consequently, the Northern Alpine Drainage System sediment routing remained unchanged in the Savoie Domain and NAFB.

#### 6.1.4. Late Chattian to Aquitanian – sediment routing into the Rhône Valley

During the late Chattian, a routing system connecting the internal Alps to the WAFB was restored. In the orogen, the Argentera and Pelvoux External Crystalline Massifs continued their slow exhumation (Bigot-Cormier et al., 2006; Van Der Beek et al., 2010; Girault et al., 2022). As suggested by Jourdan et al. (2012), the reappearance of Alpine sourced sediments in the domains could be simply the consequence of river headward erosion or capture (Fig. 18). The layout of the Alpine drainage network was seemingly about the same as for the middle to early late Rupelian. To the south, the containment of the Southern Alpine Drainage System in the Barrême area was probably amplified by

its incorporation into the Digne Nappe as a wedge-top basin (e.g., Cru-meyrolle et al., 1991; Ford et al., 2006). A main change occurred in the northern part where Alpine sedimentation reached for the first time the upper Rhône Valley (Bas-Dauphiné Domain) through the Western Alpine Drainage System 2. Considering the orogenic growth, this connection seems more compatible with a combination between slope changes related to the advance of the orogenic front and river capture rather than disappearance of Middle Oligocene relief in front of the orogen. We can therefore interpret the Western Alpine Drainage System 2 as the Palaeo-Isère. Poorly preserved in the entire Rhône Valley, late Chattian/Aquitanian deposits are mainly preserved in the most subsiding areas: the Bas-Dauphiné and Camargue domains. However, given the presence of Alpine material in the Sisteron Domain located at the entrance of the Montbrun-les-Bains Valley (Fig. 18), we assume that the connection with the Montmirail Domain in the middle Rhône Valley was ensured through a drainage system (passing through the current Esclançon-Esparron and/or Faucon-du-Caire domains) that could therefore be considered as the Palaeo-Durance. In the Provence Domain, the presence of Alpine sourced material in the Sables de Figons formation (S4) attests to the first routing system from the Alpine domain to the southern Rhône Valley. These sediments originated from either the Palaeo-Durance or the Palaeo-Isère implying some routing of Alpine sourced material through the northern or middle Rhône valley, and their final transit to the Chattian-Aquitanian Sea. This attests to the establishment of a N-S longitudinal transfer of Alpine-derived sediments within the Rhône Valley, in a system that can be referred to as the Proto-Rhône. This routing configuration therefore initiated at the end of the Chattian/Aquitanian, nearly 5 Myrs earlier than previously described in the literature (Kalifi et al., 2022), and prior to the transition to the marine Molasse (Fig. 20c). On a broader scale, this period coincides with a shift in the geodynamic regime of the Gulf of Lion. According to Bache et al. (2010) and Jolivet et al. (2015), during the Oligocene rifting phase, the continental margin maintained a relatively high topographic position. At the beginning of the Aquitanian, which marks the transition to the post-rift stage, subsidence progressively increased first slowly and, from the Burdigalian onward, increased more rapidly, thereby allowing for the rapid accumulation of sediments. Consequently, at the Oligocene–Miocene transition, the gradual increase in accommodation space within the Gulf of Lion resulted from the progressive opening of a new sink. This process likely triggered a reorganisation of the source-to-sink system within the WAFB, characterized by a southward migration of the main depocenter and sediment transport pathways toward the area of greatest accommodation. Consequently, the downstream part of the Rhône Valley evolved from a (more or less overfilled) depositional area to a transfer zone influenced by the post-rift phase in the Gulf of Lion. Although this new routing system is initiated around the Oligocene–Miocene transition, it does not appear to become fully operational until the late Burdigalian. At that time, Marine molasses are encountered throughout the Rhône Valleys and significant volumes are by passing toward the Gulf of Lion where thick marine series are accumulated (Gorini et al., 1993; Guennoc et al., 2000; Bache et al., 2010; Jolivet et al., 2015). This timing is consistent with the S–N to N–S reversal of sediment routing in the NAFB (16–15 Ma; Berger et al., 2005; Kuhlemann and Kempf, 2002) and with its connection to drainage in the Rhône Valley (Fig. 20d). As a result, from the Burdigalian–Langhian transition onward, the Gulf of Lion becomes the main sink for all sediment derived from the Western and Central Alps.

## 6.2. Origin of transgressions in the Rhône Valley

### 6.2.1. Rupelian Sea

The occurrence of a Rupelian transgression in the Rhône Valley has been discussed many times (Touraine, 1976; Dumas, 1986; Nury, 1987; Pairis, 1987; Dumas, 1988; Lesueur, 1991; Busson et al., 1992; Châteauneuf and Nury, 1995; Châteauneuf et al., 2019; Sissingh, 2001; Semmani et al., 2023; Suc et al., 2025) but never fully documented and

often qualified as “marine incursions”. Here, we bring new field and biostratigraphic data attesting for an open but also perennial marine domain in the Bas-Dauphiné (NP23 calcareous nannofossils) (Fig. 3) and Montmirail domains (NP22 calcareous nannofossils and marine facies) (Fig. 5). Added to previous works, it indicates the presence of a seaway throughout the entire Rhône Valley (Figs. 14 and 15) between late early and early late Rupelian. The occurrence of a Rupelian Sea raises the question of the origin of a connection between the Rhône Valley and an ocean. Considering our present knowledge of the palaeogeography during the Rupelian, there are three possible pathways: (1) the Alpine Sea (i.e. the Flysch Basin) to the east, (2) the North Sea through the Rhine and Bresse grabens to the north and (3) the Massif Central through the Limagne graben to the west. The Alpine Sea is the most probable origin for many authors (Dumas, 1988; Lesueur, 1991; Sissingh, 2001; Châteauneuf et al., 2019; Semmani et al., 2023) and two marine pathways have been proposed to connect with the upper and the lower Rhône Valley. The first one, named the “Crest Sill” (Sissingh, 2001), supposedly linked the Bas-Dauphiné and Dévoluy domains through an unsupported E-W corridor crossing the northern part of the Diois-Baronnies massif. The second marine pathway (Semmani et al., 2023) supposedly connected the Alpine Sea with the Durance Domain through a set of synclines in the northern part of the Provençal fold and thrust belt. We consider these hypotheses unlikely for two reasons: (1) these pathways would imply a slalom of transgressions between a set of Pyrenean-Provençal structures of various orientations that probably acted as topographic barriers since the Eocene (e.g., Flandrin, 1966; Gidon et al., 1971; Montenat et al., 2005; Bestani et al., 2015; Bestani et al., 2016; Balansa et al., 2022); (2) Westward marine transgression entering the Bas-Dauphiné and Durance domains from the Frontal Basin would supply them with Alpine material, which is not the case during the Rupelian. Consequently, it is unlikely that the Rhône Valley transgressions originated from the connections with the Alpine Sea. Located just a few dozen kilometres to the west of the Rhône Valley, the Limagne grabens belonging to the European Cenozoic Rift System (Bergerat, 1987; Bergerat et al., 1990; Merle et al., 1998; Merle and Michon, 2001; Dèzes et al., 2004) and enclosed in the Massif Central could be a candidate for a marine connection with the Rhône Valley Basin. Although they were mainly continental basins (e.g., Gorin, 1975; Huguéney et al., 1999; Roche et al., 2018),  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios on biogenic calcite (Briot and Poidevin, 1998) and the occurrence of marine species (Rey, 1971; Turland et al., 1994; Bodergat et al., 1999) attest for episodic marine incursions probably originating from the Paris Basin to the north (Gorin, 1975; Bodergat et al., 1999). As the Massif Central did not contain high topography during the Oligocene (Barbarand et al., 2001), it seems possible that a marine seaway formed between the Limagne grabens and the Rhône Valley Basin (Figs. 14 and 15). However, the timing of marine deposits in the Limagne area is poorly constrained and needs to be revised in order to evaluate the synchronicity with the Rupelian transgression in the Rhône Valley Basin. Although it has never been documented in previous studies, the connection with the Bresse graben and Rhine graben seems to us the most plausible (Figs. 14 and 15). The upper Rhine Graben experienced marine sedimentation during the Rupelian (e.g., Schuler, 1988, 1990; Pharissat, 1991; Berger et al., 2005; Picot et al., 2008; Pirkenseer et al., 2018) and the first transgression that occurred during the top NP21/base NP22 (Picot et al., 2008) was synchronous with the transgression in the Rhône Valley. As suggested by Kuhlemann and Kempf (2002) and Picot et al. (2008), marine transgressions may also have flooded the Bresse Graben, thereby making a connection with the upper Rhône Valley to the south highly plausible.

### 6.2.2. Chattian and Aquitanian Sea

Before this study, the presence of the sea in the Rhône Valley during the Late Oligocene and Early Miocene have been poorly discussed and the flooding of the entire Rhône Valley was supposed to have occurred during the late Burdigalian (17.8 Ma; Kalifi et al., 2022). Here, we show

that a marine domain existed from the Camargue to the Bas-Dauphiné as early as the late Chattian. During this period, the sea was absent in the Frontal Basin and in the Rhine Graben (Berger et al., 2005; Pirkenseer et al., 2018). Consequently, we propose that S3 and S4 transgressions originated from the south, as the Gulf of Lion had recently opened (Grand Faraman and Arlesienne basins; Figs. 17, 18 and 19; Oudet et al., 2010). Concerning S4 transgression, Kalifi et al. (2022) suggested that late Aquitanian marine inflows in the Sub-Alpine chains and Bas-Dauphiné Domain came from the Ligurian Sea and connected the Nice and Lausanne areas through an internal seaway. To support their hypothesis, Kalifi et al. (2022) used the shallow marine deposits of the St-Disdier Sandstones trapped in the Dévoluy Domain as some potential remains of an Aquitanian Sea in the Frontal Basin. However, our field-work showed that these deposits are highly likely to be part of the Rupelian S2 sequence. Moreover, the Oligocene–Miocene transition corresponds to the period of activation of certain Alpine nappes (e.g., the Digne Nappe; Gidon, 1997; Célini et al., 2022), and thus to the progressive incorporation of the Frontal Basin into the orogenic wedge in the form of transported basins (wedge-top basins). Consequently, it would be very unlikely for the marine domain — already well established in the Rhône Valley during the Chattian — to migrate back into the foreland domain.

### 6.3. Climatic vs tectonic control of the WAFB sedimentation

We showed that the dynamic of the sediment routing system is strongly related to the dynamic of the orogenic wedge. We here question the influence of climatic controls. Our results show that depositional sequences are well correlated between the Frontal Basin and the Rhône Valley Basin, both in terms of timing of shifts and nature of sedimentation (Fig. 11), with alternating periods of clastic dominated sedimentation (mainly during S2 and S3 HST) and carbonate/evaporite dominated sedimentation (mainly during S2 and S3 TST) over time ranges of several million years. The simultaneous occurrence of carbonate–evaporite sequences indicates that sedimentation across the WAFB was sometimes driven more by climatic factors than by tectonic influences. Comparison of  $\delta^{18}\text{O}$  records (Cramer et al., 2011; Westerhold et al., 2020) and eustatic sea level curves (Miller et al., 2020) shows a good correlation between the ages of our sequence boundaries and periods of low sea level (Fig. 11). The S1 SB is not well constrained but formed during the MP17 mammal biozone, which corresponds to the minimum eustatic sea level of the Priabonian. The S2, S3, S4 SBs are broadly simultaneous with Oi1 (33.65 Ma), Oi2 (27.9 Ma) and Mi1 (23 Ma)  $\delta^{18}\text{O}$  events, that correspond to cooling periods and sea-level falls related to major glaciations stages in the southern hemisphere (e.g., Miller et al., 1991; Zachos et al., 1996; Miller et al., 2020). Conversely, our MFSs and HSTs match periods of high sea level (Fig. 11). The exact climatic control of eustatic sea level change on these sequences can be questioned: 1) low magnitude marine base level variations are not likely to drive periods of clastic sediment supply vs. carbonate–evaporite deposition, moreover when connection to the marine domain is not clear. 2) The observations outlined here suggest a correlation between low sea level periods and low clastic inputs (S2 and S3 TSTs), and high sea level periods and high clastic inputs (S2 and S3 HSTs). Warm climates are often associated with high rainfall rates, favouring erosion and the export of large amounts of clastic material into basins, as has often been documented in foreland basins (e.g., Drummond et al., 1996; Chen et al., 2018; Vaucher et al., 2023). On the other hand, cold and dry climates could limit mechanical erosion in a mountain belt that had not yet developed mountain glaciers and therefore may lead to the deposition of evaporites in lowlands, as during the Badenian salinity crisis (Mi3b glacial event; de Leeuw et al., 2010). According to the recent compilation by Suc et al. (2025), the whole of SE France was subjected to fairly humid climatic conditions (high mean annual precipitation) during the early late Rupelian (~29 Ma) and late Chattian (~23–24 Ma) (Fig. 11), i.e. during the deposition of the S2 and S3 HSTs respectively,

confirming the association of strong pulses of clastic sediments with wetter conditions. Conversely, after a phase of marked humidity, Suc et al. (2025) show a strong aridity (decrease in mean annual precipitation) during the early Rupelian, i.e. during the deposition of the evaporites and the Calcaires nummulitiques of S2 TST. Such conditions were also recorded between the late Rupelian and middle Chattian (S3 early TST), even though the uncertainties are greater (Fig. 11). The correlation between the long-term precipitation variations and variations in clastic input suggest a close relationship. However, above we had associated decreasing clastic inputs, disappearance of Internal Alps derived material and Alpine drainage migration recorded in the Frontal Basin during S3 TST (i.e. Molasse rouge bréchique) to a tectonic factor involving the exhumation of the southern External Crystalline Massifs (Pelvoux and Argentera massifs), causing a reorganisation of the drainage network. In the Savoie Domain routing system, which was not affected upstream by the exhumation of the southern External Crystalline Massifs, S3 TST is associated with the deposition of a very thick fluvial series (also recorded in the NAFB; e.g., Kuhlemann, 2000; Berger et al., 2005), which implies a massive export of sediment from the orogenic wedge probably related to increasing exhumation in the Internal Alps. This implies that even though climate plays a role in the development of sedimentation during this period, tectonics has a greater influence. Likewise, MFS1 is synchronous with the eustatic peak of the Priabonian which is of greater amplitude (~70 m higher than the current sea level; Miller et al., 2020) than that of MFS2 or MFS3 (~40 m higher than the current sea level; Miller et al., 2020), although MFS1 corresponds to a lacustrine environment and the other are marine. According to our interpretations, the Rupelian transgression spread into the Rhône Valley from the North Sea or the Paris Basin, through the Rhine/Bresse graben system or the Limagne Graben respectively, at a time of European Cenozoic Rift System climax. During the Priabonian, these basins were not rapidly subsiding as the rift was just initiating and therefore could not be connected to the North Sea, once again demonstrating that tectonics are of greater importance than climatically driven eustasy.

## 7. Conclusion

In this paper, we investigated the reorganisation of sediment routing between the underfilled and overfilled phases of the WAFB. To achieve this, we reconstructed the palaeogeographic and source-to-sink evolution of this basin between the Late Eocene and Early Miocene. Our results show that, in addition to the influence of the orogenic wedge, the structural inheritance of the foreland plate (i.e. the European plate) and the extensional events (European Cenozoic Rift System, opening of the Gulf of Lion) that affected southeastern France had a strong impact on the development of sedimentation. During the Priabonian, SE France was divided into two separated basins with different dynamics: to the west, the European Cenozoic Rift System in the Rhône Valley, and to the east, the underfilled phase foredeep linked to the migration of the Alpine proto-wedge. Although European Cenozoic Rift System was active only during the Priabonian–early Rupelian interval, post-rift thermal subsidence likely controlled the deposition of Oligocene–Miocene formations and promoted longitudinal sediment transfer. The middle to late Rupelian marked the transition to overfilled conditions and the filling of the foredeep, accompanied by the development of the first Alpine drainage networks. At this time, two different types of foreland related to inherited structure of the European lithosphere are evidenced: (1) to the north, a classical flexural, quite cylindrical with longitudinal transport took place as a southwest termination of the NAFB; (2) to the south, as a consequence of the Pyrenean-Provençal orogeny, a fragmented broken foreland developed with independent and varied drainage systems. The latter included a transverse E-W drainage system, related to E-W Pyrenean-Provençal structures, that allowed the export of Alpine material from the orogenic wedge to the Rhône Valley as early as the middle Rupelian. Between the late Rupelian and the middle Chattian,

the exhumation of the southern External Crystalline Massifs caused a migration of the major Alpine drainage divide and a break in sedimentary input from the Internal Alps. The late Chattian-Aquitainian period finally marked the progressive establishment of a major longitudinal system in the Rhône Valley (configuring the early proto-Rhône course) flowing southwards to the Gulf of Lion that opened and became the final new sink of the west alpine routing system. The connection with the orogenic wedge was ensured through two transverse systems that we interpret as the Palaeo-Durance and the Palaeo-Isère.

### Declaration of competing interest

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

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### Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.earscirev.2025.105337>.

### Data availability

Data will be made available on request.

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