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Meso-Cenozoic geodynamic evolution of the Paris Basin: 3D stratigraphic constraints

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Abstract – 3D stratigraphic geometries of the intracratonic Meso-Cenozoic Paris Basin were obtained by sequence stratigraphic correlations of around 1 100 wells (well-logs). The basin records the major tectonic events of the western part of the Eurasian Plate, i.e. opening and closure of the Tethys and opening of the Atlantic. From earlier Triassic to Late Jurassic, the Paris Basin was a broad subsiding area in an extensional framework, with a larger size than the present-day basin. During the Aalenian time, the subsidence pattern changes drastically (early stage of the central Atlantic opening). Further steps of the opening of the Ligurian Tethys (base Hettangian, late Pliensbachian;...) and its evolution into an oceanic domain (passive margin, Callovian) are equally recorded in the tectono-sedimentary history. The Lower Cretaceous was characterized by NE–SW compressive medium wavelength unconformities (late Cimmerian–Jurassic/Cretaceous boundary and intra-Berriasian and late Aptian unconformities) coeval with opening of the Bay of Biscay. These unconformities are contemporaneous with a major decrease of the subsidence rate. After an extensional period of subsidence (Albian to Turonian), NE–SW compression started in late Turonian time with major folding during the Late Cretaceous. The Tertiary was a period of very low subsidence in a compressional framework. The second folding stage occurred from the Lutetian to the Lower Oligocene (N–S compression) partly coeval with the E–W extension of the Oligocene rifts. Further compression occurred in the early Burdigalian and the Late Miocene in response to NE–SW shortening. Overall uplift occurred, with erosion, around the Lower/Middle Pleistocene boundary. © 2000 Éditions scientifiques et médicales Elsevier SAS

Paris Basin / stratigraphy / geodynamics / tectonics / Mesozoic / Cenozoic

Résumé – **Evolution tectonique méso-cénozoïque du bassin de Paris: contraintes stratigraphiques 3D** L'image stratigraphique 3D du Bassin de Paris, bassin intracratonique méso-cénozoïque a été obtenue par la corrélation d'environ 1 100 puits (diagraphies) selon les principes de la stratigraphie séquentielle. Il enregistre fidèlement les principaux évènements tectoniques affectant la partie

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occidentale de la plaque ouest-eurasiatique : l'ouverture et la fermeture de la Téthys et l'ouverture de l'Atlantique. De la base du Trias au Jurassique terminal, le Bassin de Paris est un domaine subsident, de superficie supérieure à l'actuel bassin, dans un contexte extensif avec un changement de mode de subsidence durant l'Aalénien (ouverture de l'Atlantique central). Il enregistre les différents stades de l'ouverture de la Téthys ligure (base de l'Hettangien; Pliensbachien terminal;...) et son passage à un domaine océanique (marge passive, Callovien). Le Cretacé inférieur est caractérisé par des discordances de moyenne longeur d'onde selon une compression NE–SW (néo-cimmérien - limite Jurassique/Crétacé et intra-Berriasien - et Aptien terminal) contemporaines de l'ouverture du Golfe de Gascogne. Ces discordances s'accompagnent d'une diminution importante de la vitesse de subsidence. Après une période de subsidence en contexte extensif (Albien à Turonien), un régime de compression NE–SW s'installe dès la fin du Turonien avec une « phase » de plissement majeur à la fin du Crétacé. Le Tertiaire est une période de très faible subsidence dans un contexte compressif. La seconde période de plissement se produit du Lutétien à l'Oligocène inférieur (compression N–S) en partie contemporaine de l'extension E–W associée aux rifts oligocènes. D'autres paroxysmes de compression se produisent à la base du Burdigalien et au Miocène terminal en réponse à une compression NE–SW. Une surrection généralisée du bassin, avec érosion, se produit aux alentours de la limite Pléistocène inférieur/moyen.

Bassin de Paris / stratigraphie / géodynamique / tectonique / Mésozoïque / Cénozoïque

The Paris Basin is certainly among the most extensively studied sedimentary basins. The first studies were published in 1746–1756 by Guettard [1, 2]. In 1867, Guettard, Monnet & Lavoisier [3] drew the first log. Multiple studies have been carried out on the relationships between low amplitude deformations and sedimentation [4–6] see [7]. The Paris Basin is classically considered as an intracratonic basin in a regime of decreased thermal subsidence, coming into existence during a period of rifting in Permo-Triassic times [8–10].

In the early 1990s, detailed analyses of both facies evolution and subsidence trends for the Triassic to Cretaceous time interval suggested that the subsidence history of the Paris Basin could be described as reflecting the superposition of (1) a long-term component and (2) several acceleration/deceleration trends which can be related to changes in intraplate forces [11, 12]. At the same time, geochemical analyses of clay minerals [13–15] indicated different diagenetic events at about 190 Ma (220–250 °C), 150 Ma and 80 Ma. Together, these data suggest a more complex tectonic evolution than previously thought.

The development of sequence stratigraphic concepts and a resumption of petroleum exploration during the 1980s led to a better stratigraphic knowledge of the Paris Basin which can be used as a base for a new geodynamic model. The purpose of our study is to use 3D sedimentary geometries at a basin-scale (sediment thicknesses and palaeogeography) in order to better constrain the geodynamic evolution of the basin. These geometries are based on well-log correlations, using the principles of sequence stratigraphy. Some attempts to apply sequence stratigraphy were made in the late 1980s [11, 16, 17–19].

Our reappraisal of the stratigraphy of the Paris Basin started six years ago. The objective of that project was to obtain high-resolution 2D and 3D geometries of stratigraphic cycles with lithologies and facies in order to constrain the different wavelengths of the tectonic movements and their controls on the nature, geometry and hierarchy of the stratigraphic cycles. This study is part of the GEOFRANCE 3D project: *Imagerie 1:1 000 000 de la géologie de la France - Bassins sédimentaires : interface socle-couverture et grandes surfaces de discontinuité*.

1. 3D geometrical data: a sequence stratigraphic correlation of well-logs

The 3D sedimentary geometries established in this study are based on well-log correlations and not on seismic data, because seismic data acquired by petroleum companies are not in the public domain in France and because in the Paris Basin depositional geometries show little relief (small formational thicknesses over large areas) and are mostly below seismic resolution. Around 1 100 wells have been correlated using the principles of high-resolution sequence stratigraphy, i.e. the stacking pattern of parasequences [20–22] which have already been applied to the Paris Basin in earlier studies [23–25]. The principle of correlation is to identify the transgressive-regressive cycles of shortest duration (20 000 years to 400 000 years) which can be identified on well-logs and which are called parasequences [20, 21] or genetic units [22]. This requires a calibration of the well-logs in terms of sedimentary environment which is achieved by comparison with cores and/or outcrops. The vertical stacking of the parasequences in lower order composite transgressive-regressive units of longer duration provides the means for correlation to the next well. Two lower orders of transgressive-regressive cycles have been recognized: 10–40 My (major cycles) and 1–15 My (minor cycles).

The stratigraphic record [26] is determined by three main factors (1) subsidence (tectonics s.l.), (2) the sea-level variations (eustasy) and (3) the availability (supply/production) of sediments (sediment flux). The accommodation space is defined as the space created by tectonics and eustasy which can be filled by sediments. A stratigraphic transgressive-regressive cycle is the result of the variation of the ratio between A (accommodation) and S (sediment flux). When $A>0$ and $A>S$, retrogradation or transgression occurs, when $A<0$ or $A>0$ and $A<S$, progradation or regression occurs [27]. The surface separation between retrogradation and progradation is called a maximum flooding surface. The surface separation between progradation and retrogradation is called a flooding surface.

The time lines resulting from correlation of the stacking pattern are calibrated by biostratigraphy. Dating is poor for the Triassic, because of the continental environment of most sediments. For example, the Norian has not been palaeontologically characterized in outcrop. In contrast the Jurassic sediments are well dated, based on ammonite zonation and brachiopod marker-beds [17, 28, 29]. Lower Cretaceous sediments which are partly continental or deltaic, were dated by palynomorphs and ammonites [30], Upper Cretaceous chalks by ammonites (Cenomanian and Turonian), benthic foraminifers [31] and Monciardini in [32] and echinoderms. The ages of Tertiary sediments are mainly based on foraminifers, palynomorphs and nannofossils, see [33, 34].

Isopach maps were compiled for time intervals between 3 and 20 My. They are not subsidence maps, however, because of the limited changes in water depth (no more than 150 m in the Paris Basin), which appear to be small in comparison to formational thicknesses, these maps are good approximations of subsidence variations in both space and time. A few higher-resolution 2D sections have been drawn along a W–E transect from Rambouillet, SW of Paris (west) to Nancy (east, see *figure 1a* for location).

2. Geological framework and crustal structure of the Paris Basin

The present-day topography of the Paris Basin *(figure 1b)* shows a morphological contrast between planation surfaces [35] in its central (Brie surface) and western (Paris Basin 'main' surface) parts and the very well developed cuestas in the east. Today the Paris Basin is surrounded by four Cadomian/Variscan basement massifs: the Armorican Massif in the west, the Massif Central in the south, the Vosges in the east and the Ardennes in the northeast. The sedimentary threshold between the Armorican and Central Massifs is called the Poitou High, that between the Massif Central and the Vosges, the Burgundy (Bourgogne) High. Eastwards, the Vosges basement and the Paris Basin are bounded by the Rhine and Bresse Grabens. Both the Paris Basin and the Cadomian/Variscan basements are incised by the present-day river network (Loire, Seine, Meuse and Moselle).

Cretaceous and Tertiary deformation and erosion have exhumed Mesozoic sediments and underlying basement. This process was particularly effective in the eastern part of the basin, exposing the large 'rings' of Mesozoic sediments *(figure 2)*. To the northwest, towards the English Channel, and to the north, towards Belgium, the Paris Basin grades into numerous smaller basins: the north Baie de Seine Basin, the central English Channel Basin, the Portland-Wight Basin, the Weald Basin and the Bruxelles Basin, which are mainly Cretaceous to Tertiary features [37].

The structure of the continental crust beneath the Paris Basin is poorly known. One deep seismic line has been shot in the northern part of the basin, between Evreux in the southwest and Valenciennes in the northeast (ECORS project [38], see *figure 3a* for location). The upper crust is characterized by southwest dipping oblique reflectors and the lower layered crust thins westward towards the Armorican block and disappears in the northeast below the London-Brabant block *(figure 3a)*. The only fault which cuts across both the upper crust and the MOHO is the Bray Fault.

The mean depth of the MOHO [39, 36] *(figure 3b)* is around 35 km becoming shallower toward the present-day basement outcrops. The highest gradient occurs toward the southeast, perpendicular to an axis from the Massif Central (Clermond-Ferrand area) to the Vosges-Schwarzwald (Mulhouse area).

The nature of the upper part of the Paris Basin basement has been established from deep drilling, gravimetric and magnetic data (Autran et al., in [32]), and its Variscan structure has been interpreted in different ways [40]. However, it is now established that the basement belongs to four different Variscan domains *(figure 3)*: the Armorican domain (central-Armorican zone and Cadomian block) in the west, the internal domain (Liguro-Arverne zone and Morvan-Vosges zone) in the south and southeast, the Saxo-Thuringian zone which pinches out to the west in the central part of the basin along the Bray Fault, and the Rheno-Hercynian zone in the north.

Few seismic data are available to constrain the geometry of the faults and folds which were active during Meso-Cenozoic sedimentation. Two kinds of data are available, (1) from outcrops, and (2) from the published map of Perrodon and Zabek [10] showing the main faults identified in Triassic sediments on industry seismic lines. Based on these data, drilling reports (available in the *Service de Conservation des Gisements d'Hydrocarbures*, French Department of Industry) and unpublished PhD theses, a new map has been compiled *(figure 4)*.

The extension, stratigraphy and geodynamic significance of the Permian sediments below the Meso-Cenozoic deposits are still poorly known. Autran et al., in [32], Mascle [41] and Perrodon & Zabek [10] drew different boundaries for the Permian basins. The only point of agreement is the discontinuous and isolated nature of these small basins, a consequence of non-deposition and/or post-depositional subsequent erosion. The most important one is the Saar-Nahe Basin, extending from Germany in the east to the eastern part of the present-day Paris Basin [54, 55]. The relationships with the other French Permian basins, now interpreted as syn- to post-orogenic extensional gravitational collapse basins [56, 57], have not really been discussed. But because of crustal thickening during the Variscan continental collision below the Paris Basin south of the Variscan front, such an origin may be assumed for the Permian basins located below the present-day Paris Basin.

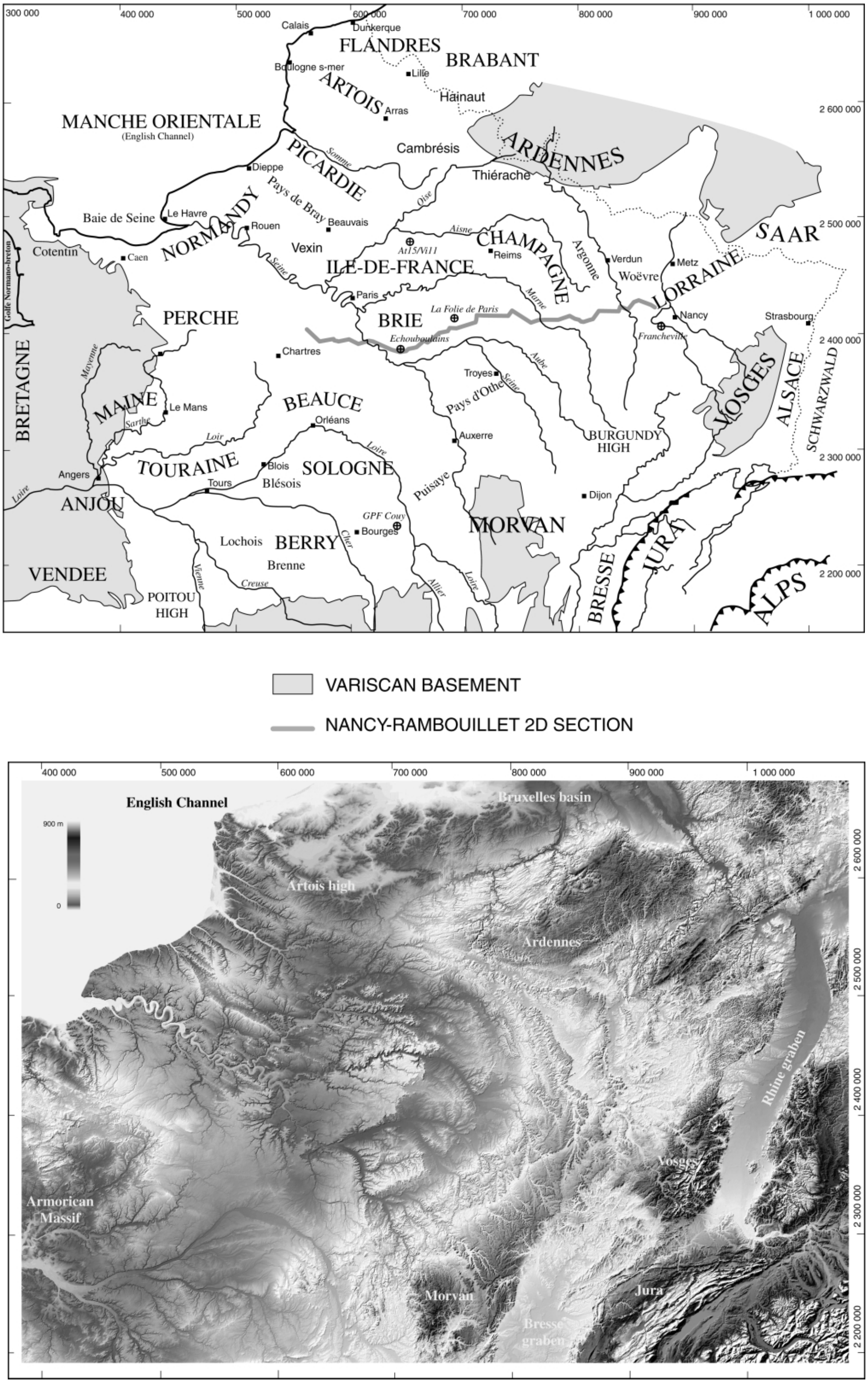


Figure 1a. Main geographic units of the Paris Basin and surrounding areas, **1b.** main geomorphological and geological units of the Paris Basin and surrounding areas (DEM).

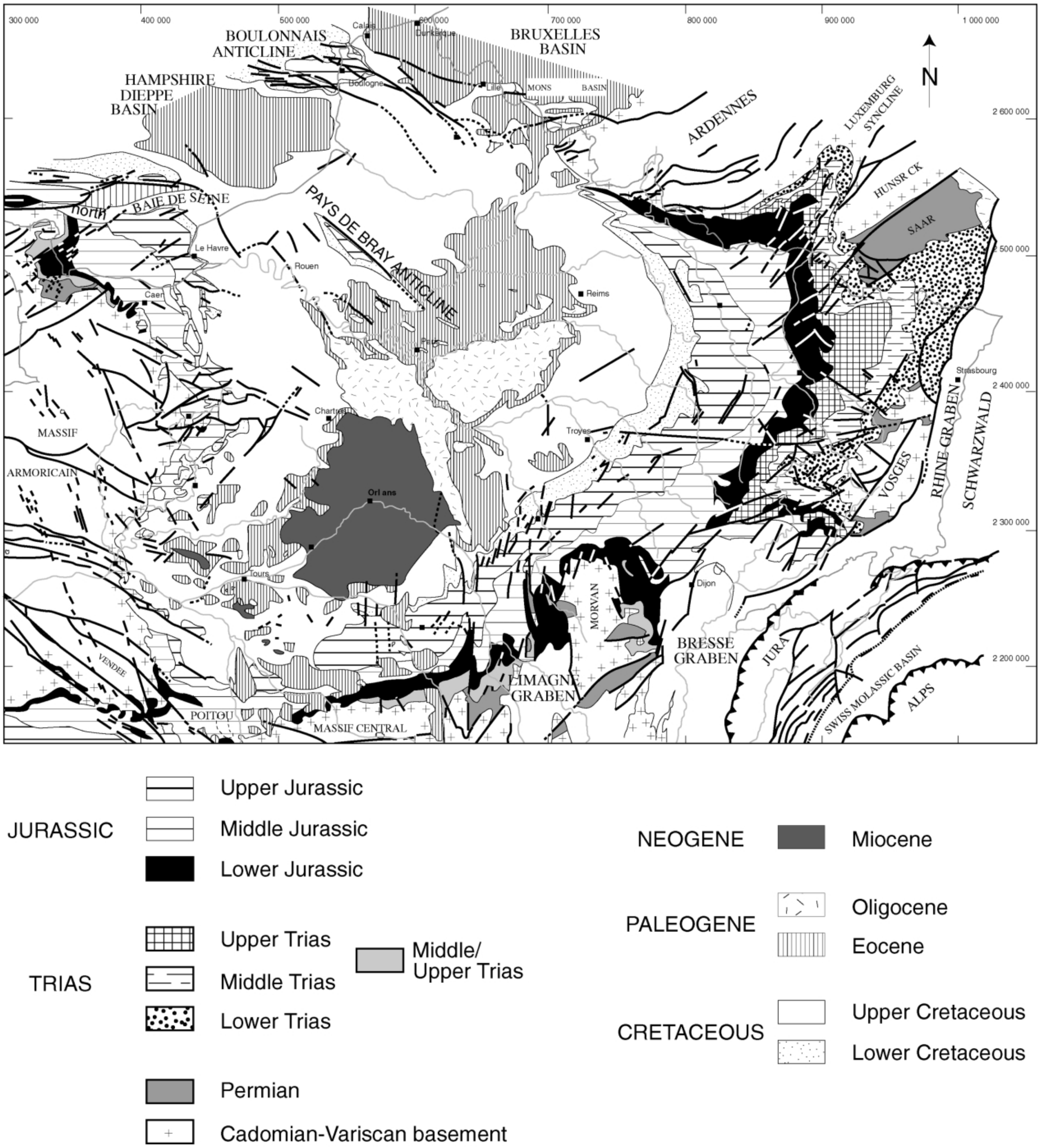


Figure 2. Geological map of the Paris Basin, simplified from the geological map of France at 1:1 000 000 [36].

3. Major stratigraphic cycles and unconformities

The major difficulties of a stratigraphic analysis are (1) to define a hierarchy of the different cycles and (2) to identify the 10–40 My cycles, that record the tectonic evolution of the basin [58, 59] and that can change laterally (boundaries, asymmetry, etc.) in response to variations of the tectonically-controlled accommodation space and/or the sediment supply.

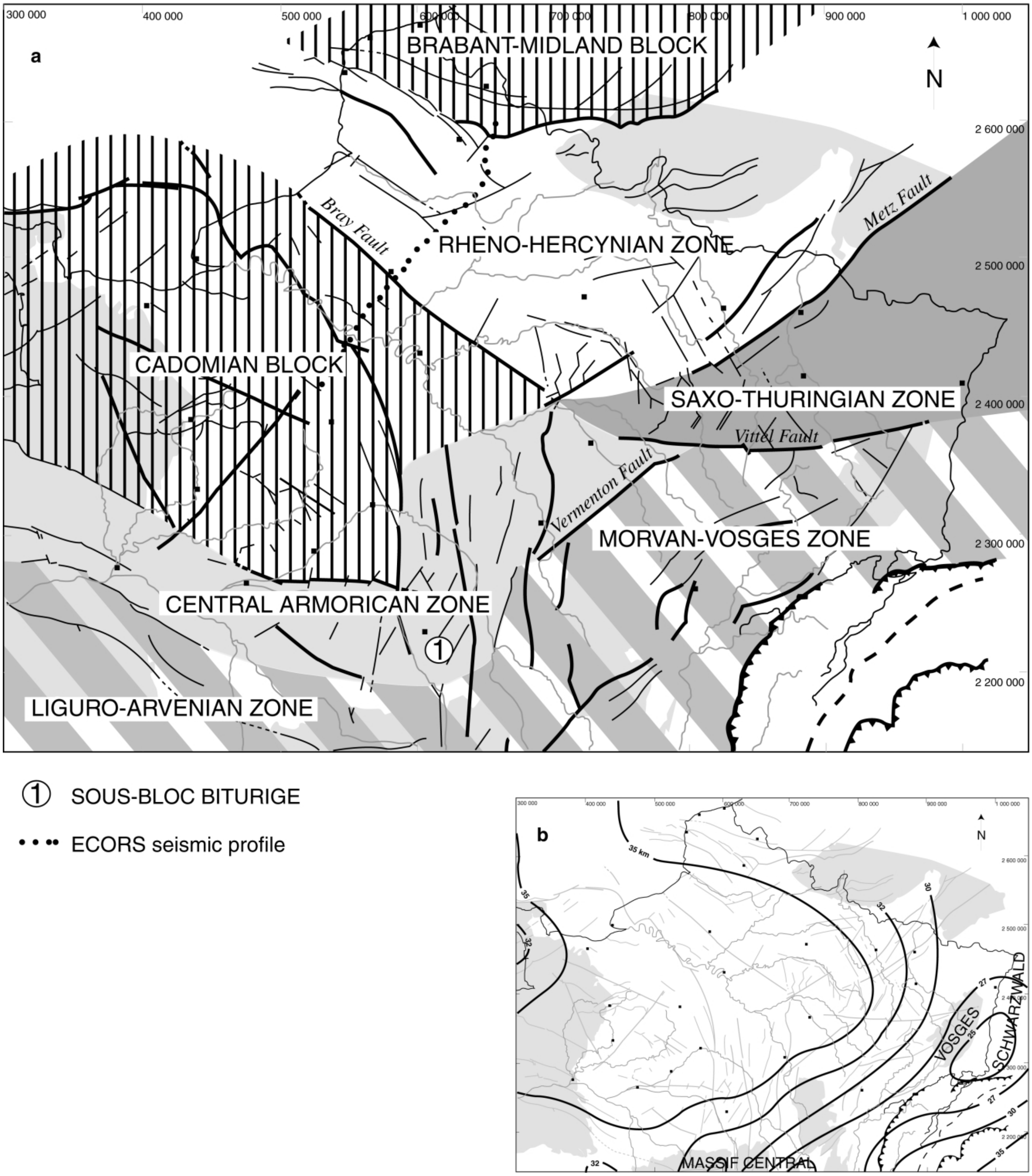


Figure 3a. Main structural domains of the Cadomian-Variscan basement below the Paris Basin, **3b.** Moho isobaths (both from [40, 36]).

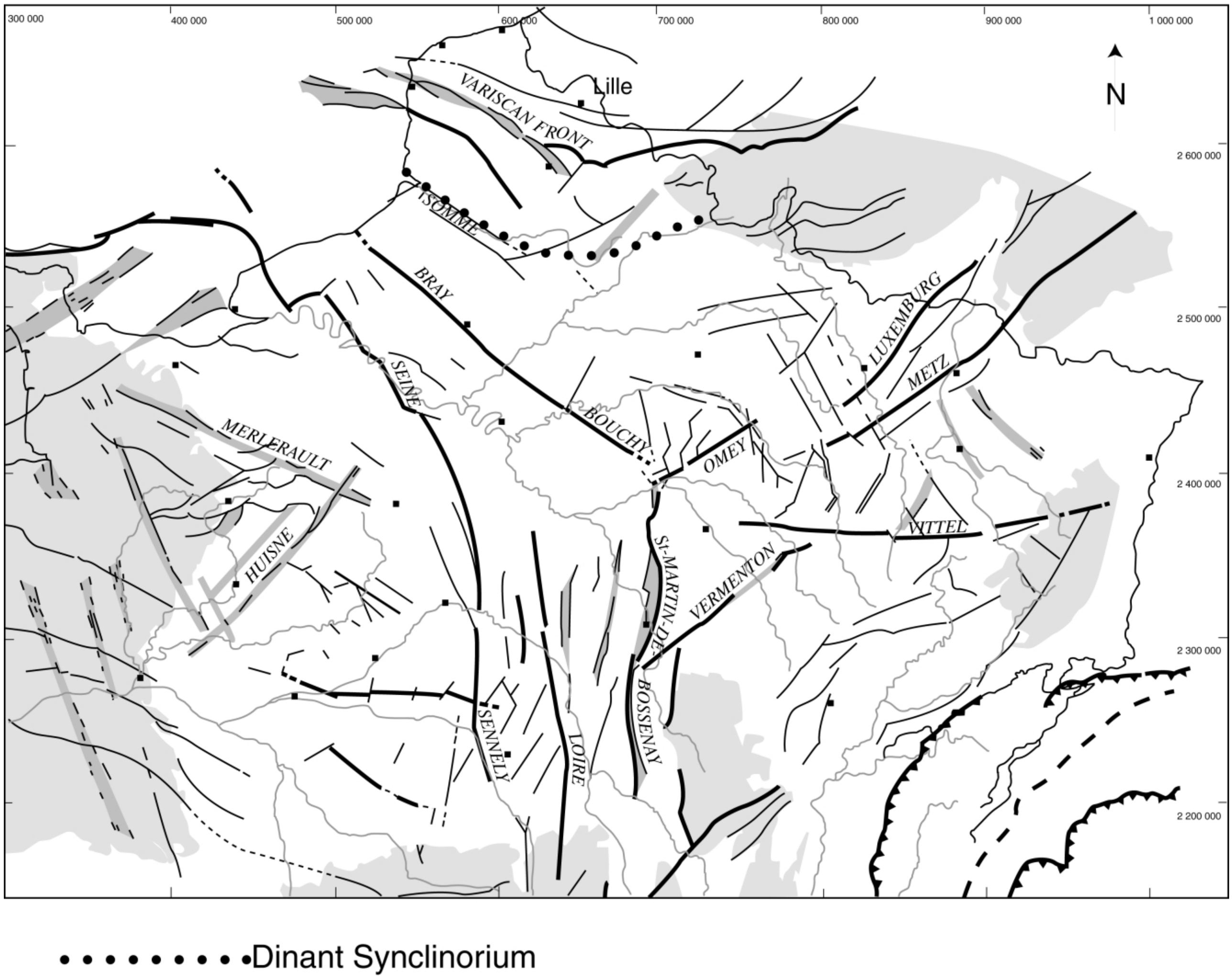


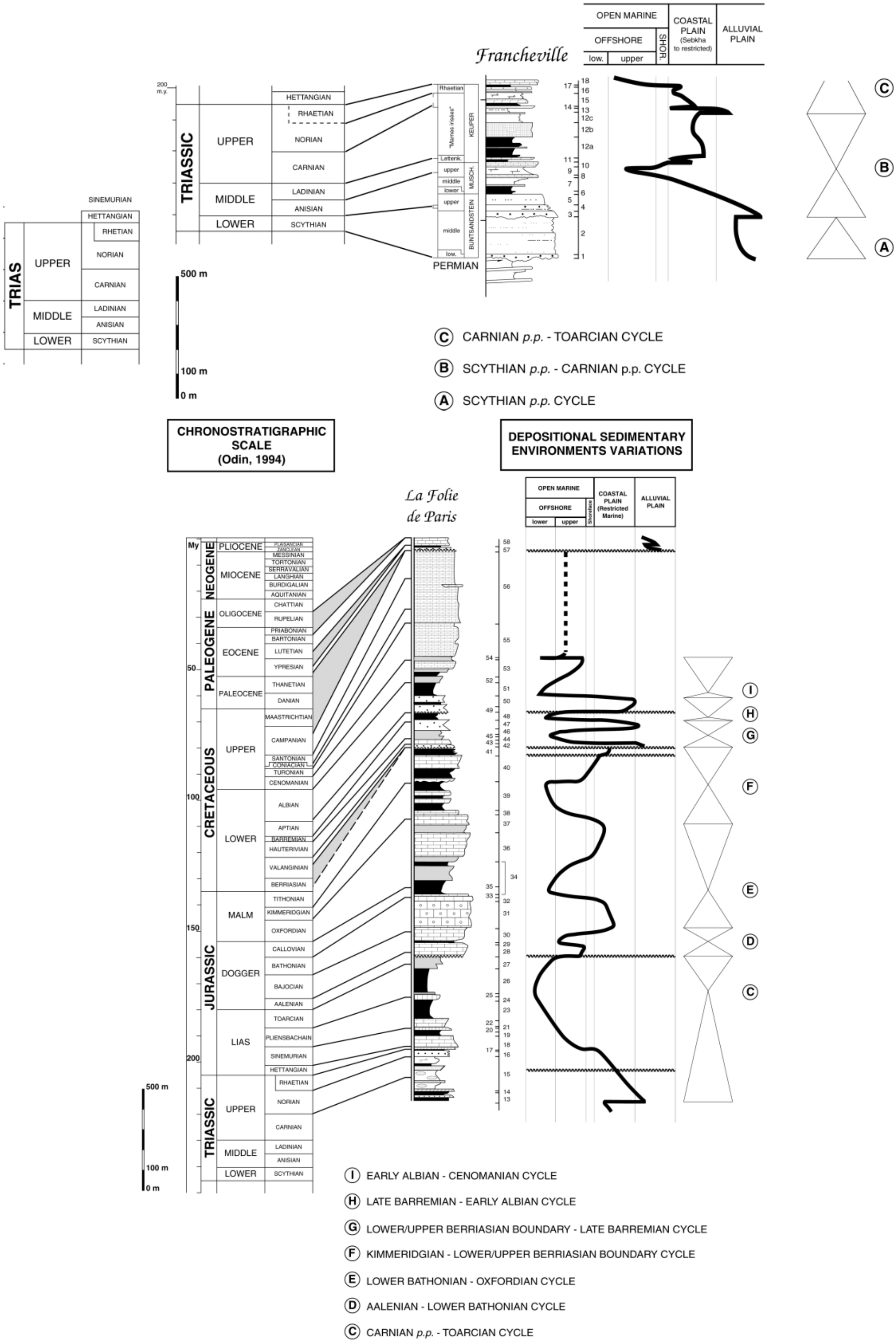
Figure 4. Major faults of the Paris Basin, compiled from Perrodon and Zabek [10] and drilling reports (available at the *Service de Conservation des Gisements d'Hydrocarbures*, French Department of Industry) for subsurface faults, from Colbeaux et al. [42] for the north (Boulonnais, Flandres, Artois, Ardennes), from Le Roux [43, 44] for the east (Lorraine), from B. Pomerol [45] for the south (Nivernais, Puisaye, Pays d'Othe), from Sapin [46], Debeglia and Debrand-Passard [47], Debrand-Passard and Gros [48], Rasplus [49], Belliard-Sens [50], Gély et al. [51], Lorenz [52] for the southeast (Sologne, Berry, Touraine) and from Mary and Giordano [53] for the west (Maine, Perche). No magnetic or gravimetric data have been used. The real existence and the importance of some faults are still debated (more especially for the southeast).

3.1. Major unconformities

Major unconformities *(figure 5)* are here defined as large-scale erosional surfaces (truncations more than 10 km wide) contemporaneous with tectonic deformation of the underlying sediments (independent of their wavelength and amplitude). The major unconformities, except the intra-Norian one, have been known for a long time (see [60]): they have been considered as evidence of epeirogenic activity, but their significance and the amplitude of the associated tectonic movements have never been evaluated.

At least 11 major unconformities have been identified:

- intra-Norian [61, 24, 62];
- Aalenian [63–65];
- Early Cretaceous unconformities (Jurassic/Cretaceous and Lower/upper Berriasian boundaries, [66]);
- late Aptian;
- Late Cretaceous (near Cretaceous/Tertiary boundary, [67]),
- intra-Tertiary unconformities (base Lutetian, [68]; late Bartonian, [69, 70]; intra-Lower Oligocene; base Burdigalian, [71]; end Miocene; Middle Pleistocene, [72]).



Francheville
OPEN MARINE
OFFSHORE
low.
upper
SHOR.
COASTAL PLAIN (Sebkha to restricted)
ALLUVIAL PLAIN
TRIAS
UPPER
MIDDLE
LOWER
SINEMURIAN
HETTANGIAN
RHETIAN
NORIAN
CARNIAN
LADINIAN
ANISIAN
SCYTHIAN
TRIASSIC
RHAETIAN
PERMIAN
KEUPER
MUSCH
BUNTSANDSTEIN
500 m
100 m
0 m
C CARNIAN p.p. - TOARCIAN CYCLE
B SCYTHIAN p.p. - CARNIAN p.p. CYCLE
A SCYTHIAN p.p. CYCLE
CHRONOSTRATIGRAPHIC SCALE (Odin, 1994)
DEPOSITIONAL SEDIMENTARY ENVIRONMENTS VARIATIONS
La Folie de Paris
OPEN MARINE
OFFSHORE
lower
upper
Shoreface
COASTAL PLAIN (Restricted Marine)
ALLUVIAL PLAIN
My
NEOGENE
PALEOGENE
CRETACEOUS
JURASSIC
PLIOCENE
MIOCENE
OLIGOCENE
EOCENE
PALEOCENE
UPPER
LOWER
MALM
DOGGER
LIAS
PLAISANCIAN
ZANCLEAN
MESSINIAN
TORTONIAN
SERRAVALIAN
LANGHIAN
BURDIGALIAN
AQUITANIAN
CHATTIAN
RUPELIAN
PRIABONIAN
BARTONIAN
LUTETIAN
YPRESIAN
THANETIAN
DANIAN
MAASTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
ALBIAN
APTIAN
BARREMIAN
HAUTERIVIAN
VALANGINIAN
BERRIASIAN
TITHONIAN
KIMMERIDGIAN
OXFORDIAN
CALLOVIAN
BATHONIAN
BAJOCIAN
AALENIAN
TOARCIAN
PLIENSBACHIAN
SINEMURIAN
HETTANGIAN
I EARLY ALBIAN - CENOMANIAN CYCLE
H LATE BARREMIAN - EARLY ALBIAN CYCLE
G LOWER/UPPER BERRIASIAN BOUNDARY - LATE BARREMIAN CYCLE
F KIMMERIDGIAN - LOWER/UPPER BERRIASIAN BOUNDARY CYCLE
E LOWER BATHONIAN - OXFORDIAN CYCLE
D AALENIAN - LOWER BATHONIAN CYCLE
C CARNIAN p.p. - TOARCIAN CYCLE

3.2. Major cycles

Nine major cycles *(figures 5a, b and 6)* have been identified from the Triassic (Scythian) to the early Upper Cretaceous (Cenomanian). Because of the lack of good facies models (and corresponding calibration of the well-logs) in the Upper Cretaceous chalk, no cycles have been defined for the Upper Cretaceous times. Despite a low accumulation rate, three Tertiary cycles could be defined. The unconformities described above do not necessarily correspond to the boundary of these major stratigraphic cycles.

The **Scythian *p.p.* cycle** (*Couches d'Annweiller – Grès Vosgien*) is made up of alluvial sediments (braided to straight channels) deposited during a time of global base-level rise ('transgression', [73, 74]).

The base of the **Scythian p.p.–Carnian p.p. cycle** (*Conglomérat principal – Marnes irisées inférieures*) is a major unconformity with local erosion (alluvial-channel scale – *Conglomérat principal* unconformity). In the eastern part of the Paris Basin (Lorraine), the transgressive succession is composed of alluvial sediments (alluvial fans – *Conglomérat principal*, Lower Buntsandstein - to anastomosing rivers **–** *Grès à Voltzia*, Upper Buntsandstein, Lower to Middle Anisian), overlain by coastal plain deposits (*Couches rouges*, *grises*, *blanches*, Middle Muschelkalk, Upper Anisian) and littoral (*Calcaire à entroques*, Upper Muschelkalk, Anisian/Ladinian boundary) to open marine sediments. The maximum flooding surface is of Ladinian age (top of the *Calcaire à cératites*, Upper Muschelkalk). The regressive trend can be subdivided into two phases: (1) a sharp transition from littoral facies (*Calcaire à térébratules*, uppermost Muschelkalk, Ladinian) to coastal/alluvial plain sediments (anastomosing rivers, Lettenkohle - lowermost Keuper, Upper Ladinian) and (2) a phase of vertical aggradation of evaporitic (halitic) coastal plain deposits (*Marnes irisées inférieures*, Lower Keuper, Carnian, [61, 24, 62]). The base of the Lettenkohle coincides with a major downward shift of facies (Lettenkohle unconformity).

The **Carnian p.p.** (*Grès à roseaux*) **– Toarcian cycle** is strongly asymmetric, dominated by the transgressive trend.

The transgressive succession starts with the *Grès à roseaux*, siliciclastic deposits of anastomosing rivers (Middle Keuper, Middle Carnian), followed by lacustrine to coastal plain (evaporitic, dolomitic) deposits of Carnian to Rhaetian age (*Marnes irisées supérieures* and Rhaetian sandstones). The evaporitic deposits grade westward into braided alluvial fans and lacustrine deposits of the Chaunoy sandstones [25, 75]. The dolomitic deposits of the uppermost *Marnes irisées supérieures* are overlain by the protected marine Rhaetian sandstones and open marine facies of the Lower, Middle and basal Upper Lias (alternations of clays and marls with muddy bioclastic limestones, *Calcaire à gryphées*, Hettangian/Lower Sinemurian to *Schistes carton*, Lower Toarcian). The boundary with the underlying cycle (base of the *Grès à Roseaux*) is again a major unconformity with local erosion (channel or valley-scale). The Rhaetian/Hettangian boundary is a major deepening event. The intra-Norian truncative unconformity (intra *Marnes irisées supérieures*) corresponds, in the central and eastern part of the basin, to a slight facies change [62].

The maximum flooding surface corresponds to the top of the black shales known as the *Schistes carton* (paper schist), i.e. the Lower/Middle Toarcian boundary. These *Schistes carton* were deposited below storm wave-base and are the main petroleum source rock of the Paris Basin [23, 76].

The regressive succession is documented by open marine deposits located below storm wave-base (*Marnes à Bifrons et à Voltzi*), overlain by shallower marine deposits and protected marine sediments (bay facies according to palynologic data, B. Courtinat, pers. comm. – *Grés supraliassiques* s.l.).

The age of the Aalenian truncative unconformity is poorly known: in some places (south and southwest of the basin), the associated hiatus ranges from the base of the Upper Toarcian to the base of the Middle Bajocian (Humphriesianum

Figure 5a. Triassic stratigraphy of the eastern part of the Paris Basin represented by the well 'Francheville'; **5b.** Meso-Cenozoic stratigraphy of the Paris Basin represented by the well 'La Folie de Paris': chronostratigraphy, sedimentary environments, stratigraphic cyles (see location on *figure 1a*, early Tertiary-Danian is missing here).
59: *Calcaire de Provins* – **57**: *Argiles de Provins* – **56**: *Craie blanche* (White Chalk) – **55**: *Craie argileuse* (Clayey Chalk) – **54**: *Marnes à Actinocamax plenus* – **53**: *Craie glauconieuse* (Glauconitic Chalk) – 52: *Gaize "cénomanienne"* – **51**: *Argiles du Gault* (Gault Clay), *Marnes de Brienne* – **50**: *Argiles de l'Armance*, *Sables des Drillons*, *Sables de Frécambault* – **49**: *Sables verts* (Lower and Upper Greensands) – **48**: *Argiles à plicatules* – **47**: *Sables et argiles bariolées* (including at the base the *Sables de Congy*) – **46**: *Argiles ostréennes* – **45**: 'Wealden' - *Sables de Perthes* – **44**: 'Wealden' - *Sables de Chateau-Landon* and de *Chateaurenard* – **43**: 'Wealden' - *Calcaire à spatangues* – **42**: 'Wealden' - *Sables de Griselles* – **41**: Purbeckian facies – **40**: Portlandian facies (including *Oolithe de Bure*) – **39**: *Marnes à nanogyres* – **38**: *Calcaire rocailleux à ptérocères* – **37**: Sequanian facies – **36**: Argovo-Rauracian facies (uppermost part, time-equivalent of the *Sables de Glos/ Hennequeville*) – **35**: RIO: Repère Inférieur Oolithique – **34**: Argiles de la Woëvre or Argiles de Massingy – **33**: Dalle nacrée – **32**: *Calcaire de Comblanchien* (time-equivalent of the *Calcaires de Langrune* and *de Marolles-en-Hurepoix*) – **31**: *Oolithe blanche* – **30**: *Grande oolithe* – **29**: *Marnes à* Ostrea acuminata **28**: *Calcaires à entroques* and *à polypiers* (time-equivalent of the eastern Oolithe de Bayeux) – **27**: *Grès supraliasiques* s.l. – **26**: *Marnes à* H. Bifrons and A. Voltzi – **25**: *Schistes carton* (paperschist) – **24**: *Banc de Roc* – **23**: *Argiles à amalthées* – **22**: *Calcaire à* P. Davoei – **21**: *Marnes à* Z. Numismalis **20**: *Calcaire ocreux* – **19**: *Argiles à Promicroceras* – **18**: *Calcaire à gryphées* – **17**: *Argiles Levallois* – **16**: *Grès rhétiens* – **15**: *Marnes irisées supérieures* – **14**: *Dolomie de Beaumont* – **13**: *Grès à Roseaux* – **12**: *Marnes irisées inférieures* (a: *Argiles à anhydrite*, b: *Formation salifère* (Salt), c: *Couches à esthéries*) – **11**: Lettenkohle – **10**: *Calcaire à térébratules* – **9**: *Calcaire à cératites* – **8**: *Calcaire à entroques* – **7**: Middle Muschelkalk (a: red, b: grey and c: white layers) – **6**: Lower Muschelkalk – **5**: *Grès à Voltzia* – **4**: *Couches intermédiaires* – **3**: *Conglomérat principal* –**2**: *Grès vosgien* – **1**: *Grès d'Annweiler*

zone, Thierry et al., in [32]). Where all ammonite zones have been identified (Normandy area for example), the major discontinuity occurs at the Toarcian/Aalenian boundary with condensation of the Opalinum zone (base Aalenian). Other minor discontinuities exist within the Concavum zone (top Aalenian) and at the Aalenian/Bajocian boundary. A Toarcian/Aalenian boundary age is assumed for this unconformity, but in fact, this event ranges from the Toarcian/Aalenian boundary to the latest Aalenian (Concavum zone).

The **Aalenian–Lower Bathonian** (Zigzag zone, Yeovilensis subzone) **cycle** is documented by the evolution of two carbonate platforms, separated by a marly interval. The transgressive succession starts with a shallow marine bioclastic platform (*Calcaire à entroques* s.l.) with some reefs [77]. The maximum flooding surface (Parkinsoni zone, Acris subzone) is represented by open marine alternations of marls-dominated and bioclastic limestones (*Marnes à* Ostrea acuminata). The regressive succession is the beginning of a new carbonate platform, dominated by ooids and bioclastic sands [78 and references therein, 79, 80]. Because of the aggradational geometry of the Bathonian carbonate platform, the flooding surface is not very well expressed and is still controversial. For Jacquin et al. [81], mainly based on 1D data, this change occurred at the Middle/Upper Bathonian boundary. For Gaumet [80], based on 3D data, this flooding surface is of uppermost Lower Bathonian age (Zigzag zone, Yeovilensis subzone).

The **Lower Bathonian** (Zigzag zone, Yeovilensis subzone)–**Oxfordian cycle** marks the end of the highly aggradational Bathonian carbonate platform [29, 79, 80, 82], composed of ooids and bioclastic shoals (*Grande Oolithe*, *Oolithe blanche*) interfingered with carbonate muds deposited in a protected marine environment (*Calcaire de Comblanchien*). The transgressive succession ended with shallow marine bioclastic and ooidic sands (*Dalle Nacrée*). The maximum flooding surface occurred during the Middle Callovian (Jason zone, Jason subzone) during deposition of open marine terrigenous shales. The regressive trend is characterized by the evolution of a new carbonate platform of late Callovian to Oxfordian age, mainly made up of reefs and widespread muddy (chalky) carbonate rocks, known as the Argovian, Rauracian and Sequanian platforms (these terms are facies and not stages), interfingered with coastal plain to shoreline siliciclastic sediments (*Sables de Glos/Hennequeville*, [83, 84]).

The **Kimmeridgian**–**Lower/Upper Berriasian boundary cycle** is composed, during the transgressive succession, of an alternation of organic-rich siliciclastic shales [85–87] with muddy carbonate sediments deposited about storm wave-base. The maximum flooding surface occurred during late Kimmeridgian (*Marnes supérieures à nanogyres* or *à Nanogyra virgula*, Autissiodorensis zone). The regressive trend is represented by muddy and bioclastic limestones going from marine/littoral sediments (‘Portlandian’ facies) to evaporitic coastal plain deposits (‘Purbeckian’ facies). Palaeo-oceanographically, the Kimmeridgian sea looks more restricted (circulation) than the Callovo-Oxfordian one.

The regressive trend ends with two tuncative unconformities [66]: an intra-‘Purbeckian’ one dated at the Jurassic/Cretaceous boundary and a ‘Purbeckian’–‘Wealdian’ one dated at the Lower/Upper Berriasian boundary (intra-Berriasian). Jacquin and Rusciadelli in [66] call the first one the Jurassic/Cretaceous unconformity (JCU=BCU, base Cretaceous unconformity) and the second one the Ryazanian unconformity (RU) and group them as the late Cimmerian unconformity.

The **Lower/Upper Berriasian boundary–late Barremian cycle** initiated a major change in the nature of the sedimentary environments. The Jurassic sediments were deposited on a carbonate platforms with occasional siliciclastic shales deposited in open marine environments, whereas the Lower Cretaceous deposits (Berriasian to Albian) are mainly siliciclastics, with very well developed deltaic (fluvial and wave-dominated) and coastal plain (bay, lagoon) facies.

This cycle started with extensive alluvial sediments (Griselles Sandstones), punctuated at the base by a marine flooding event of Upper Berriasian age [66], overlain by wave-dominated deltaic deposits. They correspond to the ‘Wealden’ facies. The maximum flooding surface occurred during the late Hauterivian [66]. The regressive succession is characterized by fluvial-dominated deltaic deposits. Within these continental deposits, the flooding surface is difficult to identify. It could be of late Barremian or early Aptian age.

The **late Barremian–early Albian cycle** is the shortest cycle (6–9 My), mainly expressed in open marine argillaceous sediments. The maximum flooding surface was of lower Aptian age (*Argiles à plicatules*, Deshayesi zone). The flooding surface occurred on top of the *Sables verts* (Greensands s.s.) which could be early Albian or late Aptian (age of the *Sables verts* in the Albian type-area, [30]).

The age of the late Aptian truncative unconformity is based on outcrops in Normandy: the unconformity is overlain by the *Sables ferrugineux* part of the Jacobi zone (late Aptian, [88, 89]). In the central part of the basin, there was no deposition or erosion during Upper Aptian (Magniez et al., in [32]). This truncative unconformity corresponds to a downward shift of facies from open marine to fluvial and estuarine deposits, just below the flooding surface.

The **early Albian–Cenomanian cycle** records two major changes in the palaeo-oceanography of the basin, and also in the nature of the sedimentary environments. The transgressive succession (Lower/Middle Albian) is mainly composed of siliciclastic tidal-dominated facies —the so-called ‘Greensands’ s.l. facies— overlain, during the regressive succession, by a carbonate platform made up of chalk [90–92]. The beginning of the trangressive trend is strongly aggradational. The maximum flooding surface is located within the open marine siliciclastic shales of late Albian age (Gault clays, Middle/Upper Albian boundary). In the Paris Basin,

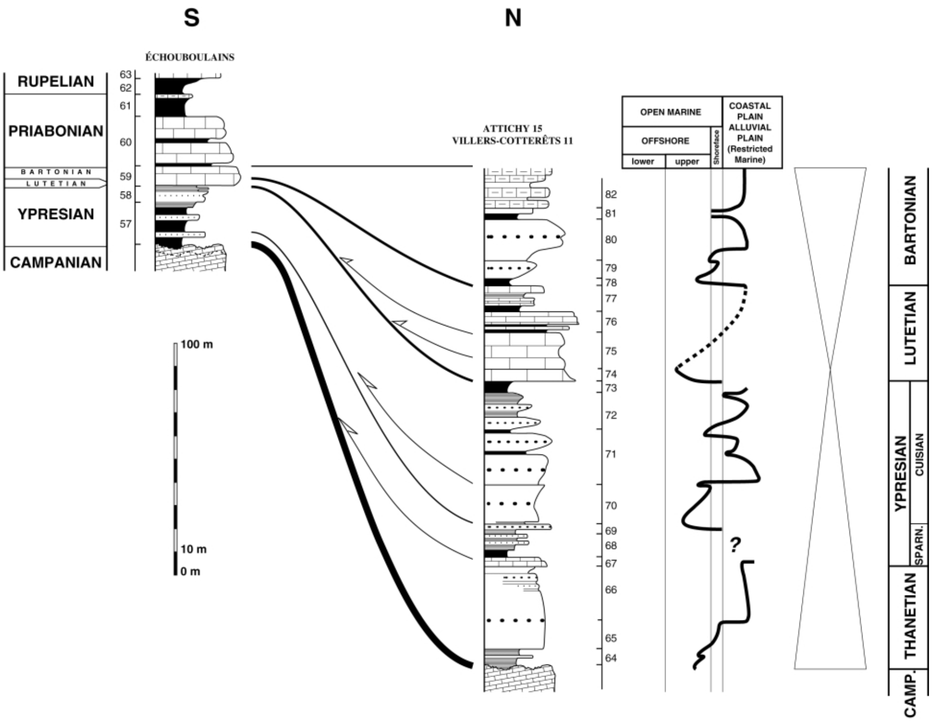


Figure 6. Tertiary stratigraphy of the Paris Basin represented by two wells (core-drills) located near Soissons (north: Attichy 15 and Villers-Cotterêts 11) and Montereau (south: Echouboulains - see location in *figure 1a*).
82: *Calcaire de Saint-Ouen* – **81**: *Horizon d'Ezanville, Calcaire de Ducy, Horizon de Mortefontaine* – **80**: *Sables de Beauchamp* – **79**: *Sables d'Auvers* – **78**: *Argiles de Saint-Gobain* – **77**: *Marnes et caillasses* – **76**: *Calcaire à cérithes* – **75**: *Calcaire à* Ditrupa and *à milioles* – **74**: *Calcaire à* Nummulites laevigatus – **73**: *Argiles de Laon* – **72**: *Sables de Pierrefonds-Cuise* – **71**: *Sables d'Aizy* – **70**: *Sables de Laon* – **69**: *Argiles* and *Sables à cyrènes* – **68**: *Argiles* and *lignites du Soissonnais* – **67**: *Calcaire de Clairoix* – **66**: *Sables de Bourguillemont* – **65**: *Sables de Bracheux* – **64**: *Argiles de Vaux-sous-Laon* – **63**: *Calcaire de Brie* – **62**: *Argiles vertes de Romainville* – **61**: *Marnes supragypseuses* (Argenteuil, Pantin) – **60**: *Calcaire de Champigny* – **59**: *Calcaire de Provins* – **58**: *Grès de Montpothier* – **57**: *Argiles de Provins*

the Cenomanian ends with the deposition of a black shale 'bed' (multiple decimetre-thick layers of 1–2 metres cumulative thickness): the *Actinocamax plenus* Marls (Geslinianum zone). The Cenomanian chalk is called the 'Glauconitic Chalk'.

The **post-Cenomanian–Upper Cretaceous chalk** comprises two main groups of formations [93, 94]: the Clayey chalk of Turonian age, and the White chalk of late Turonian to Senonian (Coniacian to Campanian) age. Maastrichtian chalk is only preserved on both sides of the present-day basin: in the Mons 'Basin' (Hainaut, southern part of Belgium, [95, 96]) and in the Cotentin (Normandy, Armorican domain, [97]). In the central part of the basin, Maastrichtian chalk is only known as reworked pebbles occurring at the base of Tertiary sediments [98] or within palaeoweatherings of Paleocene age (residual clays-with-flints – *Argiles à Silex*, [99]). Our knowledge of the chalk depositional profile and facies is too poor to define transgressive-regressive cycles. Based on palaeontological and geometrical (basement marine flooding) data, Juignet [90, 91] assumed a long-term transgressive trend within the chalk, with a maximum flooding surface during Campanian time.

The Cretaceous/Tertiary truncative unconformity is a major break in the basin evolution with a shift to erosion and continental environments. The open marine chalk facies are overlain by soils and continental deposits grading laterally into marine limestones ('Dano-Montian' facies). Biostratigraphic data [100, 101] suggest a Late Cretaceous age for this discontinuity: in the Champagne area, Upper Maastrichtian chalk pebbles are reworked at the base of the lowermost Danian limestones (Mont-Aimé-Vertus Formation).

The **Tertiary deposits** of the Paris Basin have been extensively studied since the beginning of the nineteen century [102]. Good litho-, bio-stratigraphic and palaeogeographic syntheses have been made (Paleogene: [33, 34, 68]; Neogene: [103]). Three major transgressive-regressive cycles are here defined *(figure 6)*.

(1) The Danian to late Bartonian (Marinesian of the French geologists – base of the *Calcaire de Saint-Ouen* or base of

the *Marnes à pholadomyes*) cycle is composed of six smaller cycles: two Danian cycles (not very well expressed, [100]), one Thanetian cycle (maximum flooding surface: NP6, [104]), one Ypresian cycle (maximum flooding surface: P7/NP11), one Lutetian cycle (maximum flooding surface: P11/NP15) and one Bartonian cycle (maximum flooding surface: end P12/NP16). According to the palaeogeographic maps of Gély and Lorenz [33], the maximum flooding surface of this major cycle occurs within a carbonate platform of Middle Lutetian age (P11/NP15). Uncertainties in the definition of the most landward facies between two lacustrine formations, the *Calcaire de Saint-Ouen* (Upper Bartonian) and the *Calcaire de Champigny* (Priabonian), do not allow us to define the time of the flooding surface. During the late Ypresian a major hiatus occurred (P9-P10 p.p./NP13 p.p.-NP14 p.p.).

(2) The late Bartonian to Late Oligocene/Early Miocene (Aquitanian) cycle is not very well defined on its top: no Upper Oligocene sediments have been recognized in the Paris Basin and Aquitanian lacustrine (*Calcaire de Beauce*) sediments lie paraconformably above Lower Oligocene lacustrine (*Calcaire d'Etampes*) sediments [103]. The maximum flooding surface occurs within the bioclastic limestones (*Faluns*) of Pierrefitte [105]. The passage to continental conditions is very sharp with the deposition of eolian sediments overlying the marine deposits (*Sables de Fontainebleau*).

(3) A Miocene cycle is only recorded in the southwestern part of the basin, along the present-day Loire River (Ligerian domain of the French geologists). It starts by alluvial deposits (*Sables de l'Orleanais* and *de Sologne* p.p., [49]) overlain by bioclastic limestones (the so-called *faluns helvétiens*). The maximum flooding surface could be of late Langhian age [106]. Upper Miocene regressive sediments are poorly known [103].

During the Pliocene and Lower Pleistocene (?) small amounts of mainly continental sediments accumulated in the Paris Basin (*Sables de Sologne* p.p., *de Lozère* and *du Bourbonnais*, [107–109]).

Late Lower/early Middle Pleistocene was the time where the present-day rivers incised: Seine and Somme (800 000 years, [72]), Meuse and Moselle (600 000 years), a process which started at 2.5 Ma [110].

3.3. Controls on the stratigraphic cycles: the quantification of accommodation space

Stratigraphic cycles can result from both variations in accommodation space (A) or/and sediment supply (S). 1D accommodation has been measured in different wells of the Paris Basin *(figure 7)*, [111, 112]). Accommodation can be defined, for a given time interval, by the thickness of the decompacted sediments accumulated corrected for the water-depth (for marine sediments) or altitude (for continental sediments), for more details on depth/altitude corrections see [65, 113]. This procedure is the same as in the construction of a curve by backstripping [114, 115], but, in this study, no eustatic correction has been made, because of the poor knowledge of the quantified amplitude of these variations. The used chronostratigraphic time scale is that of Odin [116]. Accommodation space has been quantified on 88 time intervals. No data are available for the pre-Carnian deposits.

Destruction of accommodation space *(figure 8)* occurred during the early Sinemurian (201–199 Ma), the late Aalenian-early Bajocian (178–174 Ma), the late Bajocian-Lower Bathonian (168–165 Ma - in the eastern part of the basin only), the late Kimmeridgian-Lower/Middle Berriasian boundary time (142–133 Ma - few data), the Upper Aptian-early Albian (111–107 Ma), the late Albian-early Cenomanian (97–94 Ma) and overall from the Late Cretaceous to present-day with some creation of accommodation space during the Ypresian (around 49 Ma) and the Priabonian (around 36 Ma) time. The two periods of maximum space destruction are around the late Kimmeridgian–Lower/Middle Berriasian boundary and during Upper Aptian–early Albian times. The Cretaceous/Tertiary boundary marks a major change in the evolution of accommodation space.

The highest rate of creation of accommodation space *(figure 8)* occurred during the Lower Carnian (229–225 Ma) to early Norian (217.5 Ma), the Rhaetian-Hettangian (207–202 Ma), from the late Sinemurian to the beginning of the Aalenian (195–179 Ma, with a paroxysm beginning in the early Toarcian, 186 Ma, to 179 Ma), from the late Bathonian to the late Kimmeridgian (161–142 Ma, with two maxima: 161–155/156 Ma, Callovian, and 152–142 Ma, Middle Oxfordian to late Kimmeridgian), the late Barremian-Lower Aptian (115–112 Ma), the base Middle Albian (106 Ma, not very important) and the late Turonian-early Santonian (89–86 Ma, few data). The most important period of creation of accommodation space ranges from the Middle Oxfordian to the late Kimmeridgian.

The Aalenian-Lower Bathonian, Kimmeridgian-intra Berriasian, late Barremian-early Albian and late Albian-Cenomanian cycles result from variations in accommodation space with removal of the accommodation during the regressive trend (A<0).

During the Carnian p.p.–Toarcian, Lower Bathonian–Oxfordian and intra Berriasian-late Barremian cycles, with a continuous creation of the accommodation space, the regressive trend was partly due to an increase in sediment supply (A<0, S>A). Granjeon [117], using diffusive stratigraphic models, showed that the intra-Berriasian-late Barremian cycle was controlled by sediment supply. The two other ones, the Carnian p.p.-Toarcian and Lower Bathonian-Oxfordian cycles, may have been controlled by changes in both sediment supply and creation of accommodation space.

These cycles of accommodation space variations can be of tectonic or/and eustatic origin. The importance and nature of the tectonic controls can only be discussed based on 3D geometrical data.

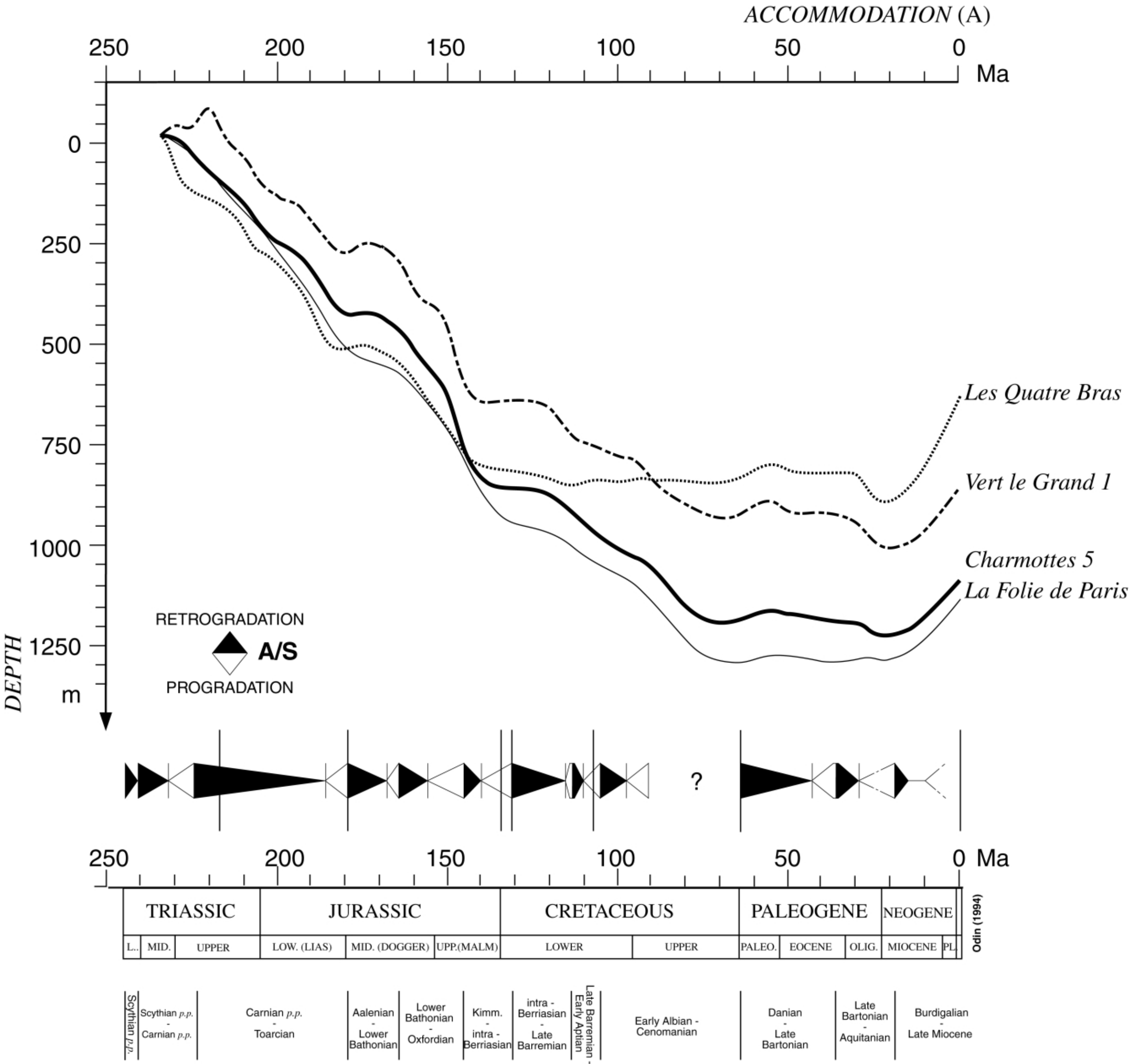


Figure 7. Accommodation curves calculated for the wells ‘Les Quatre Bras’, ‘Vert-le-Grand 1’, ‘Charmottes 5’ and ‘La Folie de Paris’: correlation with the 10–40 My duration stratigraphic cycles.

4. 3D geometry of the sediments

4.1. Scythian p.p. and Scythian p.p.–Carnian p.p. cycles

Because of the alluvial nature of the Scythian sediments, the uncertainties in the well-log correlation are too high to draw time lines across the basin. One isopach map *(figure 9)* has been compiled from the top of the basement, which includes the Permian, to the Ladinian maximum flooding surface (*Calcaire à cératites*), another one *(figure 10)* from the second surface to the base of the *Grès à roseaux* (Carnian p.p.).

The Scythian to late Ladinian sediments (Buntsandstein and Muschelkalk–Scythian p.p. cycle and trangressive trend of the Scythian p.p.–Carnian p.p. cycle, *figure 9*) are preserved along two main subsiding areas, (1) a narrow (100–

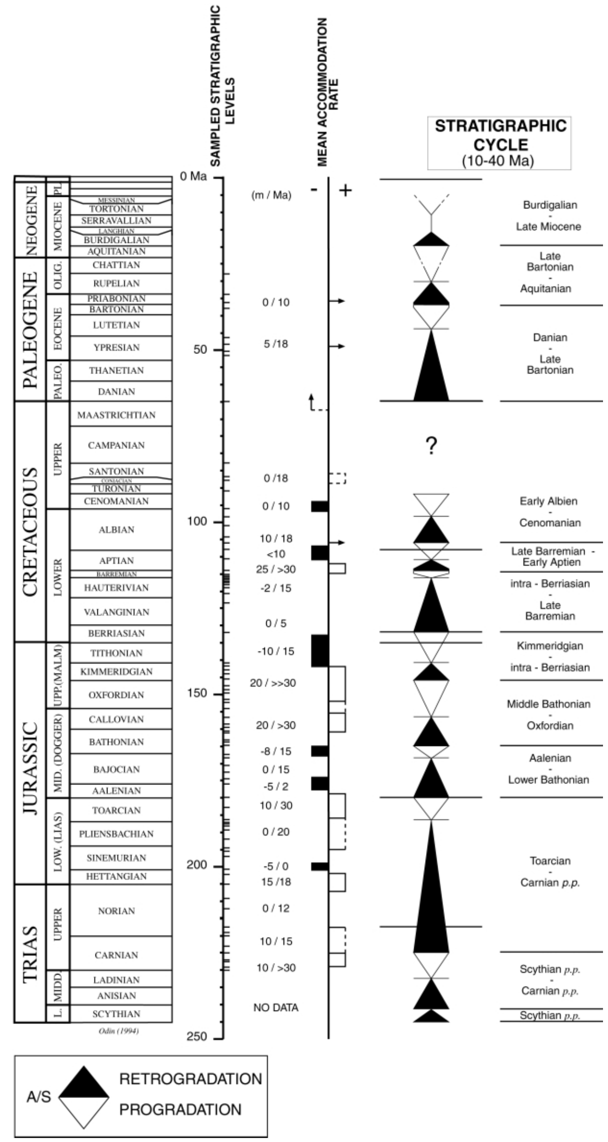


Figure 8. Major 10–40 My duration stratigraphic cycles, unconformities and associated accommodation rate variations (mean values).

120 km wide) NE–SW-trending one, between the Sologne and Champagne areas, and (2) a broad eastern one (Lorraine) which form the western boundary of the German Basin [118, 119]. No sediments have been preserved on the northwestern half part of the present-day Paris Basin. Scythian sediments onlap southward onto the Morvan-Vosges domain, northward onto the Rheno-Hercynian domain and westward onto the so-called *dôme de Songy (figure 9)*. Anisian and Ladinian sediments onlap westward onto the Armorican border. This westward onlap is contemporaneous with a change of the depositional environment going from braided rivers (*Conglomérat principal*, Lower Buntsandstein, Scythian) to coastal plain (Middle Muschelkalk, Anisian) passing westward into alluvial (anastomosing rivers) areas [73], subsequently flooded by the open marine environments of the *Calcaire à cératites*.

During late Ladinian and Lower Carnian times (uppermost Muschelkalk and Lower Keuper, regressive trend of the Scythian p.p.–Carnian p.p. cycle, *figure 10*), the subsiding areas coincide more or less with the early Triassic depocenters. They have the same arcuate shape, with (1) in the west, a NE–SW trend within the same boundaries (the Orléans-Laon line toward the northwest and a line parallel of the Vermenton fault toward the southeast) and (2) in the east, an E–W trend. The main difference is the evolution of a single E–W-trending subsiding area, located north of the Vittel fault, over the previous less subsiding domain (*dôme de Songy*). This period is characterized by a strong relationship between sediment supply (mainly clay and salt) and tectonics. The salt is located in the highest subsiding areas and is mainly aggradational which means similar rates of salt production/clay supply and subsidence. A lateral slight decrease of the subsidence rate led to shaly anhydrite deposition. It is a period of high accommodation space creation (10 to >30 m/My). Westward, medium wavelength tilting is contemporaneous with salt deposition (westward migration of the salt depocenters). N–S-trending short wavelength (around 10 km) folds are coeval with the deposition of the uppermost salt [62].

Control of sedimentation by extensional faults has been assumed for this period [120]. However, our well-log correlations [74, 62, 24] do not agree with the Goggin's interpretation (no Muschelkalk in the southwestern part of the basin although it has been palaeontologically characterized in the well GPF Couy 1 ([121], location on *figure 1a*); wrong interpretation of the top Chaunoy Sandstones unconformity, [62, 25]). The nature of the short wavelength Triassic tectonic controls (Scythian to Rhaetian) is still controversial: which basement fault might have been active? Where there are short wavelength flexures or brittle faults active? Whatever their nature, the horizontal component of displacement along these faults must be very low (no tilted blocks on seismic lines).

4.2. The Carnian p.p. (*Grès à roseaux*)–Toarcian cycle (to the Aalenian unconformity)

This cycle can been subdivided into three intervals: two represented by the transgressive trend, divided by a major deepening event occurring between the Rhaetian and the Hettangian, and one the regressive trend. An isopach map has been compiled for each of them *(figures 11, 12 and 13)*. To better constrain the wavelength and the amplitude of the Aalenian unconformity, an isopach map has been compiled for the Pseudoradiosa/Aalensis zones (late Toarcian, *figure 14*).

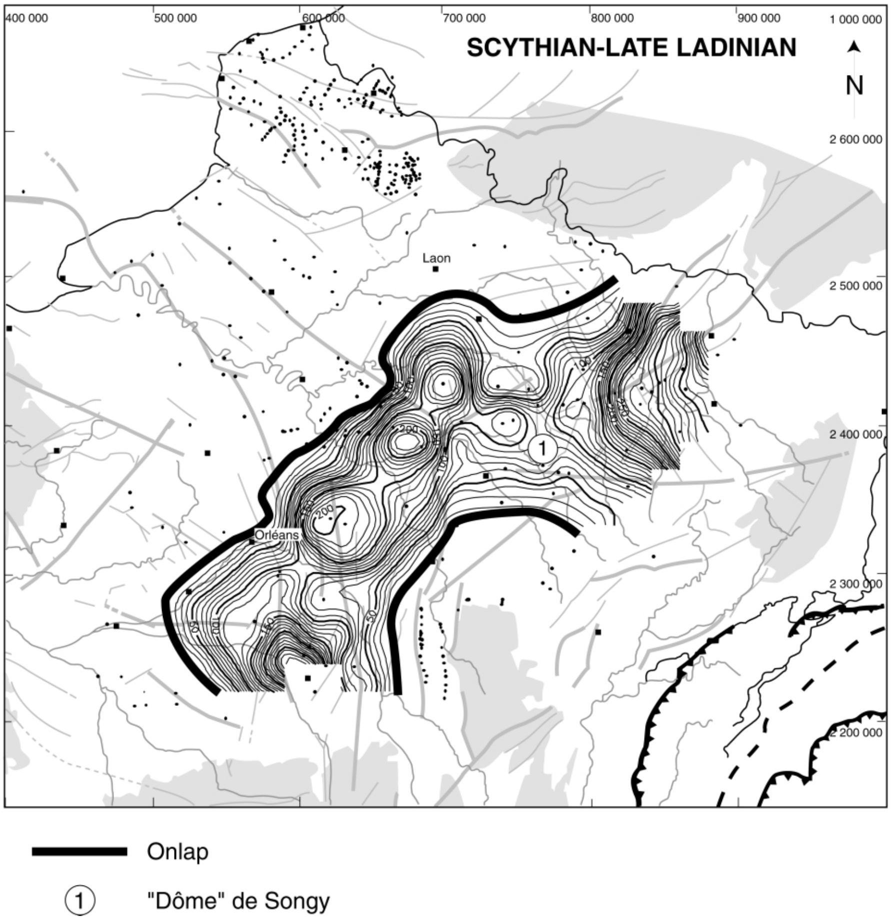


Figure 9. Isopach map from the base-Scythian (top of the basement) to the late Ladinian (maximum flooding surface of the *Calcaire à cératites*): Buntsandstein and Muschelkalk - Scythian p.p. cycle and trangressive trend of the Scythian p.p.-Carnian p.p. cycle.

The Carnian p.p.–Rhaetian isopach map *(figure 11)* suggests a medium wavelength flexural control (a few 100 km) along the same two main directions, ENE–WSW and NNE–SSW, like in the underlying Triassic sediments. The highest thickness gradients are located in the same areas (the Orléans–Laon line toward the northwest and a NW-trending line parallel to the Vermenton fault toward the southeast). The eastward thinning is accentuated by the intra-Norian unconformity, spliting this succession into two parts. In the central part of the basin, this unconformity corresponds to the boundary between evaporitic below and dolomitic coastal plain deposits above. In the eastern part, contemporaneous erosion removed much or all of the underlying deposits, in places down to the *Grès à roseaux* (southeast part). Westward, south of the Bray fault, the intra-Norian unconformity corresponds to a discontinuity at the top of the Chaunoy sandstones [25]. Overall it records a medium wavelength tilting to the WNW with erosion increasing to the southeast [24, 62]. Before this unconformity, braided alluvial fans (Chaunoy Sandstones, [25, 75]), supplied by the Armorican domain and prograding to the east, were bounded westward by the Orléans-Laon line. Post-unconformity deposits are more or less isopachous. These coastal plain deposits flooded the northern and western parts of the basin, onlapping further northwest. The intra-Norian unconformity is contemporaneous with a reorganization of the subsidence pattern. The basal unconformity of this cycle (*Grès à roseaux*) records the beginning of subsidence in the central part of the present-day Paris Basin, which was no longer the western limit of the German Basin [122, 118]. Little is known about the chronostratigraphy and palaeogeography of eastern Normandy Triassic sediments (Cotentin - Courel et al., in [32]); probably, the Portland-Wight Basin was bounded by a high in this area.

The base Hettangian to Lower Toarcian *(figure 12)* was a period of strong, short (multiple of 10 km) and medium (a few 100 km) wavelength tectonic controls [65]. Medium

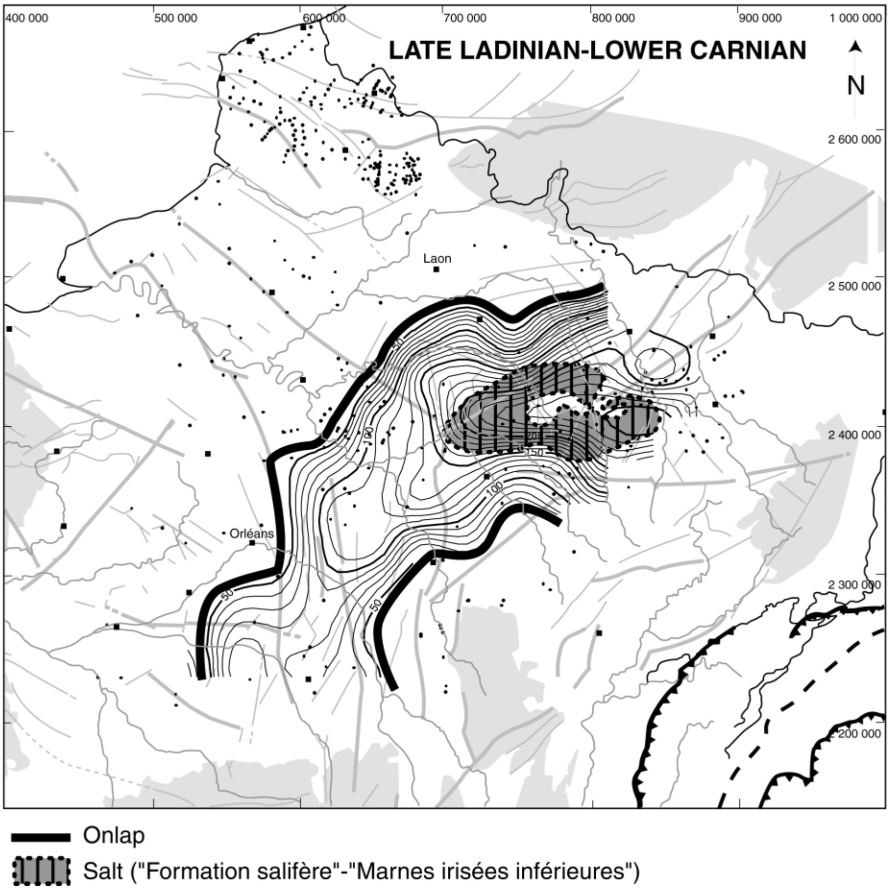


Figure 10. Isopach map from the late Ladinian (maximum flooding surface of the *Calcaire à cératites*) to the Lower Carnian time (base *Grès à roseaux*): late Muschelkalk and Lower Keuper - regressive trend of the Scythian p.p.-Carnian p.p. cycle. Distribution of Lower Keuper (Carnian) salts (from Bourquin et al. [62]).

wavelength control occurred along two main directions, N–S and NE–SW. The most subsident area was located north of the Bray-Bouchy and Luxemburg faults (Champagne/ Argonne areas). Clear short wavelength control existed along N–S faults (Sennely, Saint-Martin-de-Bossenay, [123]): these faults controlled formational thickness but did not alter the open marine depositional profile. The horizontal component of displacement is low, which means a low extension rate at these faults. The depocenters changed with time (Lefavrais-Raymond in [32] and 3D accommodation measurement of Robin [124, 65, 113]): at the Hettangian/Sinemurian boundary, in late Sinemurian time (base Raricostatum zone - lotharingian 'event') and in late Carixian times (Ibex/Davoei zonal boundary).

During the Hettangian and Lower Sinemurian, continental facies occurred toward the west (Sologne) and northwest (Normandy). Upper Sinemurian and Pliensbachian show open marine facies (marl/limestone alternations). The latest Pliensbachian was a time of development of two isolated bioclastic carbonate banks (*Banc de Roc*, *figure 11*) located in the central part of the present-day basin. During Lower Toarcian, the carbonate banks were drowned, they are overlain with a sharp contact by organic-rich shales, (*Schistes carton*) condensed strata deposited below storm wave-base [76, 23, 65].

These Lower and Middle Lias sediments onlap westward onto part of the Armorican domain (Maine, Perche and western Normandy), which was flooded in Pliensbachian and Toarcian times [125]. At the same time, the Anjou and Touraine areas subsided [126]. This was the end of a general onlap history beginning during Scythian time in the eastern part of the Paris Basin. Toward the NNW, the Liassic sediments sharply pinch-out toward the Picardie-Artois domain (southern border of the Dinant synclinorium) and Liassic sediments occur only in the Boulonnais area [127]. Facies analyses of neighbouring wells (Amiens, [128]; Boulogne-

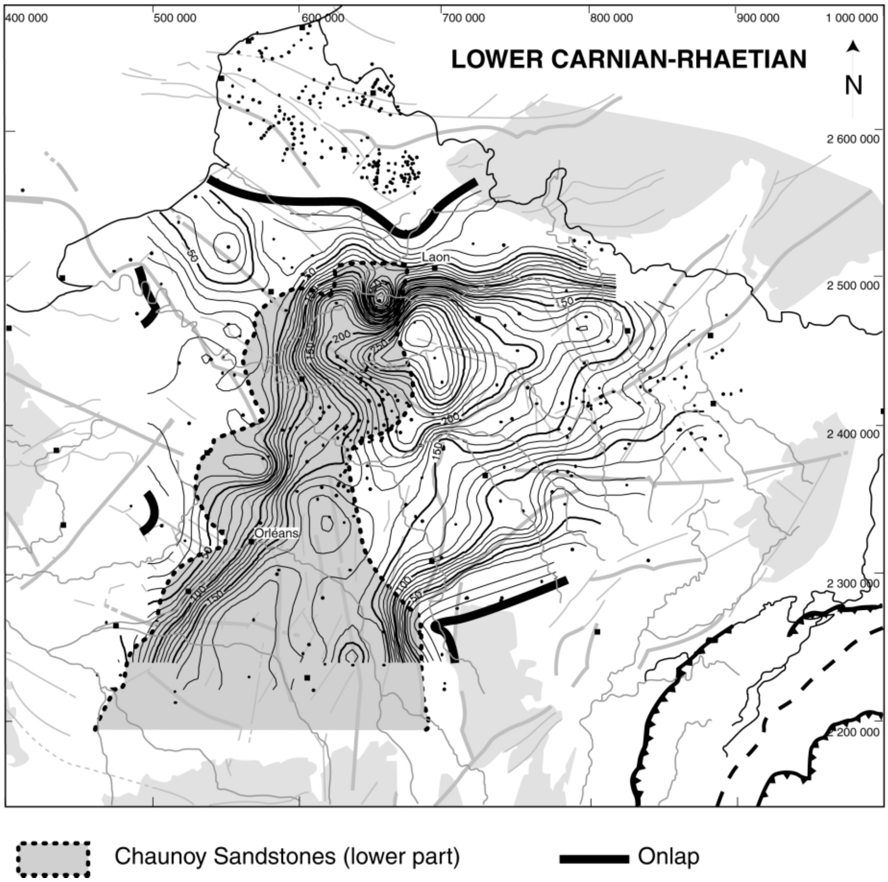


Figure 11. Isopach map from the Lower Carnian time (base *Grès à roseaux*) to the Rhaetian-Hettangian boundary: Middle–Upper Keuper, lower part of the transgressive trend of the Carnian p.p.–Toarcian cycle. Location of the Upper Keuper (Carnian–Norian ?) lacustrine braided fan-deltas of the Chaunoy Sandstones (from Bourquin et al. [62]).

sur-Mer, [129]) indicates open marine conditions. These sharp offlaps could be due to post-depositional tectonically controlled truncation prior to deposition of the Dogger limestones (Aalenian unconformity). Detailed facies and biostratigraphic analyses on both sides of the present-day Variscan basement of the Massif Central (northern part of the Aquitaine Basin and southeastern part of the Paris Basin, [130]) suggest basin continuity across the Massif Central in a N–S direction.

The Middle/Upper Toarcian isopach map *(figure 13)* is partly a residual map because of the truncation by the Aalenian unconformity *(figure 14)*. There is again a change in the wavelength of the tectonic control, with a medium wavelength flexure along a N–S direction across the Saint-Martin-de-Bossenay fault. The western border of the flexure shows the highest formational thickness gradient along the same NNE–SSW direction (Orléans/Laon line) as during the regressive trend of the Scythian p.p.-Carnian p.p. cycle *(figure 10)* and the Carnian p.p.-Rhaetian trangressive half-cycle *(figure 11)*. Higher resolution 3D geometrical data [65] indicate that this flexure started prior to the maximum flooding surface at the time of submersion of the late Pliensbachian carbonate banks, i.e. during deposition of the *Schistes carton* black shales. Lower and Middle Liassic N–S-trending faults are sealed by these Lower Toarcian organic-rich sediments.

The isopach map of the Pseudoradiosa/Aalensis zones (late Toarcian - *figure 14*), bounded above by the Aalenian unconformity, offers a good appraisal of the Aalenian finite deformation. During this time interval, truncation and condensation are limited to the area north of the Bray fault (Picardie) and to the Paris area. Domains of truncated/condensed and preserved accumulations are orientated along N–S to NNE–SSW-trending zones of 200–250 km wavelength, bounded to the southeast by an area of deposition, limited to the north by the Vermenton and the Vittel faults. Preserved depositional areas (synforms) are located in western Normandy, in the Biturige domain (*figure 3*, bounded by the Sennely and

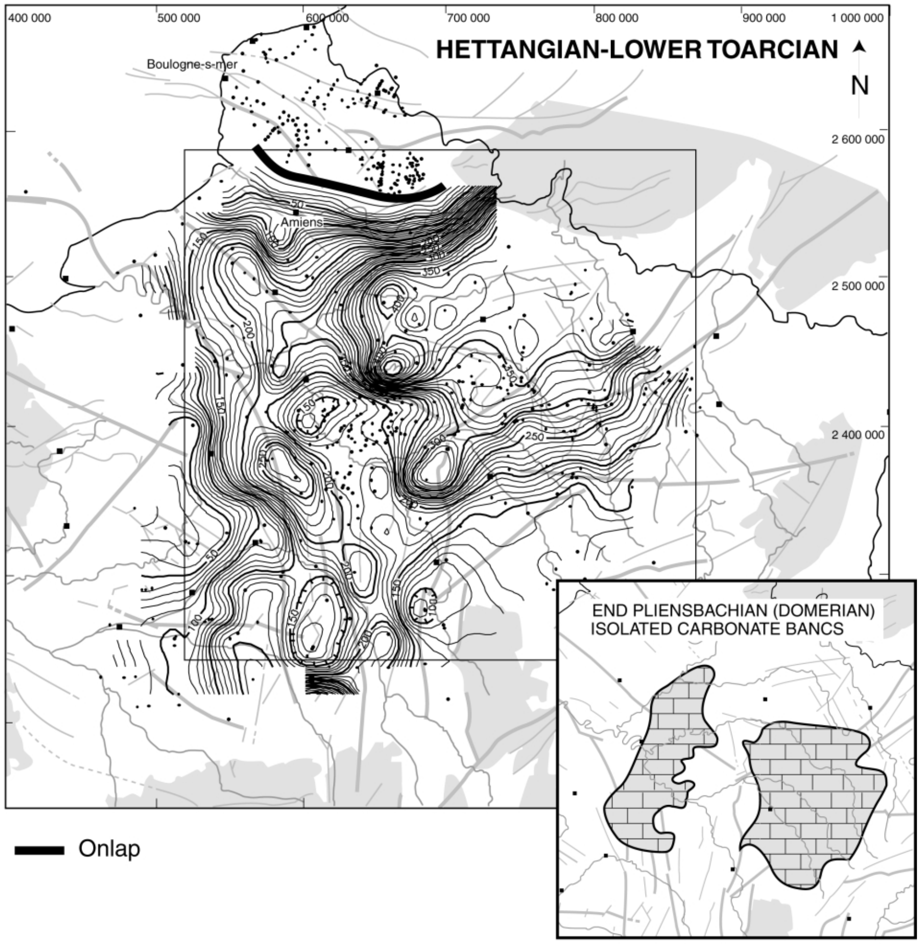


Figure 12. Isopach map from the Rhaetian-Hettangian boundary to the Lower–Middle Toarcian boundary (maximum flooding surface of the top of the *Schistes carton*): upper part of the transgressive trend of the Carnian p.p.–Toarcian cycle. Location of the end Pliensbachian (Domerian) isolated carbonate banks.

Loire faults, [47]) and along a synform located above the Saint-Martin-de-Bossenay fault. Truncation/condensation took place westward of the NNE–SSW Orléans-Laon line and in the eastern Champagne/Argonne areas.

Non-subsident and subsident areas were inverted during Lower to Middle Liassic: Normandy, a domain not subsiding during Lower/Middle Liassic became an area of sedimention; the eastern Champagne/Argonne area, a domain of high subsidence, shows condensation during late Toarcian times.

4.3. The Aalenian-Lower Bathonian (Yeovilensis subzone, Zizag zone) and Lower Bathonian-Oxfordian cycles

Two isopach maps have been compiled from the base Aalenian (Aalenian unconformity) to the Middle Callovian maximum flooding surface (Jason zone, Jason subzone - *figure 15*), and from this surface to the base of the *Calcaire rocailleux à Ptérocères* (Baylei/Cymodoce zones boundary - *figure 18*). This horizon, a key level at the scale of the Paris Basin is located a few metres above the flooding surface of the Lower Bathonian-Oxfordian cycle which is not always easy to identify because of the high vertical aggradation of this carbonate platform.

The base Aalenian to Middle Callovian isopach map (Aalenian-Lower Bathonian cycle and transgressive trend of the Lower Bathonian-Oxfordian cycle, *figure 15*) shows a major change in the location of the subsiding areas. The main directions of the depocentres are no longer NNE–SSW, but NW–SE and to a lesser degree NE–SW. The NW–SE-trending medium wavelength flexure is bounded by the Bray fault (northeast) and the southeastern prolongation of the

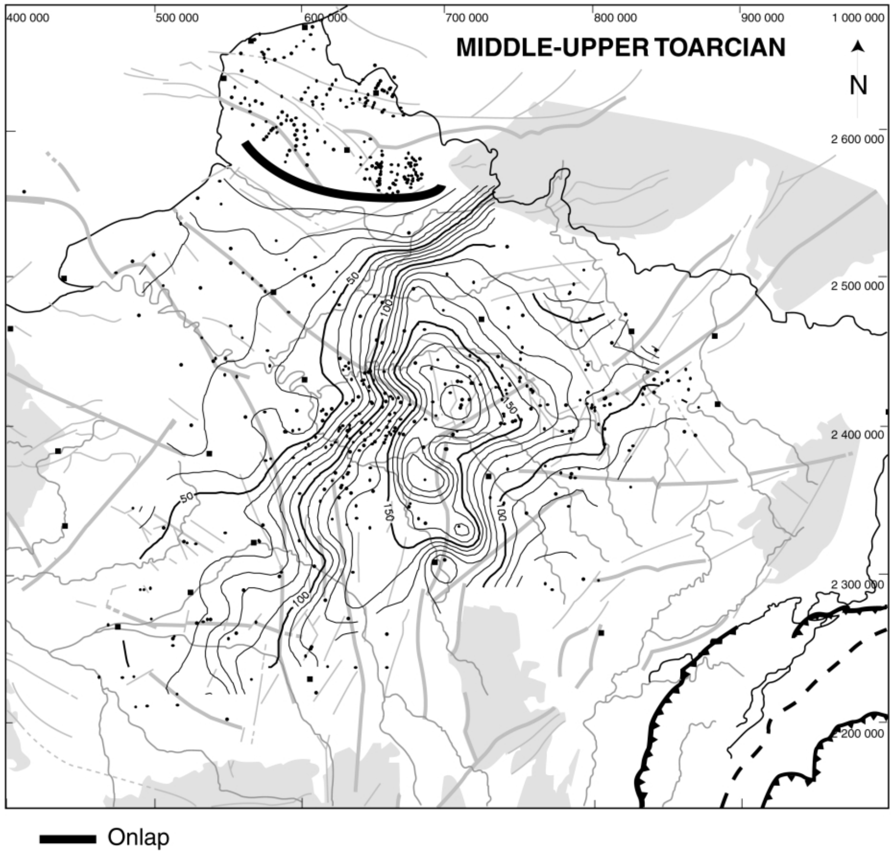


Figure 13. Isopach map from the Lower–Middle Toarcian boundary (maximum flooding surface of the top of the *Schistes carton*) to the Toarcian–Aalenian boundary (Aalenian unconformity): regressive trend of the Carnian p.p.–Toarcian cycle.

Merlerault/Eure faults (southwest). The southeastern part of the basin (Sologne, Berry) is less subsident, split into two major domains along the N–S Sennely fault, with (1) an eastern domain with low accumulation rates located between the Loire and Sennely faults, the Biturige domain [47] and (2) a western more subsiding domain with strata thinning westward. Northward, along the Artois area (southern border of the Dinant synclinorium), sediments pinch-out as a consequence of the Dogger (Upper Bajocian, but mainly Upper Bathonian, [131]) onlap in this area. In the western part, in the Armorican domain (Maine, Perche, western Normandy), Bathonian sediments onlap onto a residual relief (Lower Paleozoic Armorican Quarzite spurs, [132]). The most subsiding area is located at the same place as in Middle/Upper Toarcian times, on both sides of the Bray-Bouchy fault and along the N–S Saint-Martin-de-Bossenay fault.

Along an E–W section between Nancy and Rambouillet *(figure 16)*, the Aalenian–Upper Bajocian (Parkinsoni zone, Acris subzone) transgressive succession displays a complex geometry which could be explained by a shorter wavelength flexural control and/or a variable carbonate production rate along the transect. In the southeastern part of the basin (field data), small synsedimentary faults, sealed by Middle Bajocian sediments (Humphriesianum zone), have been identified [133]. In the southwestern part, NNW–ESE synsedimentary anticlines were active (roll-overs?) at the Middle/Upper Bajocian boundary [134]. 80-m-thick shallow marine sediments in the east (Lorraine area) grade laterally westward into highly condensed deposits only a few decimetres-thick (*Oolithe de Bayeux* in Normandy, [17]).

On a 2D profile *(figure 16)*, the Upper Bajocian (Parkinsoni zone, Acris subzone) to Lower Bathonian (Zizag zone, Yeovilensis subzone) regressive succession reflects again a medium wavelength flexure. At the end of this cycle (Macrescens subzone, Zigzag zone), a first isolated carbonate bank, located between the Seine and Bray faults and north of Paris, shifted to the north of the Bouchy/Vittel faults and to the west of the Saint-Martin-de-Bossenay fault (*figure 17)* - [80]): this could mean a westward medium wavelength tilting.

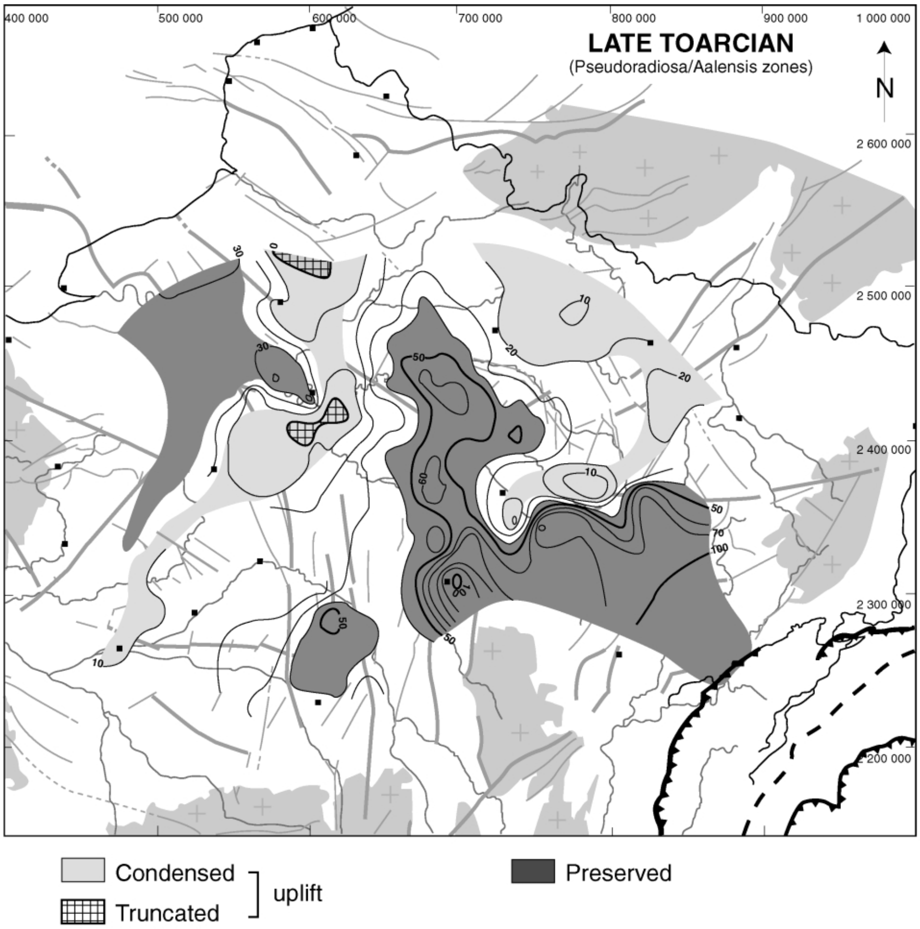


Figure 14. Aalenian unconformity condensations and truncations: isopach map of the Pseudoradiosa/Aalensis zones (late Toarcian) bounded on top by the Aalenian unconformity.

The Lower Bathonian (Zigzag zone, Yeovilensis subzone) to Middle Callovian (Jason zone, Jason subzone) transgressive sediments document the growth and flooding of an isolated platform, the Burgundy bank of Purser [78, 135], bounded on both sides by two "deeper" siliciclastic areas: the *sillon marneux* in the west and the *sillon lorrain* in the east (*figure 17*, [80]). From the Lower Bathonian (Zigzag zone, Yeovilensis subzone) to the base Upper Bathonian (Hodsoni zone), vertical aggradation occurred. During Upper Bathonian (from the Hodsoni zone to the Discus zone, Discus subzone), the isolated platform migrated westward [80], with a progradation in the *sillon marneux* and a retrogradation in the *sillon lorrain*. The cause, carbonate production or tectonism, is still unknown. At the end of this, a forced regression in the *sillon marneux* occurred (*Calcaires de Langrune* and *de Marolles-en-Hurepoix*, [136, 80]). The late Bathonian (Discus subzone) marks the end of the Burgundy bank: detailed measurements of both accommodation and carbonate production [80] suggest a very low accommodation (A) rate, but contemporaneous with a dramatic fall of the carbonate production (S) rate higher than the accommodation rate, leading to a transgression/retrogradation (low A, but A>S). In the southwestern (Berry), western (western Normandy) and northern (Boulonnais) parts of the basin, field evidence [132, 52, 137] suggests short wavelength flexural controls and medium wavelength tilting during the Bajocian-Bathonian with small synsedimentary faults. Their chronology and relationships with the central part of the basin are still poorly understood.

The Middle Callovian (Jason subzone) to early Kimmeridgian (Baylei/Cymodoce zones boundary) isopach map *(figure 18)* shows low spatial subsidence contrasts in comparison with the earlier periods: subsidence is homogeneously distributed over this area. NW–SE structural trends (southeast prolongation of the Merlerault/Eure fault and Bray fault) initiated during the Dogger are still recorded with slight NE–SW influences (NW parallel of the Vermenton fault).

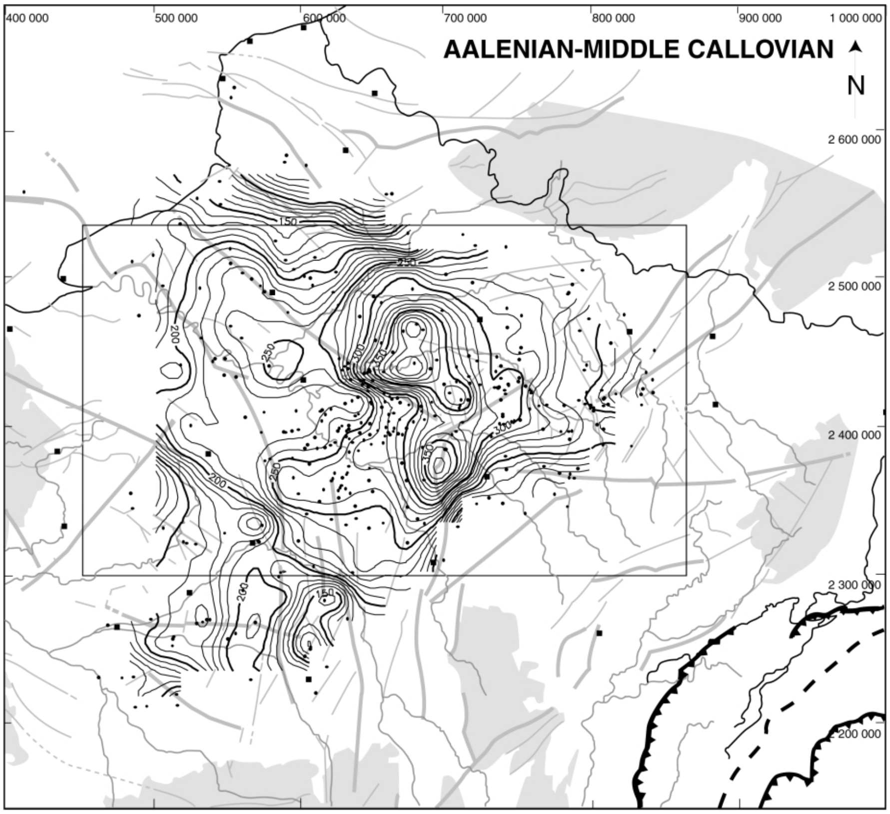


Figure 15. Isopach map from the Toarcian-Aalenian boundary (Aalenian unconformity) to the Middle Callovian (maximum flooding surface of the Jason zone, Jason subzone): Aalenian-Lower Bathonian cycle and transgressive trend of the Lower Bathonian-Oxfordian cycle.

The two most subsident areas are located between the Seine and the Loire faults (Beauce area) and north of the Bray fault (Ile-de-France/Champagne boundary). The Paris basin was occupied during Upper Callovian and Lower/Middle Oxfordian, by a progradational carbonate platform ("Argovo-Rauracian" platforms) with downlaps over a condensed interval of iron oolites (RIO: *Repère Inférieur Oolithique* of the petroleum geologists). 3D palaeogeographic data suggest a progradation towards the south evolving to a more southwest trend from an area located in the present-day English Channel. Condensation occurs in the southeastern to southern part of the basin (Burgundy High, Morvan). Middle to Upper Oxfordian sediments aggraded and the flooding surface is again difficult to localize exactly. This was a time of high rates of accommodation space creation (20/>30 m/My).

4.4. The Kimmeridgian–Lower/Upper Berriasian boundary cycle

Two isopach maps have been compiled, from the base of the *Calcaire rocailleux à ptérocères* (Baylei/Cymodoce zones boundary - *figure 19*) to the *Marnes supérieures à nanogyres* (Autissiodorensis zone) and from this surface to the Lower/Upper Berriasian unconformity (RU - *figure 20*). To better constrain the wavelengths and amplitudes of the early Cretaceous unconformities, isopach maps of the Lower (late Tithonian) and Upper (Lower Berriasian) 'Purbeckian', drawn by Rusciadelli [66], have been used *(figures 21 and 22)*.

The transgressive succession (*Calcaire rocailleux à ptérocères - Marnes supérieures à nanogyres - figure 19*) is a period of regionally homogeneous subsidence enhancing the Middle Callovian–early Kimmeridgian trends with low thickness gradients along two main directions, ENE–WSW (northwest parallel of the Vermenton fault) and NW–SE (southeast prolongation of the Merlerault/Eure fault and the Bray fault). The most subsident areas were oriented along an ENE–WSW axis, between the Beauce and the Champagne areas. The Paris Basin was filled by an alternation of marine, organic-rich clays and limestones of wide regional extension, bounded southeastward by a more carbonate-rich system, transitional to the open ocean (shoals along the transition to the Subalpine Basin - southern Jura and Bresse, Enay in [138]). Northward, toward the London-Brabant

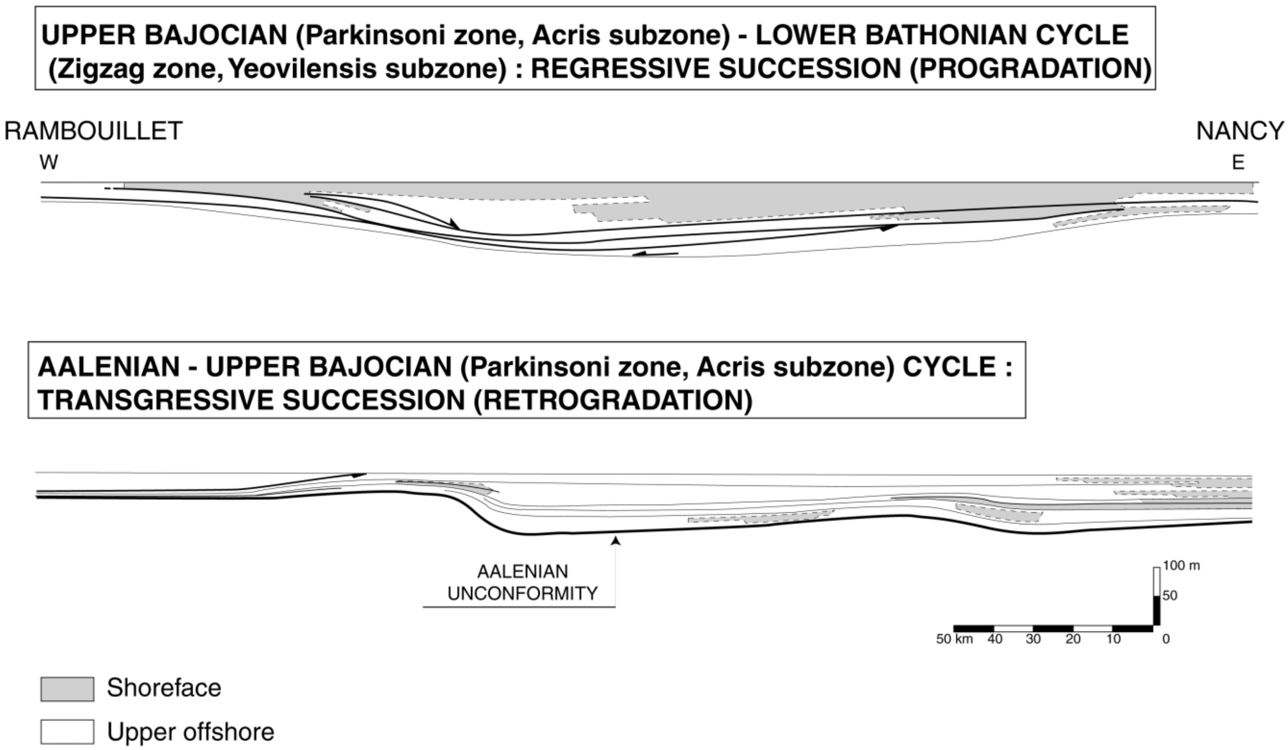


Figure 16. 2D W–E section between Nancy-Rambouillet (see location on *figure 1a*) based on well-log correlation for the Aalenian-Lower Bathonian cycle.

domain, those alternations grade into siliciclastic sandy shoreline deposits (Boulonnais, [139]). The palaeo-oceanography of this domain is still poorly understood, however the Kimmeridgian was time of (1) the maximum accommodation space creation (20/>>30 m/My), (2) the regionally most homogeneous subsidence and (3) the occurrence of one single facies: limestone/organic-rich clay alternations.

The regressive succession (*Marnes supérieures à nanogyres*-Lower/Upper Berriasian unconformity - *figure 20*) isopach map is a residual map because of two truncations (late Cimmerian unconformities), at the Jurassic/Cretaceous boundary (Lower/Upper 'Purbeckian' boundary, JCU=BCU of Rusciadelli [66]) and at the Lower/Upper Berriasian boundary ('Purbeckian'/'Wealdian' boundary, Ryazanian Unconformity, RU, of Rusciadelli [66]). The second one corresponds to a major facies change ('Purbeckian' carbonates *vs* 'Wealden' siliciclastics). 'Portlandian' facies (Tithonian ante 'Purbeckian') are mainly muddy limestones (few bioclasts and ooids - *Oolithe de Bure*) deposited on ramps extending from open marine to restricted marine (bay/lagoon) environments. This was a period of accommodation space removal (–10/+15 m/My).

The Jurassic/Cretaceous (JCU) and Lower/Upper Berriasian (RU) unconformities subcrop maps are difficult to draw because of the late Aptian unconformity that cuts down onto the Lower Berriasian and the Jurassic and therefore older unconformities are difficult to recognize. In the central part of the basin, Lower Berriasian (Upper 'Purbeckian') sediments *(figures 21, 22)* overlie toward the northeast, folded and eroded late Tithonian (Lower 'Purbeckian') deposits: the direction of the medium wavelength fold is NW–SE. Upper Berriasian sediments ('Wealden'- *figures 22, 23*) overlie toward the northwest, late Tithonian (Lower 'Purbeckian') sediments (the Lower Berriasian is missing here): Lower 'Purbeckian' strata are slightly tilted toward the east. In the NNW part of the basin, in outcrop, 'Wealden' facies overlie with an unconformity Variscan basement and Jurassic sediments in the Boulonnais [127, 140, 141] and in the Cambrésis-Thiérache-Mons area (karst filled by lower Cretaceous sediments [142–144]). These sediments document a tilting toward the south to southwest of the underlying Jurassic sediments. Because of the late Aptian erosion, no 'Wealden' sediments were preserved on the Artois (northern) and Armorican (western) domains. The unconformities document (1) a slight eastern tilt of the Armorican domain of Lower/Upper berriasian age (RU) and (2) a southwestern tilt of the Ardennes, with uplift and erosion, mainly occurring at the Jurassic/Cretaceous boundary (JCU).

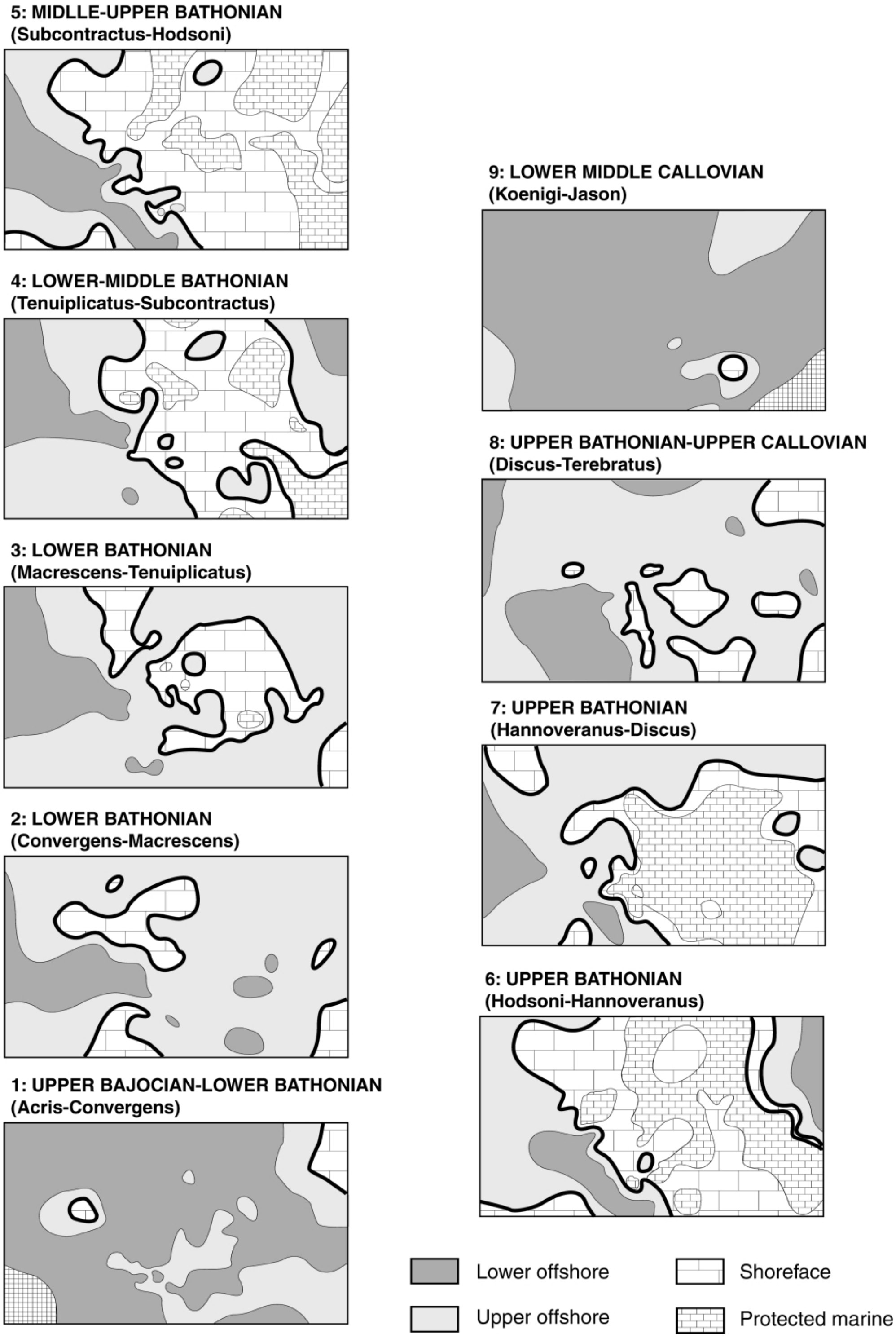


Figure 17. Palaeogeographic maps of the central part of the Paris Basin from the Lower Bajocian (Parkinsoni zone, Acris subzone) to the Middle Callovian (Jason zone, Jason subzone, from Gaumet [80], see location in *figure 15*).

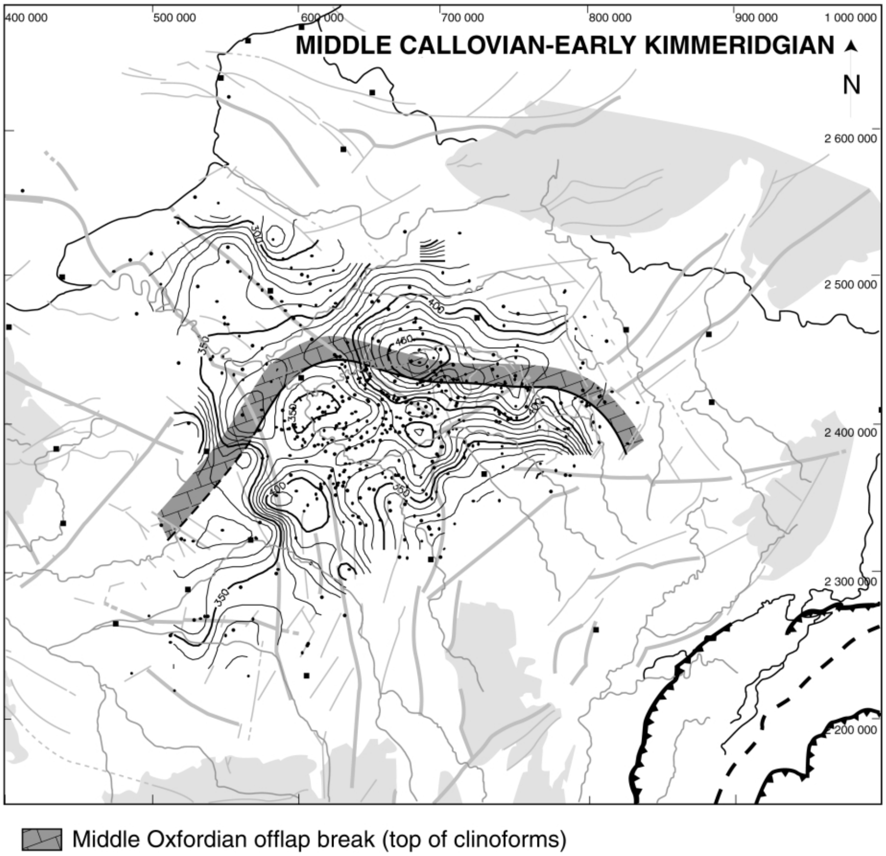


Figure 18. Isopach map from the Middle Callovian (maximum flooding surface of the Jason zone, Jason subzone) to the early Kimmeridgian (base of the *Calcaire rocailleux à ptérocères*, Baylei/Cymodoce zones boundary): regressive trend of the Lower Bathonian-Oxfordian cycle. Location of the offlap break (top of the clinoforms) at Middle Oxfordian boundary.

4.5. The Lower/Upper Berriasian boundary–late Barremian and the late Barremian–early Albian cycles

Because of the alluvial nature of the late Barremian/early Aptian sediments, the flooding surface of the Lower/Upper Berriasian-late Barremian cycle is difficult to identify on well-logs. Two isopach maps have been compiled, one from the Lower/Upper Berriasian boundary unconformity to the Lower Aptian maximum flooding surface (*Argiles à plicatules*, Deshayesi zone - *figure 23*) and one from this surface to the late Aptian unconformity *(figure 24)*.

The Lower/Upper Berriasian unconformity, Lower Aptian (Lower/Upper Berriasian boundary–late Barremian cycle and transgressive trend of the late Barremian–early Albian cycle, *figure 23*) isopach map shows a medium wavelength flexure orientated along a NE–SW axis with still NW–SE flexural directions. There is no evidence of fault control. This was a period of very low creation of accommodation space (0/5 m/My, Upper Berriasian-Valanginian, to –2/+15 m/My, Hauterivian to early Barremian) with an increase in late Barremian/early Aptian times (25/>30 m/My).

'Wealden' facies document wave- (transgressive succession) to fluvial-dominated (regressive succession) deltas in a semi-enclosed sea with a low storm activity, open toward the southeast from the Champagne (east) to the Puisaye (south). The pattern of the delta *(figure 23)* suggests a supply of sediments from the west, i.e. from the Armorican domain. Because of the very low accommodation space available, the regressive succession of Barremian age is characterized by a sharp downward shift of facies. At the beginning (Griselles sandstones) and at the end of this cycle, the Paris Basin was entirely part of the continental domain.

The Aptian (Deshayesi zone)–late Aptian unconformity (part of the regressive trend of the late Barremian-early Albian cycle, *figure 24*) isopach map indicates low thickness variations: the isopachs show a pattern similar, however less pronouced, to that of the underlying 'Wealden'

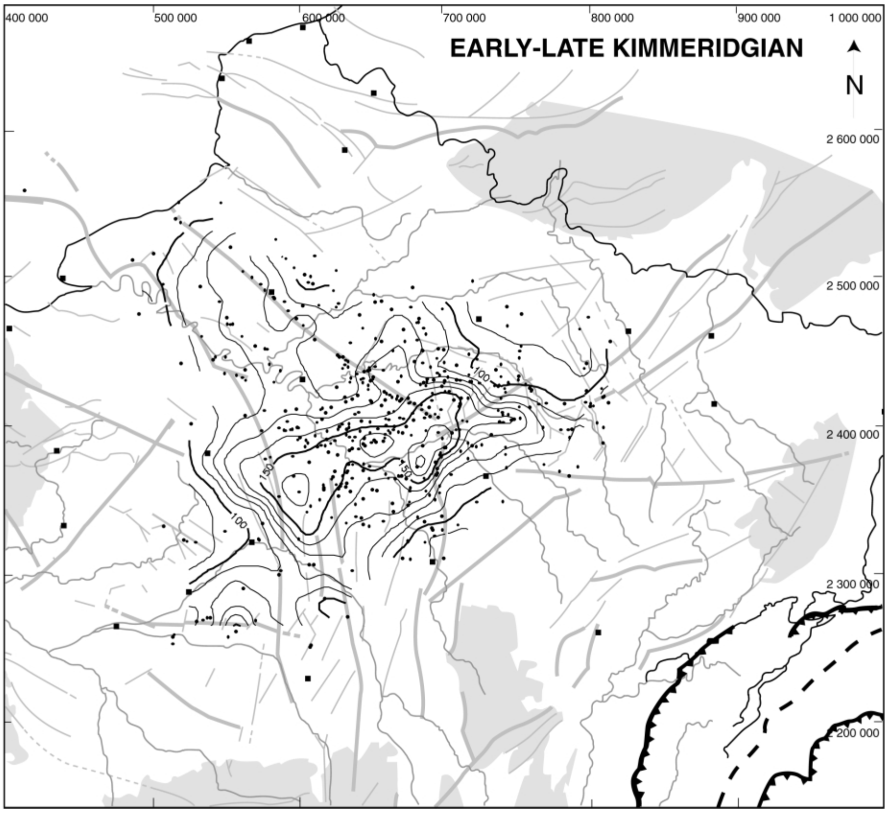


Figure 19. Isopach map from the early Kimmeridgian (base of the *Calcaire rocailleux à ptérocères*, Baylei/Cymodoce zones boundary) to the late Kimmeridgian (maximum flooding surface of the *Marnes supérieures à nanogyres*, Autissiodorensis zone): part of the transgressive trend of the Kimmeridgian-Lower/Upper Berriasian boundary cycle.

deposits with a slight NE–SW flexure. Two thickness anomalies (thinning) are obvious along N–S-trending faults (Villeneuve-sur-Yonne and Saint-Martin-de-Bossenay) along which local erosion occurred at the time of the late Aptian unconformity. The late Aptian was a time of major accommodation space removal (<–10 m/My). Because of the late Aptian unconformity, Aptian sediments were truncated along the Ardennes and Armorican margins of the basin. As a consequence, the Aptian palaeogeography is poorly known.

The late Aptian unconformity subcrop map *(figure 25)* shows in the west (Maine, Touraine, Berry) truncation of the Upper Jurassic to 'Wealden' sediments tilted eastward and in the north (Picardy to Champagne, southern border of the Dinant synclinorium) truncation down to the Variscan basement. In the northern part, the erosional truncation clearly accentuates the effects of the late Cimmerian unconformities. In the Boulonnais area, the Jurassic to 'Wealden' sediments are preserved in a synform with an Albian onlap across the basement in the surrounding areas *(figure 25)*. Vertical movements were low in the Cambrésis, Thièrache and Ardennes areas where 'Wealden' facies are still preserved. Mapping in western Normandy [145] suggests truncations and onlap onto the Lower/Middle Jurassic sediments and down to the Cadomo-Variscan Armorican basement. In the central part of the basin, Albian sediments paraconformably overlie Aptian sediments. The Paris Basin can be described as a synform with NW–SE axial orientation (medium wavelength deformation), with tilting and truncation in the north (southern border of the Dinant synclinorium: Somme, Artois) and in the west along the Armorican basin margin.

4.6. The early Albian–Cenomanian cycle

Two isopach maps have been compiled for this cycle, from the late Aptian unconformity to the maximum flooding surface (Middle/Upper Albian boundary - *figure 26*) and from this surface to the *Actinocamax plenus* Marls of Upper Cenomanian age (Geslinianum zone - *figure 27*).

The late Aptian-Middle/Upper Albian map (upper part of the regressive trend of the late Barremian-early Albian cycle and transgressive trend of the early Albian–Cenomanian cycle; *figure 26*) shows a general onlap along the western (Armorican) and northeastern (Artois, southern border of the

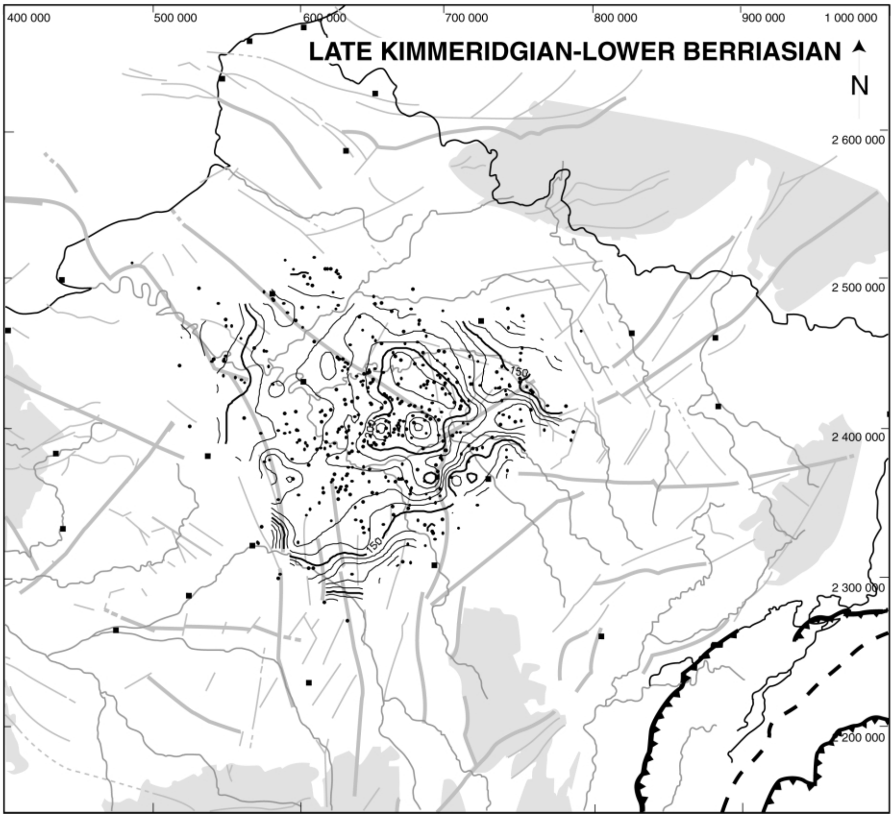


Figure 20. Isopach map from the late Kimmeridgian (maximum flooding surface of the *Marnes supérieures à nanogyres*, Autissiodorensis zone) to the Lower/Upper Berriasian unconformity (Ryzanian Unconformity, RU): regressive trend of the Kimmeridgian-Lower/Upper Berriasian boundary cycle.

Dinant synclinorium) margins of the basin, overstepping tilted and eroded zones above the late Aptian unconformity. The overlap of the onlap is less limited onto the Armorican margin. The Artois area is more widely flooded by the Greensands s.s. (late Aptian?–Lower Albian) and the Gault Clay (Middle/Upper Albian). Over this area, the Greensand s.s. occurences are patchy which can be interpreted as due to an irregular topography or a local low amplitude flexural control. The onlap starts in the NNW part of the basin with the deposition of the late Aptian *Sables ferrugineux*.

The late Aptian unconformity is coeval with a major reorganization of subsidence. There was still a medium wavelength flexural regime trending NW–SE (the same as in Middle and Upper Jurassic), but more widespread (including the southeastern part of the basin in Berry and Sologne) and with no southeastern decrease of the subsidence toward the Morvan. A larger subsiding domain in the southeast might be expected. The NW–SE flexure is bounded by the Bray and Somme faults, which means that the areas of subsidence shifted with respect to the Middle and Upper Jurassic times. In the central part of the basin, the most subsiding area is still oriented along a NE–SW direction, similar to the 'Wealden' trend. Again a thinning of sediments is noticed along the N–S Saint-Martin-de-Bossenay fault where local erosion occurs during late Albian (*Argiles de l'Armance*/*Sables des Drillons* boundary), with erosion of the Greensands s.s.

Lower to Middle Albian siliciclastic sandy facies (continental to deltaic - fluvial- to tide-dominated facies: Greensand s.l.) are located in the central to southwestern part of the basin. In the French lithostratigraphic terminology, the *Sables verts* (Greensand) are not limited to marine to estuarine glauconitic-rich sands, they comprise continental deposits. They pass laterally toward the north and the WSW into open marine facies (Gault Clay, [146]). The distribution of the sandy facies is more or less controlled by the St-Martin-de-Bossenay and Bray faults *(figure 26)*.

The regressive succession (upper part of the Gault Clay and Cenomanian 'Glauconitic chalk', *figure 27*) can be subdivided into two parts separated by a discontinuity at the Albian/Cenomanian boundary. Cenomanian deposits overlie the basement and tilted truncated Jurassic sediments along sharp contact on both northern (Flandres-Brabant) and west-

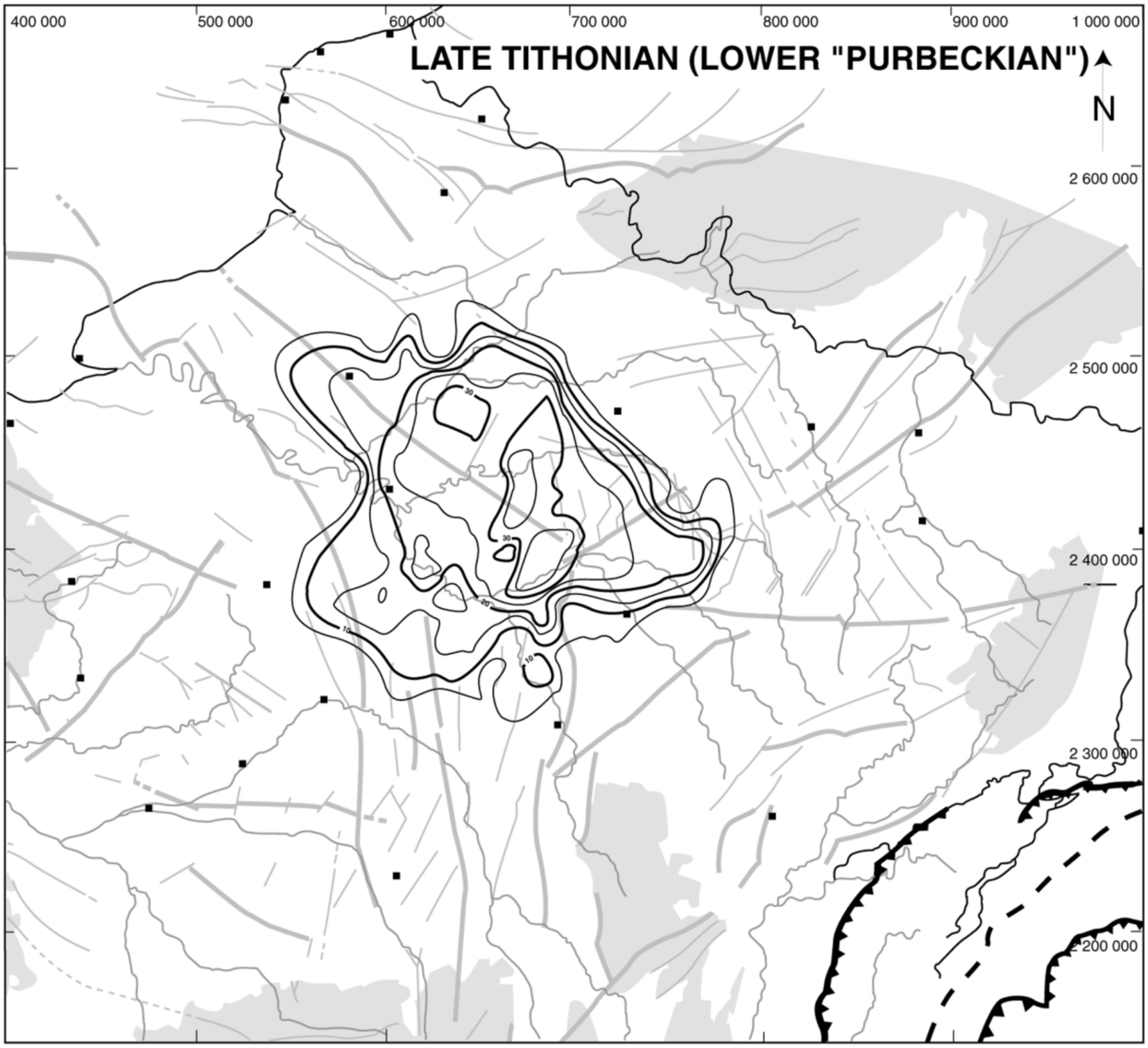


Figure 21. Isopach map from the late Tithonian (Lower 'Purbeckian') bounded on top by the Jurassic/Cretaceous boundary unconformity (JCU, from Rusciadelli [66]).

ern (Armorican) basin sides. The Armorican flooding is more important; it occurred during Lower and Middle Cenomanian time (Vendée: [147, 91]; Le Mans area: [148]; south Cotentin: Lautridou, pers. comm.). This Cenomanian overlap of the Armorican basement is well-preserved south of the Merlerault/Eure fault (south Perche, Maine, ...). To the northwest, the Cenomanian onlap does not cross the southwestern part of the Lille area *(figure 27)*.

This is still a flexural control along a NW–SE trend (between the Bray and Somme faults), the wavelength, however, is larger with low thickness variations indicating that subsidence became homogeneous. SW–NE flexures are still slightly active. Three main subsidence areas can be defined: a southwestern area (Maine, Beauce), a central area (Champagne) and a northern domain (Boulonnais, Flandres). The isopach pattern suggests a subsiding area wider than the present-day area of preserved Upper Cretaceous sediments and spreading out toward the SSE.

Geometrically *(figure 28)*, the regressive trend is first aggradational (Upper Albian Gault Clay) and then highly progradational with well expressed clinoforms and downlaps (Cenomanian chalk). The transition between the marls and the chalk is diachronous. The Cenomanian sediments are the products of two different sediment factories *(figure 27)*: a siliciclastic progradational deltaic system located in the Maine/Touraine areas, eastward bounded by the Sennely/Seine faults, and an isolated prograding carbonate platform initiated over the Ile-de-France (Paris) area and migrating

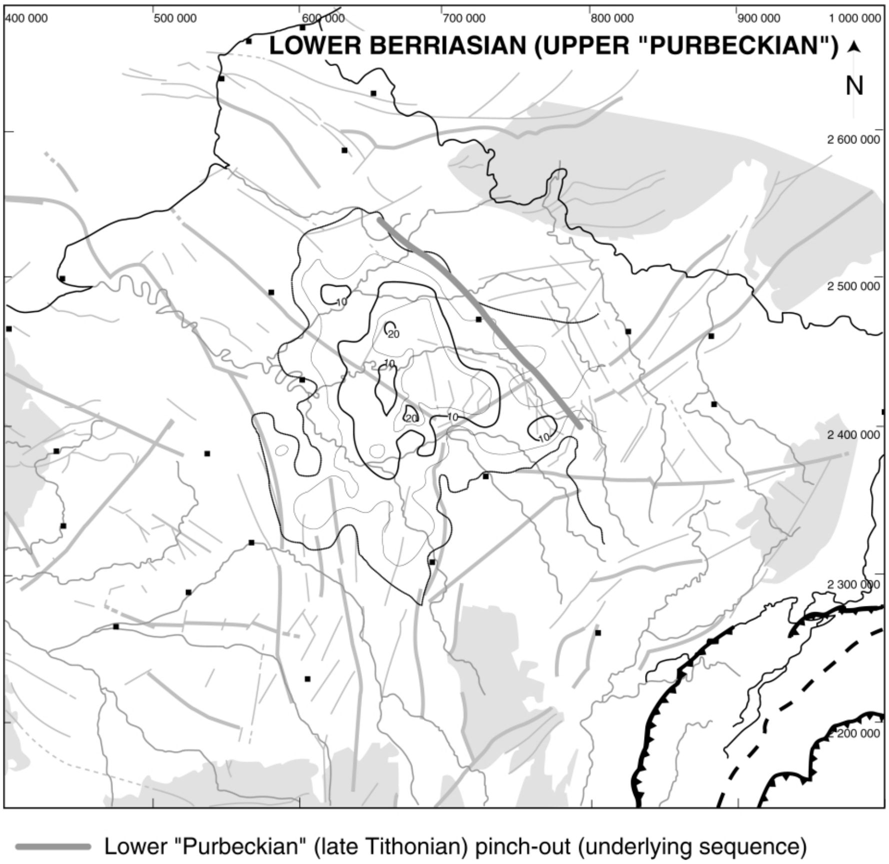


Figure 22. Isopach map from the Lower Berriasian (Upper 'Purbeckian') bounded at the base by the Jurassic/Cretaceous boundary unconformity (JCU) and above by the Lower/Upper Berriasian unconformity (Ryzanian Unconformity, RU, from Rusciadelli [66]), westward pinch-out (truncation) of the underlying Lower 'Purbeckian' sediments.

toward the southeast. The siliciclastic deltaic facies are located in the southwestern subsiding area (Maine, Beauce), where the Cenomanian onlap onto the Armorican basement is preserved.

4.7. Post-Cenomanian Upper Cretaceous chalks

The post-Cenomanian chalk to the Cretaceous/Tertiary boundary isopach map *(figure 29)*, has been compiled by Hanot [149] for the central part of the basin (Ile-de-France, Brie, Beauce). Geometrical data are only available along an W–E 2D transect between Saint-Dizier and Rambouillet (southwest of Paris - *figure 28* and location on *figure 1a*).

The Turonian Clayey Chalk is made up of chalk with some clay levels (several decimetres to metres-thick) which can be traced over the Paris and London Basins, with few lithological and thickness variations [93, 94]. From 2D data *(figure 28)*, a medium wavelength flexural regime can be deduced, accentuated by a general onlap of the Lower Turonian onto the *Actinocamax plenus* Marls in the WSW part of the basin (southern Beauce/northern Sologne areas). Laterally, the Clayey Chalk passes into siliciclastic facies along both margins of the basin (Alcaydé et al., in [32]): into sandy nearshore to continental deposits toward the southwest (Touraine, western Sologne) and into clayey open marine deposits (*Dièves*) toward the NNE (Artois to northern Champagne area).

Few geometrical data are available for the White Chalk (late Turonian to Campanian). Toward the southwest (Touraine area), the White Chalk passes laterally into siliciclastic sandy, nearshore to continental facies [150]. The abundance of Senonian planktonic foraminifera in Champagne and the occurrence of phosphatic chalk in Picardy indicate opening of the sea toward the northeast [151]. Seismic data show "valley-like structures" 1.5–2 km wide to 100–150 m deep [152]. Their interpretation is still controversial. They could be real palaeo-valleys with a NW–SE direction [153] or a diagenetic front due to continental weathering after the end-Cretaceous deformation [154]. In coastal outcrops in Normandy, the Turonian-Coniacian chalk shows large-scale

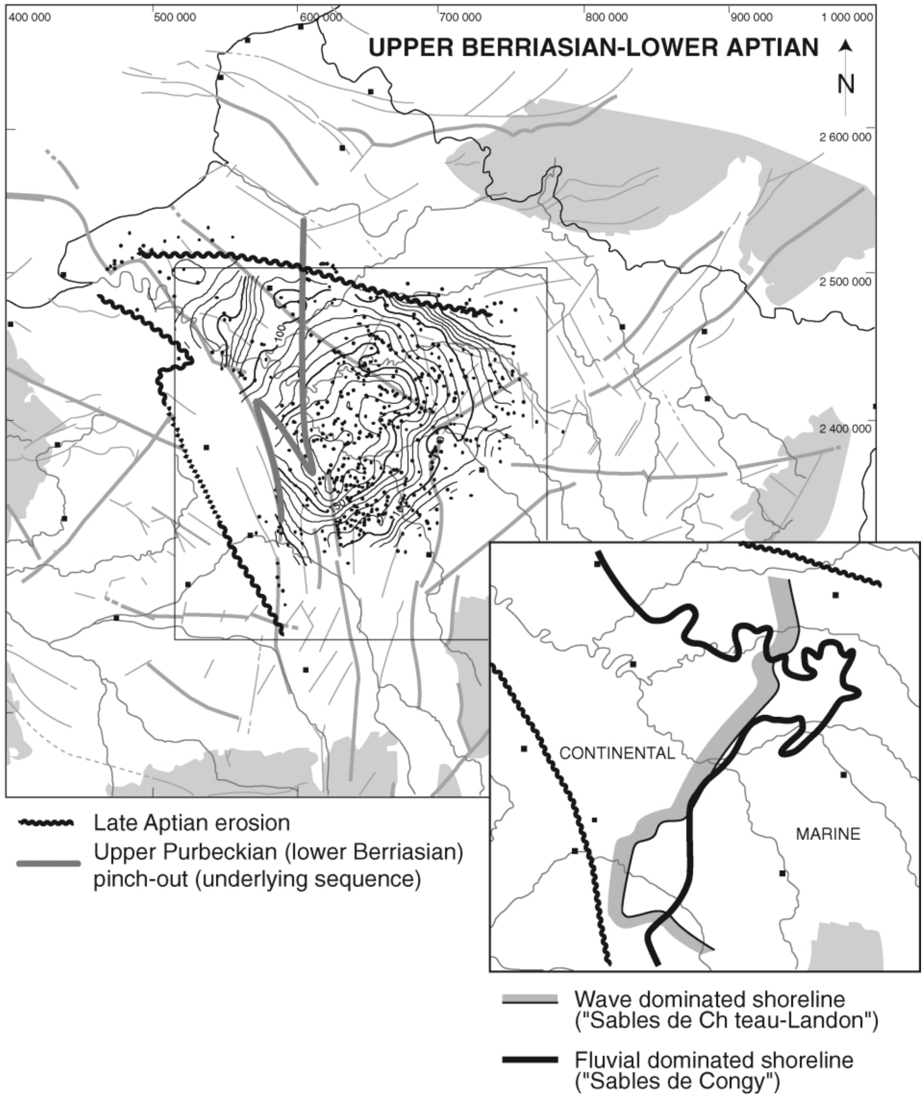


Figure 23. Isopach map from the Lower/Upper Berriasian unconformity (Ryzanian Unconformity, RU) to the Lower Aptian (maximum flooding surface of the *Argiles à plicatules*, Deshayesi zone): Lower/Upper Berriasian boundary–late Barremian cycle and transgressive trend of the late Barremian–early Albian cycle. Location of the Hauterivian (*Sables de Chateau-Landon*, wave-dominated) and Barremian (*Sables de Congy*, fluvial-dominated) 'Wealden' deltas, westward pinch-out (truncation) of the underlying Upper 'Purbeckian' sediments.

undulations (0.7 to 2 km wide, 20 to up to 70 m deep), interpreted by Quine and Bosence [155] as multistorey channels incised during relative sea-level lowstands.

Biostratigraphic data ([31] and Monciardini in [32]) suggest multiple condensations and truncations occurring along crests of synsedimentary folds of short wavelength (multiples of 10 km), precursors of the late Cretaceous/Tertiary ones (intra-Campanian truncations onto the Eure anticline, uppermost Santonian to Lower Campanian condensation and phosphatization on the Artois anticline, [156]). The tectonic movements, connected with condensation and truncation, started in the late Turonian/early Coniacian in the NNW part of the basin (Normandy-Picardy, [94]) locally even during the Middle Cenomanian. Both the Turonian and the Senonian chalk onlap onto the Variscan basement in the NNE (Brabant-London block) and probably also along the Armorican domain.

The Maastrichtian chalk is preserved along the western and eastern margins of the basin, lying paraconformably on the Upper Campanian chalk in the Mons 'Basin' [95, 96] or

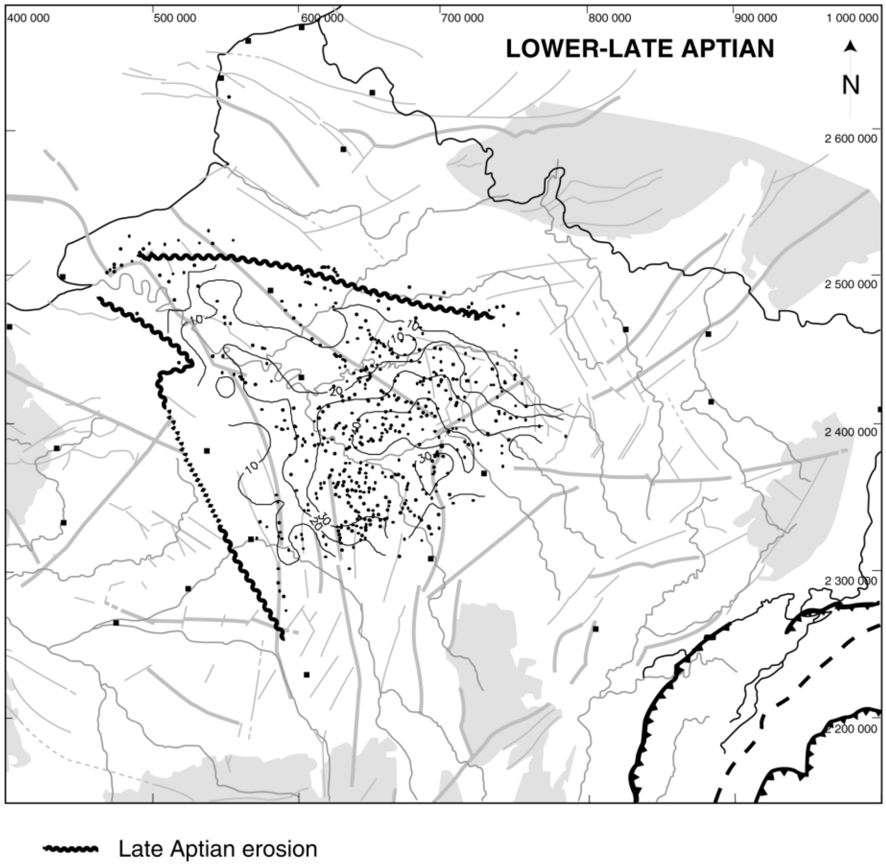


Figure 24. Isopach map from the Lower Aptian (maximum flooding surface of the *Argiles à plicatules*, Deshayesi zone) to the late Aptian unconformity (Jacobi zone): lower part of the regressive trend of the late Barremian–early Albian cycle.

unconformably on the Cadomian-Variscan basement and the Triassic and Lower Jurassic sediments in western Normandy [97].

The late Cretaceous unconformity truncates short wavelength folds. Thanetian sediments unconformably overlie Campanian to Turonian chalks along two major NW–SE-trending anticlines, the Bray (located south of the Bray fault) and the Artois anticlines (geological map of France at 1:250 000, Rouen and Amiens sheets, BRGM) and along numerous smaller ones in Picardy (NW–SE Thieux-Conty anticline, [156]) and in Champagne (NW–SE Mont-de-Champagne/Brimont anticlines, [157]; E–W Champillon syncline, [158]). The amplitude of these Late Cretaceous folds is at least of one order of magnitude higher than that of synsedimentary folds of the chalk (truncation of at least 150–200 m of Turonian to early Maastrichtian sediments at end Cretaceous along the Bray anticline vs. condensation and localized minor truncation along the Late Cretaceous folds).

The post-Cenomanian chalk isopach map *(figure 29)* mainly records the results of the end-Cretaceous deformation, with NW–SE-trending folds of short wavelength in the central part of the basin; namely, a NNW–SSE-trending anticline along the Seine fault, the NW–SE Bray anticline (dying out toward the southeast). Again isopachs suggest a subsident domain wider toward the southeast than the present-day area of post-Cenomanian chalk preservation.

The end Cretaceous unconformity documents a major compressive tectonic event with short wavelength NW–SE-trending folds.

4.8. Tertiary

Few 3D geometrical data are available for the Tertiary, mainly because of the very high facies variability in both space and time. No well-log correlation has been carried out in this study, which is based on a compilation of the scarce published data on sedimentary geometries. Only finite deformation from the base of the Tertiary or from the base of the Burdigalian to the present-day can be studied based on the isohypse maps.

Isohypses of the base Tertiary (Labourguigne and Manivit in [32], *figure 30*) show five main structural units: (1) the

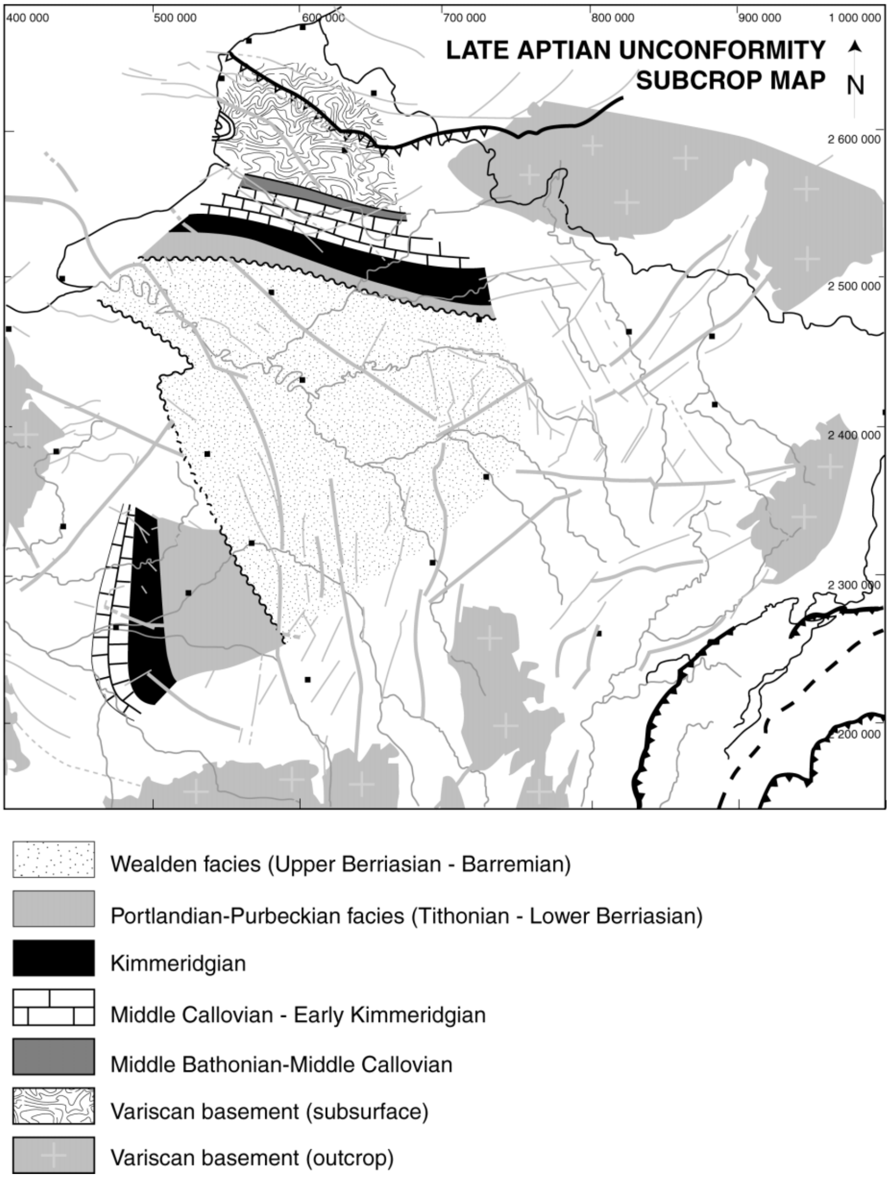


Figure 25. Late Aptian (Jacobi zone) unconformity subcrop map - subsurface data.

Flandres synform, (2) the Artois-Ardennes antiform, (3) the Ile-de-France domain, (4) the Beauce/Sologne domain and (5) the Touraine/Brenne domain.

The Ile-de-France domain, bounded southward by the La Remarde anticline, is characterized by short wavelength deformation along the older late Cretaceous NW–SE-trending anticlines and synclines. The main changes in comparison with the Late Cretaceous folds are the occurrence of the Beynes-Meudon and La Remarde anticlines and the lack of continuity of these NW–SE structures toward the southeast. In the southeastern Ile-de-France, the base Tertiary is a regular steep surface dipping toward the northwest. The Beynes-Meudon and La Remarde anticlines are westward bounded by the Seine fault. To the northwest (Picardy), the Margny-les-Compiègne and the Bray anticlines merge. These structures were later deformed by a NE–SW-trending antiform extending from Arras to Rouen.

The Beauce/Sologne domain is less deformed with two major ‘synforms’ (trough or *fossé* of the French authors): the N–S Ingrannes-Pithiviers and the NW–SE La-Ferté-Saint-Aubin ‘synclines’ separated by the Sennely/Seine Faults.

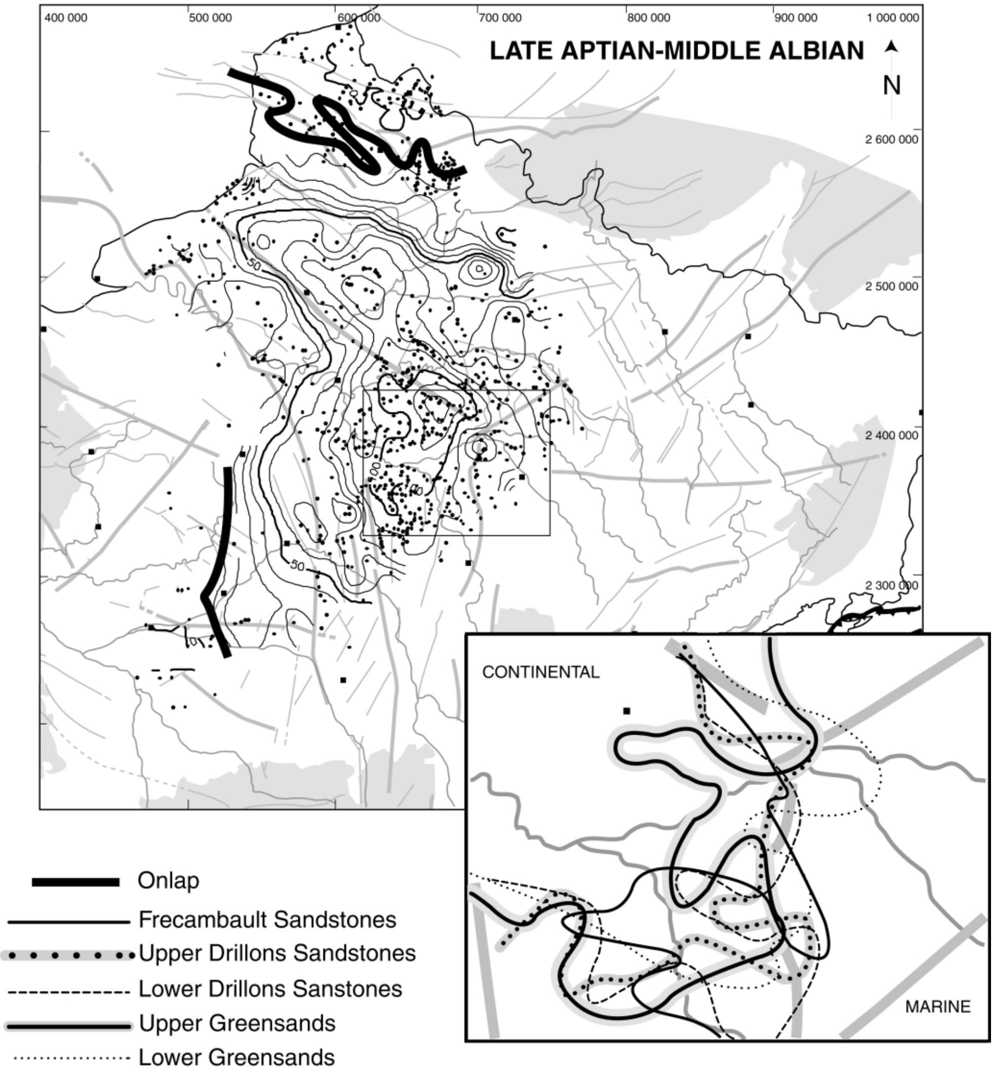


Figure 26. Isopach map from the late Aptian unconformity (Jacobi zone) to the Middle/Upper Albian boundary (maximum flooding surface of the Gault Clays): upper part of the regressive trend of the late Barremian–early Albian cycle and transgressive trend of the early Albian-Cenomanian cycle. Location of the Lower/Middle Albian shorelines for the central part of the basin.

The Touraine/Brenne domain is characterized by more or less E–W anticlines and synclines (Brenne, Esvres, etc.).

Isohypses of the base Burdigalian ([49], *figure 31*), only preserved in the SSW part of the basin (Sologne area), show a NE–SW to WSW–ENE-trending synform with a medium wavelength, asymetric to the SSE, with superposed short wavelength anticlines and synclines.

Looking at the 1:1 000 000 map of the Paris Basin ([40], *figure 2*), the most evident features are the distribution of the Meso-Cenozoic sediments along an antiform orientated ENE–WSW —the so-called *Seuil de Bourgogne* [159]— between the Morvan basement and the middle part of the Vosges basement and a large central synform where Tertiary sediments are preserved. The only chronological constraint for the formation of these features is given by the Plio-Pleistocene sediments which clearly cut across these structures. Their relationships with the late Eocene/Oligocene grabens (Rhine, Limagne and Bresse) are still poorly understood.

Another important set of data is the information contained in the Tertiary sediments including their internal unconformities *(figure 2)*.

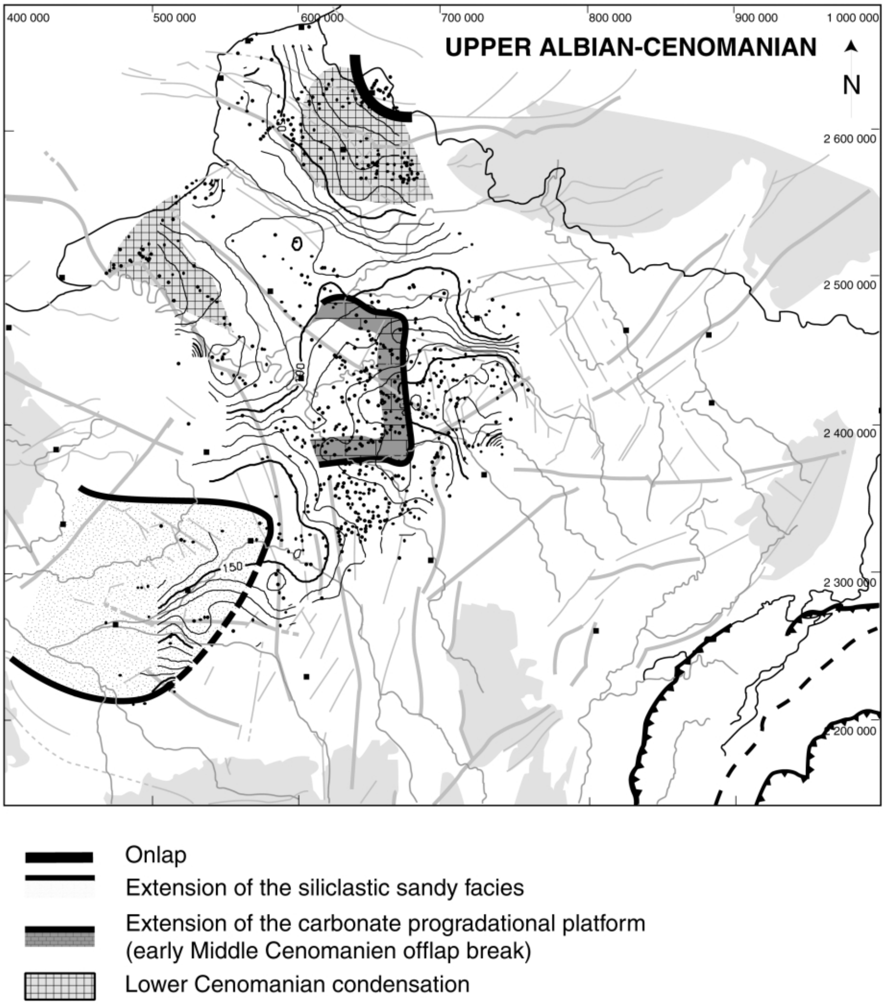


Figure 27. Isopach map from the Middle/Upper Albian boundary (maximum flooding surface of the Gault Clays) to the late Cenomanian (*Actinocamax plenus* Marls): regressive trend of the early Albian–Cenomanian cycle. Location at the early Middle Cenomanian boundary of the siliciclastic deltaic system and of the isolated progradational carbonate platform.

Lutetian sediments (lacustrine *Calcaire de Morancez*) overlie Paleocene and Ypresian sediments in the Beauce area, which is mainly a southward onlap effect;

fluvial and lacustrine deposits of Eocene age occur over all the Mesozoic sediments of the Paris Basin and over the northern part of the basement of the Massif Central;

Oligocene sediments overlie the Upper Cretaceous chalk and the'clays-with-flints' (1) in the NNW part of the present-day Paris Basin (Normandy) westward of the Seine River, and (2) along the E–W short wavelength La Remarde anticline, south of Paris;

Miocene deposits, in the southwestern part of the basin, overlie the Variscan basement (Anjou), the Upper Cretaceous chalk (Beauce and Sologne) and the Eocene continental sediments (Touraine and Brenne);

Plio-Pleistocene fluvial sediments cut across the tectonic structures of the Paris Basin, whatever their scale; Pliocene fluvial sediments crop out along a N–S corridor which coincides with the Sennely, Loire and Saint-Martin-de-Bossenay faults in the south and the Seine fault in the north.

Stratigraphically, the Tertiary can be subdivided into three periods: (1) a period of low accumulation rate, Paleocene to Lower Oligocene; (2) a period of by-pass or local accumulation - Upper Oligocene to Lower Pleistocene and (3) a period of erosion, Middle Pleistocene to present.

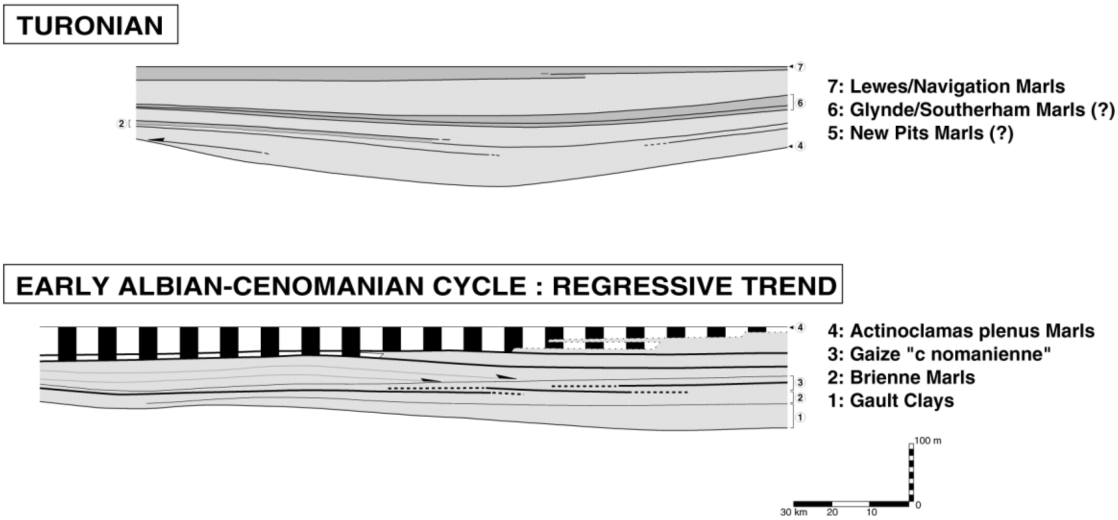


Figure 28. 2D section between Saint-Dizier and Rambouillet (see location on *figure 1a*) based on well-log correlations for Cenomanian and Turonian times.

The **period of low accumulation rate** is a period of low creation or removal of accommodation space. The transgressive periods of both Danian/Bartonian and Priabonian/Oligocene cycles coincide with the two moments of accommodation space creation, Ypresian and end Bartonian/Priabonian.

Paleocene to Middle Eocene: after the end-Cretaceous deformations, the central part of the Paris Basin is flooded by the sea and became again an area of deposition. The southern and western parts of the basin were emerged and subject to weathering and erosion. No traces of this weathering period or Tertiary sediments are, however, preserved in the eastern part of the basin, between the Morvan and the Ardennes *(figure 32)*, [160]).

Danian sediments are a few metres thick shallow-marine to lagoonal limestones overlying karsts in the Cretaceous chalk. Two episodes of marine floodings have been identified, but the Danian palaeogeography is poorly understood because of the scarcity of outcrops [100]. Synsedimentary NW–SE-trending normal faults are located on the crests of Late Cretaceous anticlines (e.g. Vigny anticline, [161]). They control both facies (location of reefs) and the occurrence of slides and slumps.

Thanetian sediments are mainly siliciclastic littoral to marine sediments, which onlap onto the chalk along a NNW–ESE line which more or less coincides with the Bray fault system (F. Mégnien in [32]). In the south, the Thanetian coastline did not reach the Danian one [100].

Ypresian siliciclastic coastal plain to open marine deposits, partly onlapping in the south, are more widespread than the Thanetian ones. At that time, the open sea was located in the northwest (Belgium, England and the North Sea, [162]). Occurrence of reworked Kimmeridgian palynomorphs at the base of the Ypresian sediments in the west [163] would indicate movements along the Bray anticline before the Ypresian. The late Ypresian *(figure 33)* was a time of fold deformation along the western Vernon and Meudon anticlines where the chalk was eroded [68].

During Lutetian time, the palaeogeography changed: the Paris Basin became a carbonate platform more and more enclosed (evaporitic facies) with time in response to uplift and short wavelength folding of the northern part of the basin (Laon area, [33, 68]). The paroxysm of these movements occurred in Lower Lutetian time. During Upper Lutetian, the northwest part of the present-day basin (Cotentin) was flooded and open marine bioclastic limestones were deposited. This area was part of the Atlantic domain with sediments different from those of the central basin [164].

From Paleocene to Lutetian, the type of weathering changed [165, 166]. The weathering profiles are dated based on their geometrical correlations with the marine deposits of the central part of the basin. The Paleocene is characterized by kaolinite-rich deep soils, indicating a warm and humid climate with little vertical tectonic movement. The Lower Eocene was a period of erosion which is replaced during the Middle and Upper Eocene by the development of silcretes and calcretes associated with calcareous lacustrine deposits [49, 53]. Lower Eocene drainage occurred along N–S-trending extensional faults (Loire and Loing faults, [155, 156]).

Bartonian: the Bartonian/Lutetian transition documents again a major sedimentary change, both lithologically with a siliciclastic Lower Bartonian (Auversian of the French authors) marine platform, and palaeogeographically with the

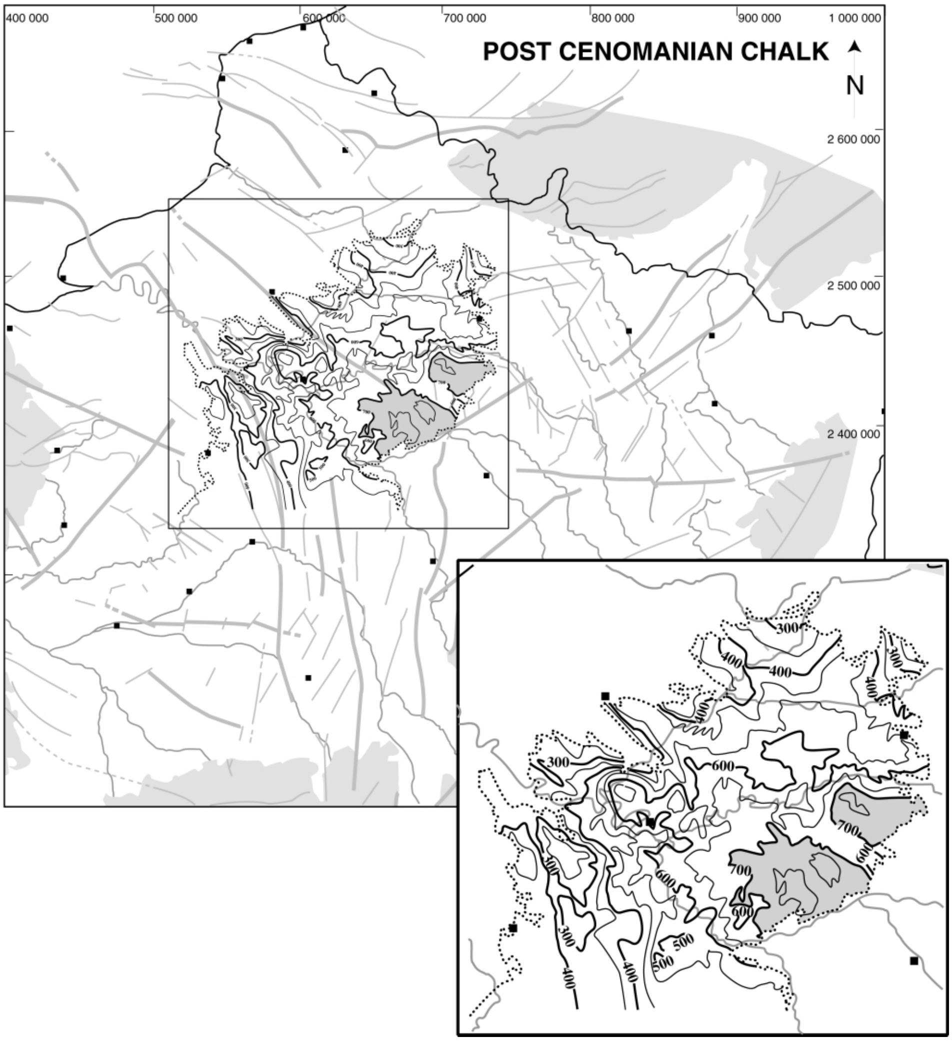


Figure 29. Isopach map from the late Cenomanian (*Actinocamax plenus* Marls) to the Late Cretaceous unconformity: post-Cenomanian chalk (from Hanot and Obert [149]).

development of a semi-enclosed sea open toward the English Channel. These siliciclastic sands pass, both laterally and vertically, into brackish and lacustrine calcareous deposits.

Locally, along the Bray anticline, late Bartonian (late P14/NP17) siliciclastic sediments unconformably overlie faulted and folded Lutetian to Upper Bartonian sediments [69, 70]. Control by wrench tectonics is assumed for the deposition of the Upper Eocene lacustrine sediments of the western part of the present-day Paris Basin and now preserved along faults like the Huisne fault (Maine, [53]). Unfortunately, no 3D stratigraphic data are available to support this idea: are these faults syn- or post-sedimentary (Oligo-Miocene) structures?

Late Bartonian–early Oligocene: during this time, lacustrine and evaporitic sediments with some brackish intercalations, suggesting a flat topography in very dry to arid climate with no fluvial input [167]. The geometries of the Bartonian to Lower Oligocene sediments are still unknown.

Lower Oligocene: the Oligocene deposits are lacustrine carbonate deposits (*Calcaire de Brie*) overlain by marine to eolian siliciclastic sands (*Sables de Fontainebleau* s.l.) which southward pinch out along a NNW–SSE-trending line between Chartres and Montargis (Ménillet in [32]). The *Sables de Fontainebleau* unconformably overlie all the Tertiary sediments to the Upper Cretaceous chalk along the E–W Remarde anticline [168]. Petrographic data [169] suggest an Ypresian, Albian and 'Wealden' origin for the Fontainebleau sands located north of the Seine River. The uppermost *Sables de Fontainebleau* shows soft-sediment deformation related to seismic events [170, 171]. In the southern and western parts of the basin (Sologne, Berry, Touraine and Anjou areas)

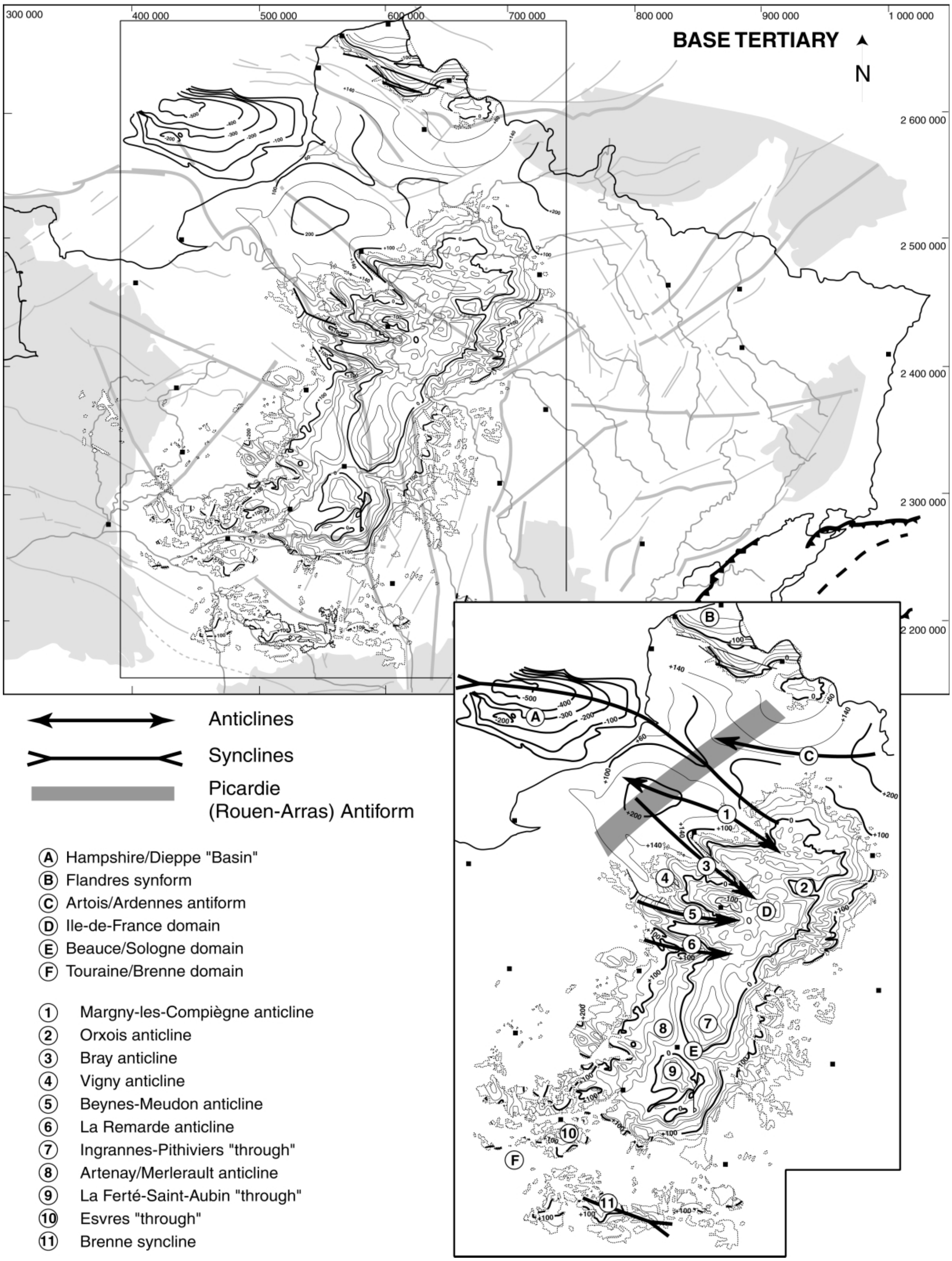


Figure 30. Isohypse map of the base Tertiary (modified from Labourguigne et al., [32] for the central part of the basin and from Hamblin et al. [37] for the Dieppe/Hampshire basin).

Figure 31. Isohypse map of the base Burdigalian (from Rasplus [49]) - Burdigalian sediments are mainly preserved in the south Beauce/ Sologne areas.

lacustrine sedimentation ended around the Eocene/ Oligocene boundary [49].

The **change from accumulation to sediment by-pass** occurred contemporaneously with a migration of the depocentres toward the SSE (Beauce and Sologne) starting at the end of the Lower Oligocene (lacustrine *Calcaire d'Etampes* overlying the *Sables de Fontainebleau*, [172]). The depocentres were now located along the present-day Loire River (Ligerian area, [108]). At the Aquitanian/Burdigalian boundary, a base-level fall (unconformity) occurred contemporaneously with the development of a large alluvial plain (*Sables de l'Orleanais* and *de Sologne*) coming from the south (Massif Central) and turning to the west along the course of the present Loire River. This alluvial plain was flooded by a tidal-dominated sea with deposition of bioclastic limestones (*Faluns* of the French authors). Locally, in the Sologne area, Burdigalian fluvial sediments are discordant above Oligocene to Aquitanian lacustrine deposits [71]. In other places, west and south of Tours (Blésois and Lochois), Lower Burdigalian to Upper Serravallian marine

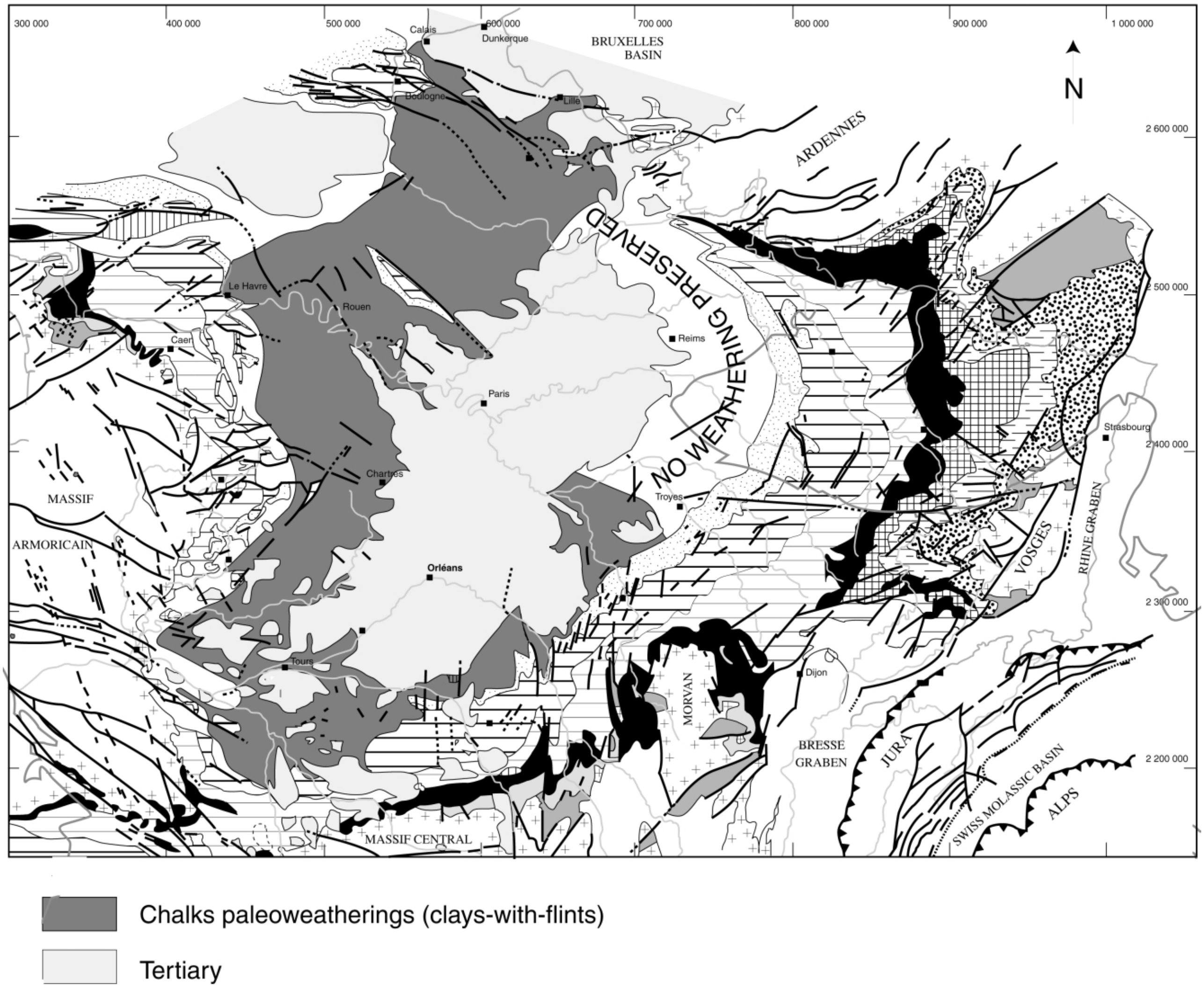


Figure 32. Distribution of weathering surface on the chalk (Clays-with-flints, from Quesnel [160]).

bioclastic sediments (*Faluns*) cut across Upper Cretaceous and Late Eocene to Lower Miocene (Aquitanian) sediments, which are slightly folded and fractured along NE–SW and NW–SE directions [106]. The *Faluns* themselves are fractured (joints) along a NW–SE direction.

A major palaeogeographic reorganization occurred during Upper Miocene times. Unfortunately, this event cannot be better determined because of the lack of well-dated marine and continental deposits [103]. It was a time of a general by-pass of fluvial sediments across the Paris Basin. Rivers coming from the south (Massif Central area) were captured [107, 108, 109, 173] westward to the Atlantic ocean (Ligerian area) during the Upper Pliocene, or northward to the English Channel during the Late Miocene/Lower Pliocene and the Lower Pleistocene.

Erosion started 800 000 years ago in the valley of the Seine and Somme Rivers [72], the capture of rivers from the south (Allier and Loire Rivers) occurred in Saalian and Weichselian times [173]. In the eastern part of the basin (Meuse, Moselle, Rhine, Saône and Doubs Rivers, [110]), finite erosion was between 2.5 and 0.6 Ma 50 to 70 m. At 0.6 Ma the incision rate increased to 30 m/100 000 years (Meuse-Rhine systems) with numerous captures (the Moselle by the Meurthe, [174]; the Aire by the Aisne, the Saulx and Ornain by the Marne, [175]). This widespread erosion indicates a global base-level fall during Middle and Upper Pleistocene times.

Synsedimentary Pleistocene tectonic structures could be the small faults observed in river terraces (Aube River along the N–S Saint-Martin-de-Bossenay Fault - Upper Pleis-

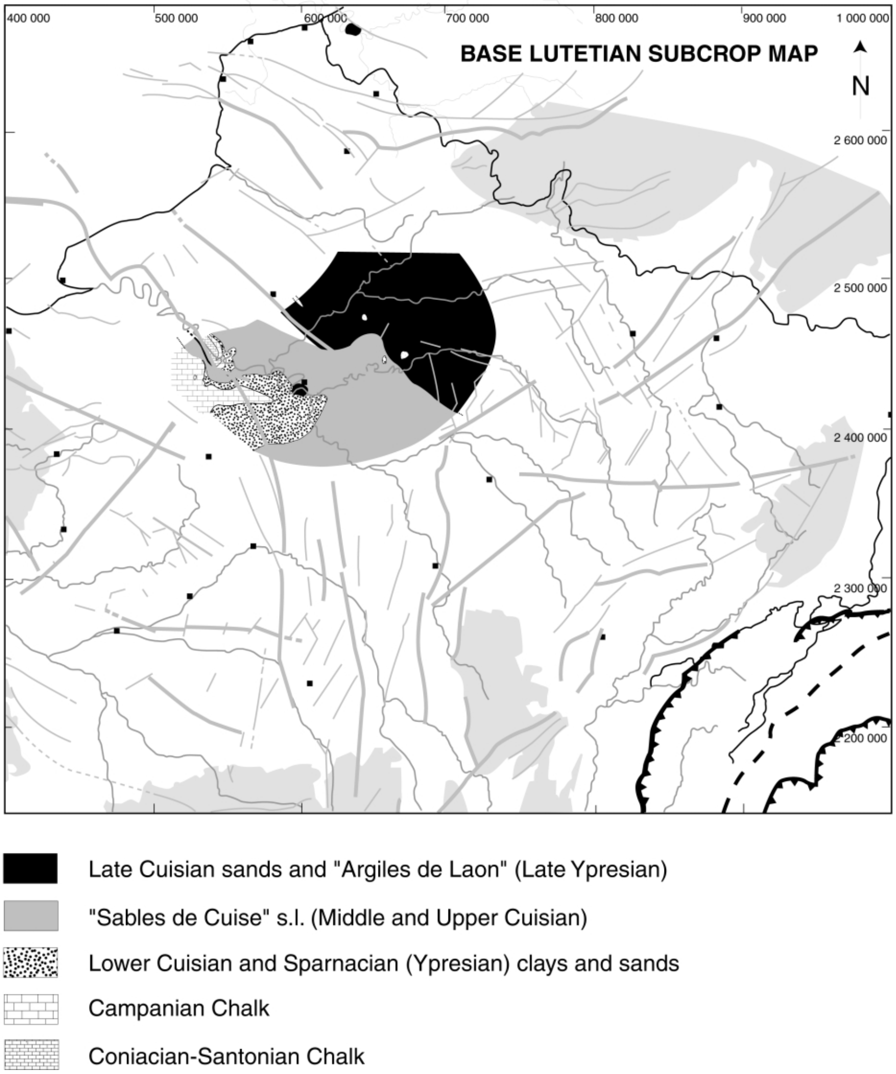


Figure 33. Ante Lutetian subcrop map (from Gély [68]).

tocene, [176]) or normal to strike-slip faults (Strait of Dover graben - *Pas-de-Calais* - of Lower Pleistocene age with Weichselian strike-slip movements along E–W to ESE–WNW faults, see Colbeaux et al. [137] for discussion).

The study of the levelling variations (geodesy) of the northeastern part of the Paris Basin [177] suggests significant recent tectonic vertical movements. The amplitude of these movements are still a matter of debate. West of a roughly N–S line between Metz (Lorraine) and Dijon (Bourgogne), the western area is rising and the eastern one subsiding. The highest uplift values occur in Normandy and Picardy along a NE–SW trend between the Seine (Rouen) and the Somme Rivers.

4.9. The main stages of the Paris Basin evolution

Based on thicknesses (subsidence), sedimentary systems, palaeogeography and variations in accommodation space, the evolution of the Paris Basin can be subdivided into five main stages:

(1) **Scythian to Toarcian** (Triassic to Lower Jurassic), with two discontinuities, the intra-Carnian (base *Grès à roseaux*) unconformity and the intra-Norian unconformity, and short wavelength tectonic control in the Lower/Middle Liassic;

(2) **Aalenian to Tithonian** (Middle/Upper Jurassic), bounded at the base by the Aalenian unconformity;

(3) **Berriasian to late Aptian** (part of the Lower Cretaceous), bounded by the late Cimmerian and late Aptian unconformities,

(4) **Albian to late Turonian** (part of the Lower and Upper Cretaceous),

(5) **late Turonian to recent**, characterized by a short and medium wavelengths deformation with paroxysms in the end Cretaceous and the Late Miocene.

5. Towards a stratigraphic and geodynamic history of the Paris Basin

5.1. Scythian–Toarcian (Triassic–Lower Jurassic)

Scythian to Toarcian times are characterized by (1) subsidence controlled by a bend-shape structure along NE–SW and E–W to ENE–SSW flexural trends, decreasing during the Lower and Middle Liassic, (2) onlaps toward the west and north and (3) a period of accommodation space creation (except during early Sinemurian).

5.1.1. Scythian–Toarcian hydrothermal events

Two major hydrothermal events are contemporaneous with sedimentation: at late Triassic (Rhaetian) and during Pliensbachian. The late Triassic 'event' is recorded from the Armorican border to the Vosges area. It has not been identified in the Ardennes area. In the Armorican domain (western Normandy), Triassic sediments are mineralized by baryte, galena and blende, and are reworked as pebbles at the base Hettangian [132]. In the northeastern Massif Central area (Morvan), synsedimentary mineralizations (blende, fluorite, pyrite, galena and baryte) are associated with late Triassic coastal plain deposits [178]. The late Pliensbachian 'event' is recorded along both margins of the basin (Armorican and Vosges domains) and within the basin. In the Armorican domain, baryte-rich dykes cut across Lower Pliensbachian sediments (Carixian) and occur as clasts in Upper Pliensbachian (Domerian) sediments [132]. In the Massif Central (Morvan), fluorite-rich dykes were emplaced at 185 Ma, at a depth of 500 m, with a temperature of around 165 °C [179]. In the southern Vosges [180], Carboniferous magmatic rocks show a Liassic magnetic overprint (205–170 Ma), a consequence of mineralizations. In the central part of the present-day basin, Rhaetian sediments contain illites crystallized around 190 Ma during a short event of less than 1 My duration, at a depth of around 500 m and with a temperature of 220–250 °C [15]. 190 Ma corresponds to the Pliensbachian stage (194–187 Ma, [116]; 195.3–189.6, [181]).

These hydrothermal events are widespread in France. The late Triassic 'event' is coeval with volcanic activity in the Alps (Dauphinois, alkaline basalts interbedded in the uppermost part of the late Triassic dolomites, [182]), in the Pyrenees/Aquitaine Basin (tholeiitic continental ophitic dolerites contemporary with Rhaetian/Lower Hettangian tuffaceous explosive volcanics, [183]) and in the Armorican Massif (200 Ma-old dolerites, [184]). In the western part of the Armorican Massif [185], dolerites show ages around 233 Ma and 210–190 Ma, rejuvenated by hydrothermal activity around 156 Ma. The Pliensbachian 'event' has been recognized in the Massif Central (Ardeche palaeomargin, illite of 190 Ma, [186]; Lodève Basin, 190–170 Ma, [187]; eastern and northern parts of the Massif central, fluorite-rich dykes at 194 Ma, [188]).

5.1.2. Tectonic and geodynamic significance of the Triassic unconformities

Little is known about Triassic deformation in western Europe. In the Paris Basin, N–S extension is assumed for the Permian and the Triassic [189, 190] based on few microtectonic data in the Massif Central [191] and on the E–W orientation of the Lower Keuper salt depocentre. Along the Ardeche palaeomargin [192] a structural study of orientated cores suggest an E–W extension in the Ladinian (late Middle Triassic)-early Carnian (early Upper Triassic) with tilting (before the deposition of the *Barre carbonatée médiane*, a time-equivalent of the Lettenkohle). During the Norian, a progressive change occurred from an E–W to a N–S extensional pattern. This E–W extension contrasts with the N–S extension assumed for the Paris Basin.

Our data suggest subsidence along two main directions, E–W and NNE–SSW, which might be consistent with both N–S and E–W extension with permutation and rotation of the stress axis. The late Ladinian (Lettenkohle) and intra-Carnian (base *Grès à roseaux*) non-truncating unconformities and the intra-Norian truncating unconformity do not record major changes in the subsidence pattern, but a migration of the depocentres. The importance of E–W subsidence trends decreased during the Triassic, what might suggest more E–W directed extensional control during late Carnian/ Norian times.

Little is known regarding the geodynamic significance of the base *Conglomérat principal* and base Lettenkohle unconformities.

The base *Grès à roseaux* (middle Carnian) corresponds to a world-scale discontinuity interpreted as climatic [193]. In the Paris Basin, in contrast to the German Basin [194], the base *Grès à roseaux* did not involve significant erosion (no evidence for incised valleys). The maximum thickness of these fluvial deposits documents a westward shift of the depocentre [62]: it records the first occurrence of a subsident area in the central part of the present-day Paris Basin, which was no longer the western end of the German Basin. A similar tectonic origin is assumed for time-equivalent deposits of the Schilfsandstein of the German Basin [195].

The intra-Norian unconformity was first identified in the Paris Basin by Bourquin and Guillocheau [61, 24] and subsequently recognized in the Bresse-Jura Basin [196]. In the German Basin, it might be the time-equivalent of the erosional truncation of the Stubensandstein [194], and of the intra-Norian discontinuity of the Barents Sea and the Dolomites [123]. In the absence of biostratigraphic data in the Paris Basin, this unconformity might be correlated with the base Norian (early Cimmerian I unconformity of Ziegler [123]), mainly recorded in areas of low subsidence and interpreted as due to a change in intraplate stress regime in response to the closure of the Black Sea back-arc Basin [123].

5.1.3. Tectonic and geodynamic significance of the Lower/ Middle Lias: a short wavelength control

The Rhaetian/Hettangian deepening event, coeval with a change of the tectonic wavelength in the Paris Basin, was at least a West European-scale event (early Cimmerian II unconformity of Ziegler [123]) in response to a slight change in the movements of the continental masses at the time of the break-up of Pangaea. The Rhaetian/Hettangian boundary and the early Hettangian record the very beginning of the rifting phase of the Ligurian Tethys [182]; this occurred just after the late Triassic volcanic and hydrothermal "events".

The Hettangian/Sinemurian, late Sinemurian (base Raricostatum zone, "Lotharingian" events), late Carixian (Ibex/ Davoei) reorganizations of the subsidence pattern and the late Domerian-early Toarcian wavelength change are contemporaneous with increasing rates in the extension of the Ligurian Tethys [197–200] along a NW–SE direction [201]. Unfortunately some of these data are in conflict with this view: this could be explained by data provided by different analytical methods or by a diachrony in deformation. In the Dauphinois and Briançonnais domains, extension occurred during Hettangian/early Sinemurian, with a paroxysm at the Hettangian/Sinemurian boundary, and in the Upper Pliensbachian (Domerian)/Toarcian/Aalenian [202]. In the Subalpine Basin (Digne area), the Lower Pliensbachian (Carixian) and the late Upper Pliensbachian (Domerian) were times of extension with increased subsidence rates, in contrast the Lower Domerian and the Toarcian were times of more equally distributed subsidence [198]. Along the Ardeche palaeomargin, extension occurred during the Sinemurian and the Pliensbachian after a Lower/Middle Hettangian subsiding period and late Hettangian/early Sinemurian uplift [199]: maximum rates of vertical displacement along faults occurred at the late Hettangian/early Sinemurian and during the late Pliensbachian (Domerian, [200]). The Toarcian was a time of regional (flexural) subsidence.

In the Paris Basin, E–W to NW–SE extension is assumed for Lower Jurassic times [189–191, 203] based on microtectonic data (WNW–ESE along the Loire fault to NW–SE, Morvan, [191]), synsedimentary N–S faults and isopach maps. All authors agree that a major change in the extensional regime occurred during the Late Triassic (Rhaetian).

Our data are compatible with E–W to NW–SE extension during the Lower and Middle Liassic and with an uplift at the Hettangian/Sinemurian boundary (accommodation space removal). The Pliensbachian/Toarcian boundary included a major change in the distribution of the subsidence in the Paris and Subalpine Basins, after the Pliensbachian hydrothermal 'event'.

5.2. Aalenian–Tithonian (Middle–Upper Jurassic)

Aalenian to Tithonian times are characterized by (1) a major Aalenian unconformity with accommodation space removal (–5/+2 m/My), (2) a change in the subsidence regime now controlled by NW–SE flexural trends, (3) a period of accommodation space creation, with a maximum period of accommodation space creation during the Kimmeridgian and (4) northward and westward general onlaps.

From the Aalenian to the Tithonian, short wavelength tectonic controls ceased and subsidence became more equally distributed (high wavelength) during the Kimmeridgian-Tithonian. Short wavelength tectonic control still occurred during the Aalenian/Lower Bajocian at the basin-scale, and during the Bajocian/Bathonian in the Armorican and Brabant domains. Two main time intervals may be distinguished: an Aalenian/Lower Callovian interval and a Middle Callovian/Tithonian interval.

5.2.1. Tectonic and geodynamic significance of the Aalenian unconformity (mid-Cimmerian unconformity)

The Aalenian is again a time during which the west European palaeogeography and subsidence pattern changed (mid-Cimmerian unconformity of Ziegler [123]), recording both the early stage of the rise of the thermal dome in the central North Sea [204] and a further step of the Tethyan rifting (late Liassic, [205]). In the Dogger, there was a major break (180–170 Ma, base Aalenian to Bajocian) in the evolution of the Pangaea, i.e. from rifting (Triassic/Liassic) to the break-up into three separated continental masses (Laurasia, East and West Gondwana, [206]) with the opening of the Central Atlantic ocean [207].

In the Paris Basin *(figure 14)*, base Aalenian truncation and erosion across N–S to NNE–SSW-trending antiforms and synforms suggest an E–W to ENE–WSW compression.

5.2.2. Aalenian–Tithonian tectonic and hydrothermal events

During the Callovian, the rate of separation of Laurasia (including Europe) and West Gondwana (including Africa) increased. It was the time of the beginning of the oceanic accretion in the Ligurian Tethys [202, 205] and of the post-rift thermal subsidence of the margins [200, 203].

Hydrothermal activity is still present along the Armorican domain [132, 208]: Middle Bathonian sediments are mineralized by baryte, fluorite, galena and blende mineralizations and pebbles derived from them occur in Upper Bathonian deposits. In the southern Vosges [180], Carboniferous magmatic rocks show an Upper Jurassic–Cretaceous magnetic overprint (150–70 Ma), a consequence of mineralizations.

In the central part of the present-day basin, Rhaetian sediments contains illites crystallized around 150 Ma (short event of less than 1 My duration), cooler than the 190 Ma event [15]. 150 Ma corresponds to the Oxfordian (154–146 Ma, [116]) or to the Tithonian (150.7–144.2, [181]).

Extension directions during Middle and Upper Jurassic are difficult to establish. An E–W extension is assumed for the Middle Jurassic (mainly based on isopachs [189] and microtectonic data on the Loire fault, [191]). This is not consistent with (1) the Bajocian/Bathonian synsedimentary brittle faults: E–W (N80) normal faults, E–W (N80) and N–S (N10) extension joints (ante late Lower Bajocian, Humphriesianum zone, Burgundy High, [133]), (2) the WSW–ENE normal faults on the crests of small synsedimentary anticlines (Lower/Upper Bajocian boundary, Berry, [52, 134]) and (3) the E–W-trending normal faults (Upper Bathonian, Boulonnais, [137]) which suggest more N–S extensional movements. These small-scale tectonic data are too scarce to constrain the stress regime contemporaneous with the NW–SE medium wavelength flexure.

The interpretation of Upper Jurassic deformation is more controversial: according to Mascle and Cazes [190], extension is E–W during the Kimmeridgian and according to Blès et al., [191] N–S during the late Jurassic, however, there are only a few arguments for either interpretation.

Our data suggest (1) a change in the location of the subsidence which would mean a change in stress regime between the Liassic and the Dogger and (2) a lack of fault activity during the Upper Jurassic, which would be inconsistent with substantial extension and could explain the scarcity of brittle faults. The demise of fault tectonics in the Middle Callovian/Tithonian could record the rift/post-rift transition in the Ligurian Tethys and the evolution to a subsidence regime controlled by cooling of the lithosphere.

The tectonic evolution of the Dogger is still poorly known and requires more 3D data; tiltings and fault movements can be expected during the late Lower Bajocian (base Humphriesianum zone) and during the lower Bathonian. These deformations have been identified at the same time in the London Basin.

5.2.3. The sedimentary record

Space and time variations of both relative sea-level and carbonate productivity control the facies and the geometries of the Aalenian-Tithonian carbonate platforms. Four main platforms have been distinguished:

(1) a Lower/Middle Bajocian platform dominated by skeletal limestones (mainly crinoids with small reefs);

(2) an Upper Bajocian/Lower Callovian aggradational bank, made up mainly of ooids and skeletal-accumulations with some carbonate mud;

(3) an Upper Callovian/early Kimmeridgian progradational platform, mainly composed of carbonate mud more or less pelletoid-rich, with some reefs, ooids and oncoids and local siliciclastic influences (*Sables de Glos/Hennequeville*);

(4) a Tithonian aggradational platform, made up of carbonate mud with few ooids and skeletal deposits.

Regarding variations in carbonate production, two main facts have to be pointed out:

(1) a major fall of carbonate production at the time of low accommodation space creation occurred during late Bathonian/early Callovian [80];

(2) the Middle Callovian/Oxfordian progadational geometry occurred at the time of high rate of accommodation space creation (A>>0) which suggests very high rate of carbonate production (S>A).

These variations of carbonate productivity are widespread in France [209].

5.3. Berriasian–late Aptian: first compression in response to the opening of the Bay of Biscay

Lower Cretaceous times are characterized by (1) a change in sedimentary systems (siliciclastic vs. carbonates for Jurassic times), (2) a low rate of accommodation space creation (0/5 to –2/+15 m/My), (3) a NW–SE-trending medium wavelength flexural regime and (4) two major unconformities (early Cretaceous and late Aptian) as a consequence of the medium wavelength deformation and truncations.

5.3.1. Tectonic significance of the Lower Cretaceous unconformities

Both Early Cretaceous (late Cimmerian unconformity of Ziegler [123]) and late Aptian (‘Austrian phase’) unconformities have the same geometries and truncations: a NW–SE-trending synform with an east-dipping basin margin along the Armorican domain and a south-dipping basin margin in the Artois (southern border of the Dinant synclinorium) and the Ardennes, with paraconformable contacts in the central part of the basin. The late Aptian unconformity enhanced the late Cimmerian ones, which as a consequence are poorly preserved. In the central part of the basin, geometrical relationships between Upper Jurassic, ‘Wealden’-Aptian, and Albian sediments suggest a higher amplitude for the late Aptian deformation *(figures 21, 22 & 25)*. This can be interpreted as a medium wavelength NE–SW to E–W compression of the Paris Basin with a localization of deformation along the Armorican and the Artois–Ardennes basin margins.

These late Aptian to early Albian movements in the Paris Basin *(figure 34)* are known from both microtectonic field studies of brittle faults [210] and subsurface data (determination of faults displacement from seismic and well data, [123, 203, 211, 212]). Unfortunately, no measurement of the horizontal component of movement is available. Field data [210] suggest an E–W (Armorican border) to NE–SW (Bray

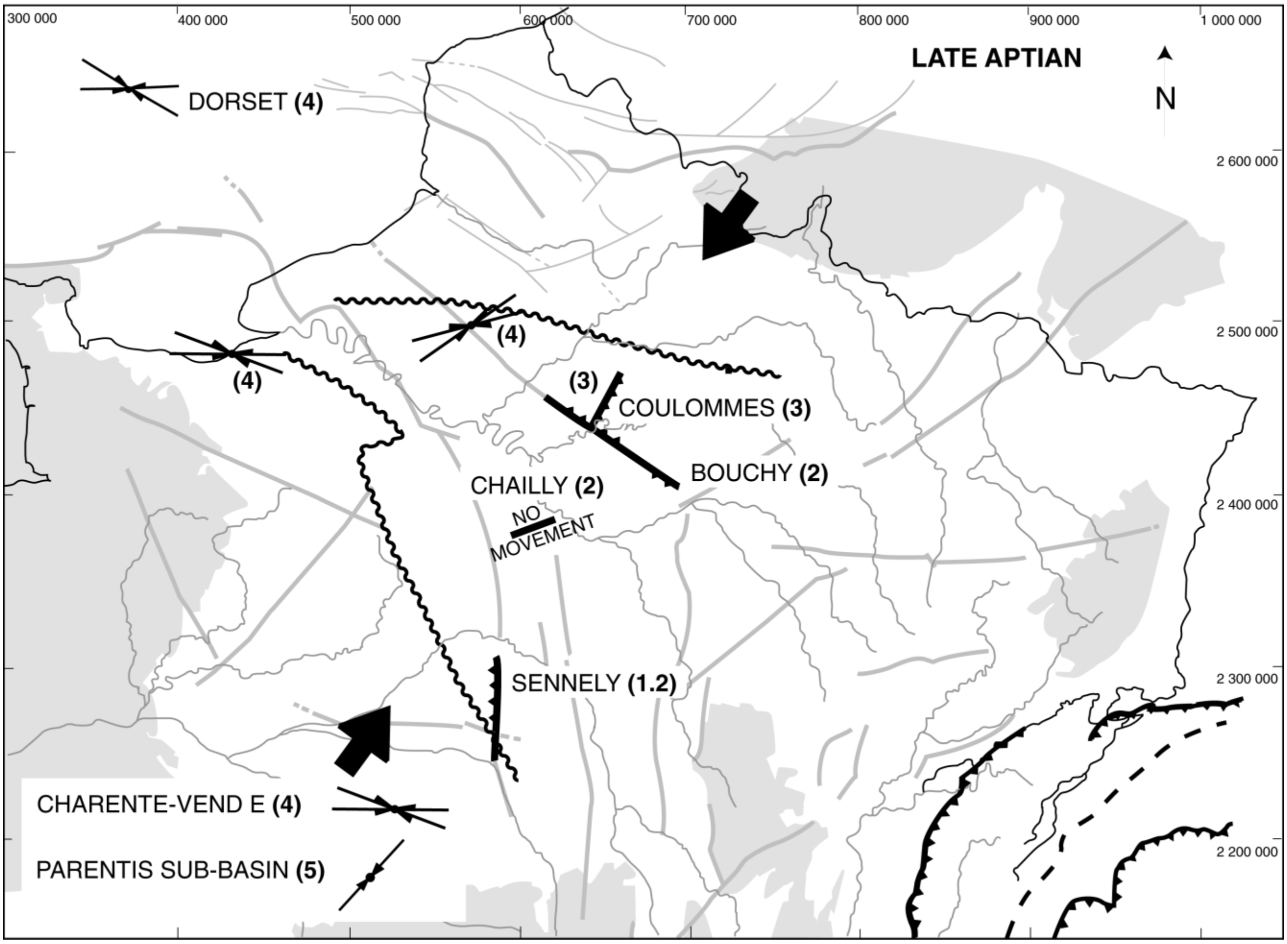


Figure 34. Late Aptian/Early Albian deformation field. 1, 2, 3: subsurface data (vertical displacement from well correlations, 1: Héritier and Villemin [211]; 2: Trémolières [203]; 3: Monchaux and Trémolières [212]). 4: microtectonic data (Bénard et al. [210]). 5: Parentis Sub-basin (Masse [216]). The dip of the Bray fault changes between Coulommes (NE dip) and Bouchy (SW dip) areas.

fault) trending axis of compression during Lower Cretaceous. Subsurface data show an inversion of N–S faults and, to a lesser degree, NNE–SSW and NW–SE-trending faults, before and contemporaneously with the deposition of the Greensand. As a consequence a late Aptian to early Albian age has to be assumed for these movements. WSW–NNE-trending faults (Chailly-en-Bière field, [203]) do not show any inversion movements. All these data are compatible with E–W to NE–SW axis of compression. The N–S inverted faults are localized in the southern part of the basin (Sennely, Sens, Saint-Martin-de-Bossenay faults, ...).

Both unconformities occur over most of western Europe. In the Wessex Basin, the major unconformity is of late Aptian/earlyAlbian age (base of the Lower Greensands). It records the end of an extensional phase along E–W-trending faults active during Late Jurassic/Early Cretaceous time [213]. In the Aquitaine Basin, two major unconformities exist at the base Cretaceous and at the Aptian/Albian boundary [214]. They record opening of the Bay of Biscay, documenting a change from possible NNE–SSW extension (late Jurassic-Valanginian) to oblique extension with a strike-slip component (Hauterivian-Lower Albian, [215]). In the Parentis Sub-basin, the Aptian/Albian unconformity records NE–SW compression [216]. In the Bresse area, a major truncation occurred between Hauterivian and Middle Albian, coeval with a tilting with a southward component (Cotillon et al., in [138]).

5.3.2. Geodynamic significance of Lower Cretaceous deformation

The North Atlantic Rift initiated at the Callovian/Oxfordian boundary in the West Portuguese Margin [217] and in the Berriasian in the West Galicia Margin [218]. During the late Aptian rifting ended and oceanic accretion started in the Bay of Biscay [219].

5.3.3. *Consequences for the sedimentary record*

The late Cimmerian unconformities were contemporaneous with (1) a sharp decrease of the subsidence rate (low rate of accommodation space creation) between Upper Jurassic and Lower Cretaceous and (2) the development of 'Wealden' siliciclastic deltas prograding from the Armorican domain toward the present-day Paris Basin and the eastern Channel Basin (Central Channel, Wessex, [220]), which indicate emersion and erosion of the Armorican Cadomian-Variscan basement.

The tilted and eroded Armorican basement and the associated low subsidence of the Paris Basin may be interpreted as the consequence of the pre-rift thermal doming [123] and the evolution to the northeastern shoulder of the Bay of Biscay rift during the Lower Cretaceous. Two consequences can be expected (1) an uplift of the western part of Europe and (2) an E–W to NE–SW compression.

5.4. Albian to late Turonian

The Albian to late Turonian interval is characterized by (1) a NW–SE medium wavelength flexural control, (2) an increase of the rate of accommodation space creation, (3) a general onlap of the sediments toward the west and the northeast and (4) a change of the sedimentary system, from siliciclastics (Greensands) to carbonates (Cenomano-Turonian chalk).

A few Albian to late Turonian brittle faults have been identified, except in the Boulonnais (northern part of the Paris Basin) with N–S compression/E–W extension of early Cenomanian age followed by an E–W extension during late Cenomanian [221]. This E–W Cenomanian extension could be an explanation for the increase in sediment thickness along a WSW trend in the Boulonnais *(figure 29)* which appears to have been part of the Weald Basin at that time.

5.5. Late Turonian–recent

The late Turonian to recent time interval is characterized by an overall decrease of the rate of accommodation space creation (late Turonian-late Cretaceous, Paleocene–Lower Oligocene) evolving to sediment by-pass (Upper Oligocene-Lower Pleistocene) and uplift (Middle/Upper Pleistocene), in a compressional framework, with a major short and medium wavelength unconformity at the end Cretaceous.

5.5.1. *Late Turonian to late Maastrichtian deformation*

The post-Turonian Upper Cretaceous time span is characterized by short wavelength synsedimentary deformation (NW–SE to E–W folds) with a paroxysm at the end of the Cretaceous times (end-Cretaceous unconformity). This mean NNE–SSW to N–S compression of the Paris Basin.

From the late Turonian to the end of the Maastrichtian, basement structures were inverted: formerly subsiding flexural areas (along the Bray and Somme faults, for example) became anticlines in the late Turonian-early Coniacian [153]. These inverted structures were previously described along the ECORS seimic line [190] but correlated with the 'Pyrenean' (Eocene) deformation.

In the Mons 'Basin', field data on brittle faults suggest NW–SE extension during the late Campanian evolving, during the early Maastrichtian, into NE–SW extension with pull-apart [222].

The late Campanian NW–SE extension is compatible with NE–SW compression. A major problem is to explain the early Maastrichtian pull-apart processes. This could suggest a more complex history for the end-Cretaceous compression which could occur throughout the Maastrichtian. This could explain the occurrence of Maastrichtian chalk in the Cotentin overlying basement and tilted Tiassic/Lower Jurassic sediments and in the 'Mons' Basin: deposition areas are located on the western and eastern margins of the inverted domain [223].

With a few exceptions [60, 67, 189], this Late Cretaceous deformation has been underestimated by geologists working on the Paris Basin. Its importance has been pointed out by geochemists working on fluid migration and diagenesis. Fluid inclusions data [224, 225] indicate for the Late Cretaceous time a geothermal gradient 25 °C higher in comparison to that of the present-day. Part of the illites contained in Rhaetian Sandstones crystallized around 80 Ma (event duration 40 Ma), under temperatures cooler than those of the 190 Ma (220–250 °C) and 150 Ma events [15]. 80 Ma corresponds to the Campanian stage (83–74 Ma, [116]; 83.5–71.3, [181]). This was probably caused by late Cretaceous uplift in addition to the thermal screen-effect of the overlying porous chalk and Cretaceous surface temperatures which reach its maximum.

5.5.2. *Tertiary deformation*

For the Tertiary time the relationships between tectonics and sedimentary basin fill are less constrained. Many tectonic studies on brittle faults have been carried out over the past 20 years. Unfortunately, little is known about the relationship between stratigraphy, fold structures and brittle faults for these times. Most authors working on brittle faults in order to define the average stress field agree on the occurrence of three major deformation stages in the Tertiary [48, 191, 203, 226, 227].

<u>Palaeocene–Late Eocene - paroxysm 40 Ma, Pyrenean phase</u>: there is general agreement on a near N–S compression prior to the Late Eocene/Oligocene extension, except in the northern part of the basin (Boulonnais, [221]; Mons 'Basin', [222]). Chronological data are poor: this stress regime could range from the Late Cretaceous to the Late Eocene. It has been correlated with the so-called 'Pyrenean phase' of Late Eocene age (40 Ma).

Late Eocene/Oligocene: this was a time of E–W extension (35–30 Ma), coeval with the development of the west European grabens system. Extension is recorded (1) in the western part of the present-day Paris Basin [228] along a NE–SW corridor bounded by the NE–SW Omey/ Luxemburg, Metz and Vermenton faults and limited to the southwest by the N–S Loire fault, and (2) possibly in the northern part (Boulonnais area: [221]). The youngest deformed sediments are the Champigny Limestones of Priabonian age (post P16 deformation) in the Fontainebleau area. No evidence of extension has been identified to the west [227].

Neogene: there is no agreement on chronology or the direction of faults: this may be due to a deformation field more complex than in the Palaeogene.

In the Morvan and North Burgundy areas [226], two major phases have been documented (1) a NE–SW compression of early Miocene (22–20 My) age and (2) a NW–SE to WNW–ESE compression of Late to post-Miocene (7–4 My) age.

In the western part of the basin, along the present-day boundary with the Armorican basement [227], the Miocene is characterized by two E–W to ESE–WNW compressions, and Pliocene by N–S extension preceding Pleistocene s.l. N–S to NW–SE compression.

In the southern part of the basin, along the present-day boundary with the Massif Central basement [229], tectonic and geomorphological data indicate E–W compression of Burdigalian to Upper Miocene (?) age. This compression is followed by a complex alternation of E–W extension and N–S compression of Upper Miocene (?) to Pliocene age. The late Pleistocene is characterized by NNW–SSE compression.

In the north, no Mio–Pliocene compression was recorded in the Mons ‘Basin’ [222], but NE–SW extension is recorded all along the French coast of the Channel between Le Havre (Normandy) and Calais (Boulonnais/Flandres) and is active until the Quaternary [230].

Present-day: the measurement of the present-day stress field [231] shows NW–SE to NNW–SSE compression of the southeastern part of the basin.

We propose here, based on the rather sparse tectonic and stratigraphic data, an attempt to reconstruct the tectonic evolution of the Paris Basin for the Tertiary. This has to be confirmed by 3D geometrical data based on sequence stratigraphy and geomorphology (surfaces and weatherings).

Paleocene to Lower Eocene: the Lower Paleocene (Danian) was a time of decreasing tectonic activity. The sea flooded the space created by the end-Cretaceous medium wavelength tilt toward the N–NE (development of a carbonate platform). During the Upper Paleocene (Thanetian) and Lower Eocene (Ypresian), medium wavelength folding of the lithosphere was again active, with onlaps of the Thanetian and Lower Ypresian siliciclastic sediments towards the S–SW; continental erosion occurred in the southern part of the basin. The Belgium and Paris Basins were parts of the same basin.

The direction of the large synform, a consequence of medium wavelength deformation, can be estimated by mapping onlaps, marine and coastal plain sediments and preserved traces of weathering. This structure is orientated along an E–W to ESE–NNW axis. This means a near N–S compression, which would explain the N–S orthogonal normal faults (Loire and Loing faults).

Middle to Upper Eocene: the Lutetian records changes on basin-scale. In the southern and western continental areas, erosion came to an end and silcretes and lacustrine/calcretes deposits (mainly of Upper Eocene age) formed. In the central area, after a major hiatus, a carbonate platform onlaps toward the south. This means a variation in the medium wavelength folding initiated at the Cretaceous/Tertiary boundary.

From the Lutetian to the late Bartonian [68, 33, 69, 70] short wavelength folding occurred. It could be the time of the NW–SE anticlines formation shown on the base Tertiary isohypses map (Labourguigne et al., in [32], *figure 30*). These deformations can be related to the Pyrenean phase s.l. (N–S compression). Unfortunately, geometrical data do not allow a determination of the kinematics of deformation and its paroxysm.

Late Eocene to Oligocene: tectonically, the Paris Basin became subdivided into two major domains of deformation separated by the N–S Loire and NW–SE Omey/Luxemburg faults *(figure 35)*.

To the east, E–W extension occurred. The Paris Basin was located on the western shoulder of the Rhinegraben [232, 233] and no sediments have been preserved. The exact chronology of extension in the Paris Basin is not known: it is, however, younger than zone P16 (Priabonian).

To the west, short-wavelength folding occurred during Lower Oligocene times (paroxysm P19-20/base NP 24): it was synsedimentary with mainly E–W-trending anticlines and synclines (La Remarde anticline, Artenay/Merlerault anticline- north of Orléans, [234]) and contemporaneous with the onlap onto the Upper Cretaceous in northwestern Normandy.

This heterogenous field of deformation can be integrated with E–W to ESE–WNW extension maily located to the east and with orthogonal N–S to NNW–ESE compression along the Seine fault. Typically, no brittle deformation occurred in the Armorican block.

These deformations are however not of exactly the same age. Extension started in the Rhine graben with the *Sel Inférieur* of Priabonian age (late NP 18/early NP19, [235]). The La Remarde anticline and the northwestern onlap are capped by the Fontainebleau Sands s.l. Its base (*Marnes à Huitres*) is a time-equivalent of the *Marnes à Foraminifères* in the Rhinegraben [235] both deposited at the time of northward longitudinal tilting of the graben still in an E–W exten-

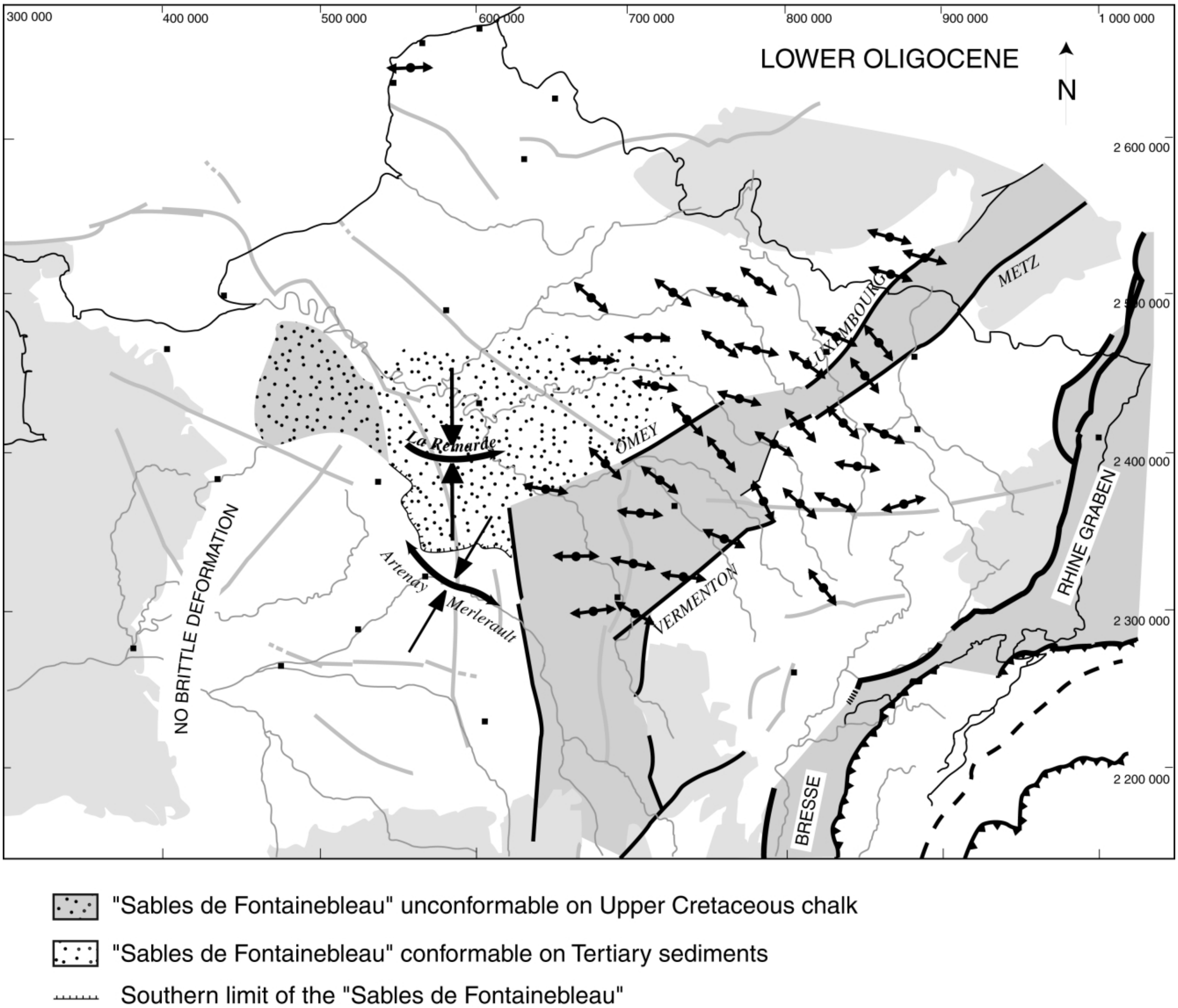


Figure 35. Lower Oligocene deformation field.

sional regime [236, 237]. Around the Lower/Upper Oligocene boundary, subsidence increased sharply in the northern part of the Rhinegraben, whereas in the south only thinner continental sediments were deposited [236, 237]. No sedimentation occurred at that time in the Paris Basin.

The second Tertiary time interval of accommodation space creation (*Sel Inférieur* of the Rhine graben = “Gypse II” of the Paris Basin, [235]) is of the same age as the beginning of the Rhinegraben extension.

Late Lower Oligocene to Middle Miocene: the end of Lower Oligocene time (lacustrine Etampes Limestones) saw a major break in the Tertiary evolution of the Paris Basin, with a medium wavelength tilt, uplift of the northern part of the basin and migration of the depocentre toward the SSW (Beauce area). This trend was accentuated during the Aquitanian (lacustrine Beauce Limestones). At the time of a major low base-level (Lower/Upper Oligocene boundary eustatic sea-level fall, [238]), the Upper Oligocene hiatus and the paraconformable contact of these two lacustrine formations are not explained.

The Burdigalian time was a time of base-level fall with the development of fluvial deposits derived from the south (Massif Central) coinciding with short wavelength deformation in the southern part of the basin (Sologne area). The direction of compression can be estimated from the comparison of the base Tertiary *(figure 30)* and the base Burdi-

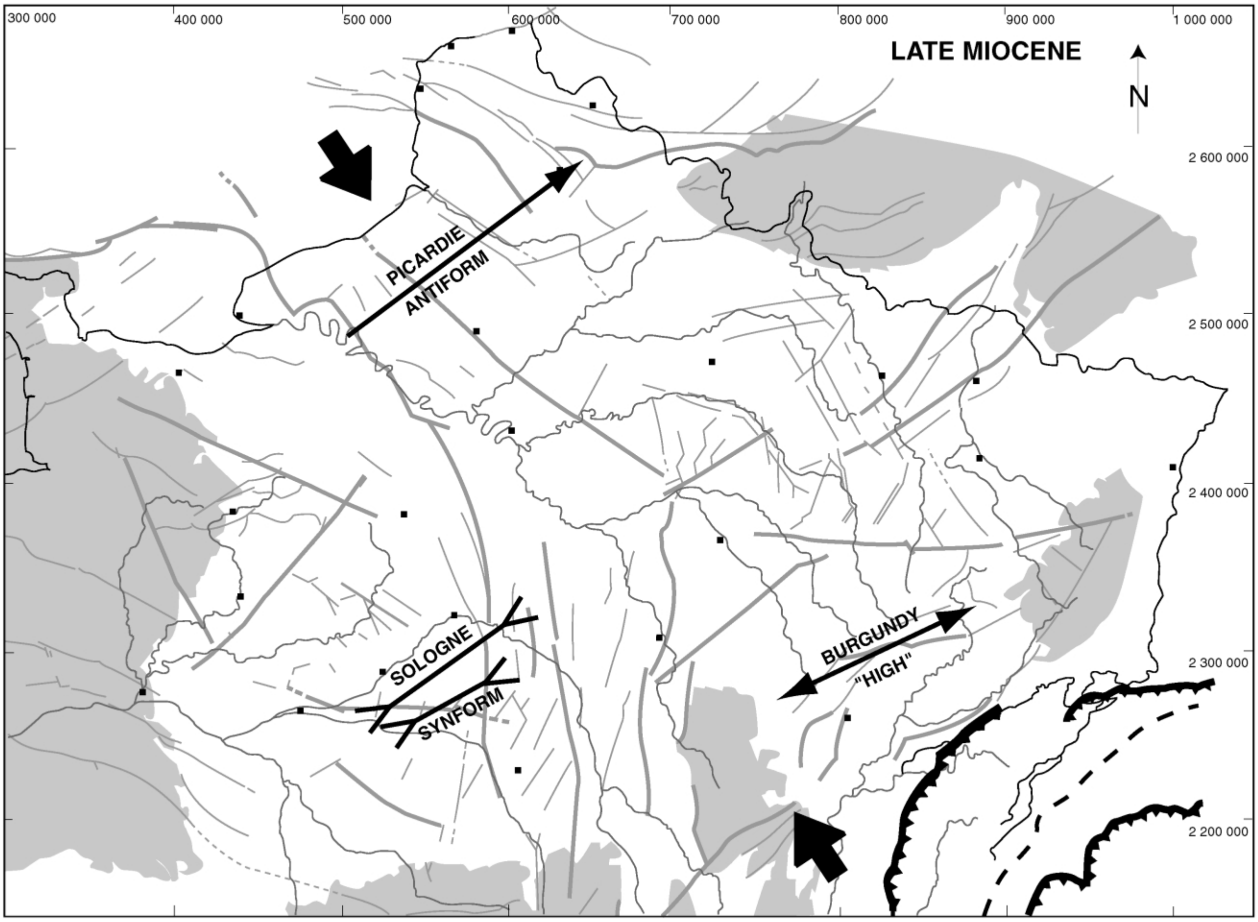


Figure 36. Late Miocene deformation field.

galian isohypses *(figure 31)*. Synclines are orientated along a NW–SE to WNW–ESE axis, which means NE–SW to NNE–SSW compression. This is in good agreement with the NE–SW compression recognized by Bergerat [226]. Westward E–W compressions [229, 227] suggest spatial variations in the field of deformation.

The Langhian-Serravallian marine flooding (*faluns helvétiens*) is partly of eustatic origin (high sea-level, [238]).

Upper Miocene to Lower Pleistocene: this was a time of general sediment by-pass; its beginning, however, is somewhat uncertain (intra-Tortonian). Pliocene rivers cut across the older structures of the Paris Basin. Upper Miocene *(figure 36)* could be the age of formation of (1) the WSW–ENE antiform (*Seuil de Bourgogne* [239, 159] - Burgundy High antiform), (2) of the NE–SW antiform between Rouen and Arras (Picardy antiform), (3) of the post-Burdigalian NE–SW to WSW–ESE Sologne synform and (4) of the post-Serravallian NW–SE fractures of the *Faluns* (Touraine area). No significant short wavelength deformation (folds) occurred.

Middle to Upper Pleistocene: the basin-scale late Lower/early Middle Pleistocene base-level fall may be due to both a major eustatic fall and a general tectonic uplift. These two processes may have operated simultaneously, tectonic activity is proven by the differential incision along the Loire River, for example. This mid-Quaternary tectonic event was identified in the Paris Basin [240–242] and in the Armorican Massif [243–244]. In this area, deformation has been interpreted as related to buckling of the lithosphere in response to NNW–SSE compression [244]. Capture of the Allier and Loire Rivers toward the English Channel during Upper Pleistocene indicates slight tectonic movements. The seismic activity of the Paris Basin [245] is of low magnitude (generally less than 3.4) and located all around the basin: (1) in the northern part along the Variscan front, (2) in the southwest (Anjou, Touraine, Brennes) along Variscan Armorican faults and (3) in the eastern part from the Marne/Meuse Rivers area to the Rhinegraben. The central part of the basin does not show any seismic activity except along some major faults like the Bray fault.

5.5.3. Geodynamic significance of the late Turonian–present-day deformations

Late Cretaceous to Lower Eocene deformation may be the record of lithosphere buckling [246] in reponse to a N–S to NNE–SSW compression.

Middle to Upper Eocene deformation ('Pyrenean deformation'), a more or less continuous process during this time interval, does not correspond to one single phase and looks like an emphasis of the late Cretaceous deformation. The exact fold kinematics are still unknown. These folds were still active during the Lower Oligocene.

Late Eocene/Oligocene extension is only recorded in the eastern part of the Paris Basin, east of the Loire/Loing faults. At that time, the Paris Basin was a transitional area between an extensional domain in the east, i.e. the shoulder of the Rhinegraben rift, and an area of compression to the west. The Oligocene was also the time of inversion of the Western Approaches and Wessex Basins [123, 247]. Surprisingly, little is known about the relationship between deformation kinematics and the basin infill history in both the inverted and rifted basins: at what time was the paroxysm of inversion? What is the significance of the northward tilting of the Rhinegraben?

Early Burdigalian deformation appears to be limited to the southwestern part of the basin. The development of a northward-flowing fluvial system is ubiquitous in other French basins (southeast Alpine foreland Basin - Digne, North-Pyrenean foreland Basin - Aquitaine Basin [248, 109]) which would mean uplift of southern France.

Upper Miocene deformation was contemporaneous with two major tectonic events: a general uplift of France occurring in early Tortonian times, coeval with the restarting of volcanic activity in the Massif Central (Cantal volcanism, [249]), and a NW–SE Late Miocene compression (formation of the Jura mountains, [226]). The NE–SW-trending Burgundy High and Picardy antiforms may be the result of lithosphere buckling in response to Late Miocene compression. The overall sediment by-pass and the capture of rivers coming from the south toward the west (Atlantic Ocean) or the north (English Channel) may be a consequence of low amplitude medium wavelength tilting at a time of general uplift, which is still recognized in the Armorican basement [250].

Similar patterns of fluvial captures have been put forward by Petit et al. [251] with rivers flowing along the Bresse and Rhine grabens. During Upper Pliocene time, the Rhine River was flowing southward to the Mediterranean Sea. This suggests a general uplift of the northern part of France from the Armorican block to the Rhinegraben. This uplift may explain that there is no record of the Lower Pliocene eustatic high sea-level in that area.

6. Conclusions *(figure 37)*

The Meso-Cenozoic evolution of the Paris Basin can be described as a long-term thermal subsidence, inherited from the Permian extension and perturbed by intraplate deformation consequence of the geodynamic events occurring in western Europe, i.e. opening and closing of the Ligurian Tethys and opening of the Atlantic. These tectonic events modified both subsidence and facies distribution in space and time. The Paris Basin is first of all an 'extensional' basin (Triassic–Jurassic) which progressively evolved to a compressional one, first only temporarily (lower Berriasian and late Aptian) and then permanently (late Turonian/present-day).

The **Scythian to Toarcian** is characterized by an arcuate subsiding area along NE–SW and E–W to ENE–SSW directions, by general onlaps in the north and west, and by the creation of accommodation space (5–30 m/My).

Major tectonic disturbances occur (1) at the time of deposition of the Lettenkohle (late Ladinian), (2) at the base *Grès à roseaux* (intra-Norian), (3) during the Norian, (4) at the Rhaetian/Hettangian boundary, (5) at the Hettangian/Sinemurian boundary, (6) in the late Sinemurian, (7) in the Carixian and (8) at the Pliensbachian/Toarcian boundary.

The base *Grès à roseaux* unconformity records the beginning of subsidence in the central part of the present-day Paris Basin.

The Hettangian to Pliensbachian time is characterized by short and medium wavelength deformation which contrasts with the medium wavelength flexural controls of the Triassic and the Toarcian. At that time, the Paris Basin records the different steps of opening of the Ligurian Tethys with a 190 Ma (Pliensbachian) diagenetic 'event' coeval with a strong hydrothermal activity.

The spatial subsidence pattern is consistent with both E–W and N–S extension during this time interval, but no tilted blocks were encountered. Extension is directed E–W to NW–SE during Hettangian to Pliensbachian.

The **Aalenian to Tithonian** is characterized by NW–SE medium wavelength flexural control beginning with some short wavelength controls during the Aalenian/Lower Bajocian and locally during the Bathonian and changing to a high wavelength flexural pattern during the Kimmeridgian/Tithonian times. The kinematics of the Dogger deformation are poorly known.

The high wavelength control is coeval with the highest rate of accommodation space creation (20–40 m/My), with progradational (Callovo-Oxfordian) and aggradational (Kimmeridgian-Tithonian) carbonate platforms and with a 150 Ma (Oxfordian-Tithonian) diagenetic 'event'.

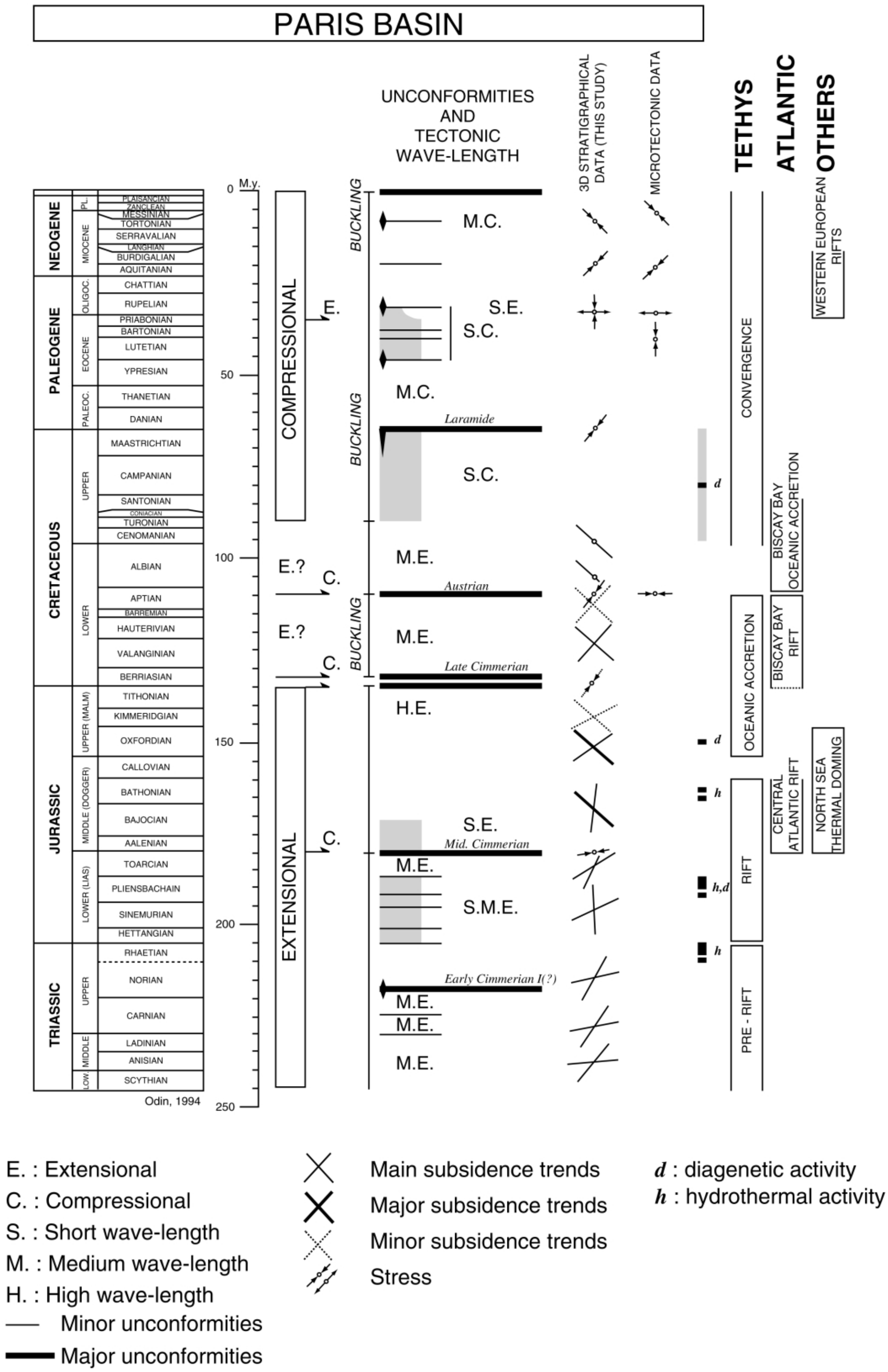


Figure 37. Geodynamic evolution of the Paris Basin.

These changes in both tectonic wavelength and carbonate platforms during the Callovian are contemporaneous with oceanic accretion in the Ligurian Tethys and the transformation of the Subalpine Basin into a passive margin.

The Aalenian unconformity, a major basin-wide discontinuity seems to reflect short wavelength E–W to ENE–WSW compression.

The **Berriasian to late Aptian** is characterized by a NW–SE medium wavelength flexural control bounded by two major unconformities (late-Cimmerian: Jurassic/Cretaceous boundary, Lower/Upper Berriasian boundary; “Austrian”: late Aptian) contemporary with a sharp decrease of the subsidence rates (0–10 m/My) and a change in the sedimentary system with the development of the “Wealden” siliciclastic deltas.

Both unconformities result from NE–SW to E–W compression along the eastern border of the Armorican domain and along the southern border of the Dinant synclinorium (Artois/Ardennes). They are contemporaneous with the early stages of extension in the Bay of Biscay (rift - late Cimmerian) and of oceanic accretion (passive margin - late Aptian). This time was also the first period of lithosphere buckling in the Paris Basin.

The **late Aptian to Turonian** is characterized by a NW–SE medium wavelength flexural control with an increase of the subsidence rate and a change from a siliciclastic system (tidal-dominated Greensands) to carbonate platforms (Cenomano-Turonian chalks) with post-unconformity onlaps toward the west and the northeast.

The **Turonian to recent** is characterized by a decrease of the subsidence rate, sediment by-pass and finally uplift and erosion, in a generally compressional setting. Three main periods of compression can be defined: (1) in the Late Cretaceous (short and medium wavelengths, ‘Subhercynian’ and ‘Laramide’) paroxysm of Senonian deformation, (2) from the Lutetian to the Lower Oligocene (short wavelength, ‘Pyrenean’) and (3) in the Late Miocene (medium wavelength). The subsidence rates sharply decreased in the Late Cretaceous time, going from 10–20 m/My to very low rates (0–10 m/My) in the Lower Oligocene. Sediment by-pass occurred from the Upper Oligocene to the Pliocene, with local deposition along the present-day Loire River. Uplift started in the late Lower/early Middle Pleistocene.

Late Cretaceous and Late Miocene middle wavelength deformations reflect lithosphere buckling in response to N–S and NW–SE compression. Late Eocene/Lower Oligocene extension occurred east of the basin, whereas compression was located in the western part. Minor compression was recorded in the earliest Burdigalian.

The present-day shape of the Paris Basin is the result of compression during the Tertiary and to a lesser degree in the early Cretaceous.

Little is known about the relationships between small-scale faults, medium wavelength flexure and stress regime. Some more work needs to be done in order to better constrain the interpretation of medium wavelength flexure in terms of stress regime.

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