Palaeogeographic evolution of northwestern Europe during the Upper Cenozoic

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ABSTRACT

A re-analysis of the stratigraphy based on recent dating of palaeo-shores of Neogene to early Pleistocene ages is proposed within a geodynamic context in the Channel and Dover Strait areas. This sector of Europe is controlled by two main geological boundaries: to the North, the Variscan Overthrust and, to the South, the northern branch of the southern Armorican Shearing Zone. These two boundaries border a domain that seems to behave rather homogeneously on a large scale controlled by plate tectonics. Since the Paleogene shorelines have been subsiding North and South of this "Channel" region. Episodic uplift largely controlled the opened or closed status of the Dover Strait during late Zanclean, by reactivating Variscan structures. Re-analysis of post-Oligocene sandy formations shows that these regions have suffered long wavelength deformations during the Neogene. These deformations, slightly diachronous from South to North, affect the limits of the Miocene and Pliocene transgressions. The periods of maximum accommodation space for sedimentation are the late Tortonian and the late Piacenzian, both coinciding with tectonic relaxation events. They explain the micropalaeontological evolution of the micropalaeontological fauna of the Channel and southern North Sea during the considered time span.

KEY WORDS
Neogene,
Quaternary,
Dover Strait,
English Channel,
stratigraphy,
palaeogeography,
geomorphology,
neotectonic,
dating.

RÉSUMÉ

Évolution paléogéographique en Europe du Nord-Ouest au Cénozoïque supérieur.
Les formations littorales et fluviatiles sableuses du Néogène et du Quaternaire ancien sont réétudiées dans les régions encadrant le Sud de la mer du Nord et la Manche, en prenant en compte leur contexte géodynamique. Deux domaines sont définis au sein d'une zone limitée au Nord par le front varisque et au Sud par la branche nord du cisaillement sud-armoricain. Les régions situées respectivement au Nord et au Sud de ces limites sont subsidentes depuis le Paléogène. Le secteur interne est soumis à des épisodes de soulèvement temporaires de grande longueur d'onde, légèrement diachroniques du Sud au Nord, qui vont contrôler la géométrie des littoraux et l'ouverture du Pas-de-Calais, vers la fin du Zancléen. Le maximum d'espace disponible pour la sédimentation sableuse est enregistré au Tortonien supérieur et au Piacenzien final, en correspondance avec les épisodes de relâchement des contraintes tectoniques. Les relations fauniques entre la Manche et la mer du Nord se trouvent ainsi expliquées.

MOTS CLÉS
Néogène,
Quaternaire,
Pas-de-Calais,
Manche,
stratigraphie,
paléogéographie,
géomorphologie,
néotectonique,
datation.

INTRODUCTION

Considering a regional scale, palaeo-shorelines provide evidences of long-term eustatic changes and help to understand the tectonic movements. Raised shoreline anomalies are found in the Channel region and may result from deformations induced by tectonics.

In western France, the Redonian Crag or Faluns and the Red Sands are Mio-Pliocene deposits. The same period is covered in Belgium by the Diest, Kattendijk and part of the Lillo formations (Luchtbal Crag). In the Netherlands, it corresponds respectively to the Breda and Oosterhout formations, while a correspondance is established with Coralline cragpartity. Between these main regions, sporadic deposits lead to stratigraphic controversies.

Most of these shore formations consist of decalcified coarse clastic or sandy bodies located at various altitudes, ranging from 3 to 190 m, often on a low platform. It is now quite obvious that elevation cannot be used in any chronological sense, but this does not mean that elevation is not important. It is by mean of elevation differences of well-dated units that tectonic events can be identified when compared with available eustatic curves (Haq *et al.* 1988; Hardenbol *et al.* 1998). On the marine shelf, platform shaping has limited the preservation of Neogene bodies, troughs excepted. Although a few oil exploration drillings have been performed at the shelf break in the Channel Western Approaches, neither the Hurd Deep, nor the Fosses Dangeard have been deeply cored. The only available marine information comes from the southern North Sea in the Murray Pit. Thus Neogene is lacking high resolution data in this region.

This study is based: 1) on the morphostructural analysis of every site in relation to preserved sedimentary wedge; 2) on the pedosedimentary and sequential record in these formations; and 3) on palaeontological dating combined with ESR (Electron Spin Resonance Spectroscopy) on quartz and ^{87}Sr/^{86}Sr dating of shells or bones. North of the Dover Strait, the various steps of the transgressions have been reconstructed on the basis of micropalaeontology (dinoflagellate cyst stratigraphy), owing to the non-oxidized status of the Diest Formation. The boundaries defined by these approaches emphasize the part played by tectonic reactivation of the Variscan substratum.

METHODS

Systematic microstratigraphic work has been carried out since 1994 on both sides of the Channel on previously described sections, in extending sand pit exploitations both in Brittany and England or in new temporary excavations related to the development of highways or industrial zones, and to the Dover Strait Tunnel realisation. Of particular interest is the increased erosion of shore cliffs as it provides higher and more complete outcrops than in the previous 30 years, especially in the Boulonnais. Some of the observations were obtained from drillings performed during the geological mapping along the French side of the Channel as well as in the Ardenne. Altitudes are given as OD (Ordonance Survey) for the British sites and NGF (French Geodetic levelling) for French sites. Palaeopedological survey and new ESR dating corroborate this research. Special attention has been paid to the beach/cliff/valley-incision geomorphic system (Figs 3; 4).

The sediments were dated by the ESR method performed on quartz and involving aluminium colour centres. The residual palaeodose is measured after exposure of 10 aliquots of quartz under black light (365-460 nm). This methodology follows the procedure published by Laurent (1993) and Laurent *et al.* (1998). Dating was systematically associated with petrographic analysis in order to detect abnormal bleaching possibilities, linked to important clay or organic interstitial accumulation and drainage story of the deposit. Nevertheless, an accuracy no better than 10% is the rule so that dating is indicative, rather than true stratigraphic diagnostic

Micropalaeontological boundaries are those defined by Berggren *et al.* (1995). The eustatic and stratigraphic reference used is the Hardenbol *et al.* (1998) chart, which integrates most of the previous works (Haq *et al.* 1988). Tectonic phase names are those defined in Haq & Van Eysinga (1998). Palaeopedology is integrated as a tool in sequence stratigraphy (McCarthy & Plint 1998).

GEODYNAMICAL CONTEXT (FIG. 1)

The main characteristics of the geological substratum will be briefly reviewed in order to understand the Channel basement behaviour. Cainozoic compressional foreland deformations in western Europe developed in a fore-arc setting with respect to the Alpine subduction system (Ziegler 1992). All the sites investigated in this

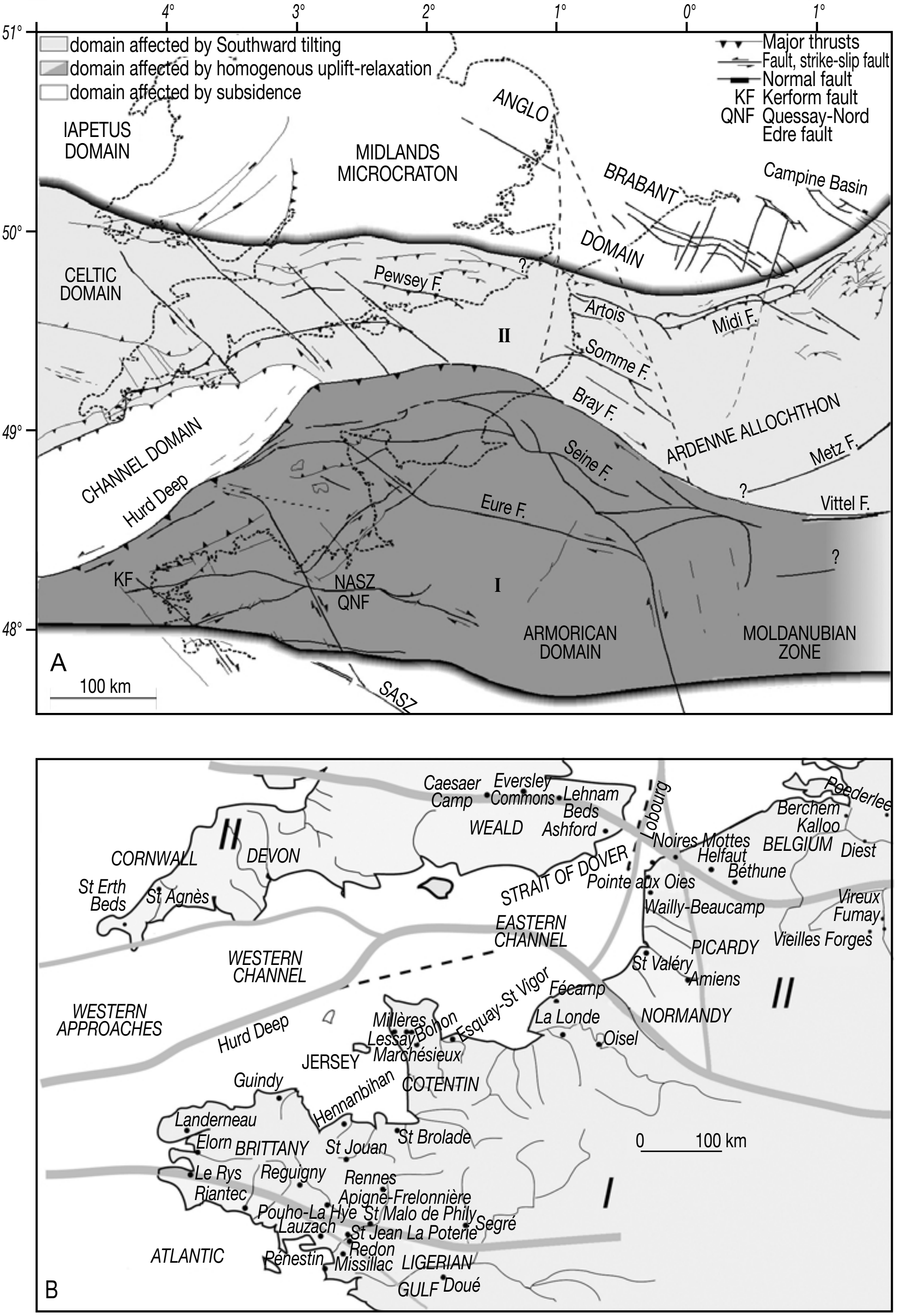


FIG. 1. — **A**, map of the geodynamic domains from the investigated region of western Europe, modified from Mansy *et al.* in press; **B**, location of the main sites.

study are located between two major Variscan lines, the northern Variscan front overlapping the Anglo-Brabant massif and the northern branch of the southern Armorican shear zone (SASZ). Jurassic to Cretaceous N-S extension along Variscan accidents induced the formation of basins, which were inverted by steps since the late Cretaceous-Paleogene. After the Paleocene, the zone located north to the Variscan overthrust has been subsiding.

The Dutch Rhine graben and the North Sea basin have also been subjected to accelerated subsidence since the Plio-Pleistocene in response to the Neogene-present-day compressional stress field in western Europe (Cameron *et al.* 1992). This movement has been accentuated by huge sediment import from the Alps into the North Sea basin. The southern boundary of the area, including southern Brittany, the entrance of the Channel, Cornwall and Scillies, belongs to the Western Approaches of the Channel system and is subsiding by steps as a passive margin and as a side of the Western Channel "graben" since the opening of the Biscay gulf (Evans 1990) during early Cretaceous.

When a tectonic compressive event occurred in the Cainozoic, at the southern boundary of the European plate, the Channel area was submitted to a generalized, long wavelength uplift (Fig. 1, domains I-II). These events occurred during the Eocene (Wyns 1991) and during several phases of Neogene and Quaternary (Van Vliet-Lanoë *et al.* 1998a, b), from 7 to 5 Ma and around 1-0.8 Ma. After each compression, differential movements occurred during relaxation with increased individual block dynamics and micro-basin sedimentation. These movements were controlled by the fault geometry in the upper crust as stressed by Matte (1998) and Mansy *et al.* (in press). A second flexural deformation has been activated at least since Oligocene time in conformity with the fault system of the Channel graben (Hurd Deep). The excavation of the main palaeo-cliff seems to be the result of a progressive regional uplift related to the Attic phase (Had & Van Eysinga 1998), a tectonic phase more or less synchronous with the late Miocene (Van Vliet-Lanoë *et al.* 1998b), in combination with rising sea-levels from 10 to 7 Ma (Haq *et al.* 1988; Hardenbol *et al.* 1998). Secondary N-S extensional movements occurred during the Pliocene relaxation event with mainly local subsidence on the southern coast of Brittany, the North of Dover Strait or along the Hampshire basin and Picardy coasts. These movements led to more or less W-E oriented basins filled with "Red Sands". Some of those as the Frelonnière in the Rennes basin were inverted during the Early Quaternary tectonic phase.

The Variscan front was uplifted by steps such as in the case of the inverted Wealden basin, in the southern part of the Isle of Wight and in the Bray (Ziegler 1992; Underhill & Patterson 1998; Everaerts & Mansy 2001; Mansy *et al.* in press). The Somme and the Seine are synforms that are being squeezed between the actively inverting Bray and Boulonnais zones and uplifting Armorican massif. The deep structural geometry of the Variscan faulted substratum (Fig. 1, domain II) seems to induce a southward global tilting of the northern domain, accentuated, from the onset of upper Neogene by the formation of the Jura belt (Kalin 1997), South-East to the Paris basin with drainage modifications, river or sea capture in supplement to the allocyclic tectonic uplift/relaxation and to the third order eustatic variation.

HISTORY OF THE CRAG AND EVOLUTION OF THE REGIONAL PALAEOGEOGRAPHY

Alvinerie *et al.* (1992) have recently proposed a set of palaeogeographic maps for the Atlantic shore face from the Chattien up to the Pliocene. Those maps have sustained the palaeoclimatic interpretation of the fauna assemblages (Lauriat-Rage *et al.* 1993) which record a progressive cooling during the Neogene with a North-South latitudinal gradient increasing with time.

With regards to palaeontology, many studies have been performed on Neogene crag-rich formations in the Ligerian and Normandy regions. The sequence starts with two transgressions, Aquitanian and Burdigalian, poorly expressed in Anjou (oysters shells). Helvetian Crag deposits

(Pontilevian and Savignean "faluns") cover more or less the Langhian-Serravallian highstands. The rich malacofauna of these Crag, widely spread, represent the middle Miocene. They are also found in Cotentin (Normandy: Crag of Béhou). These middle Miocene biogenic deposits are often unconformably covered by more recent Crag, the Redonian deposits, sharing partly some species with the middle Miocene, which leads sometimes to stratigraphic ambiguities. According to Dollfus (1900) who defined molluscan assemblages, it is clear that the Redonian Crag are younger than middle Miocene in Touraine, Anjou, Brittany and Cotentin. Dollfus (1900) located the Redonian in the upper Miocene, mostly in the Tortonian. Since his works, many studies have been made by: 1) intercomparison with earlier fauna (extinction, renewal, survival, local endemism); 2) regional differentiation with the Redonian (stratigraphic, palaeobiogeographic and palaeoecologic criteria); 3) intercomparison between fauna of the same age in neighbouring provinces; 4) dinoflagellate stratigraphy datation (e.g., Head 1996) of the British Crag and the Lehnam Beds.

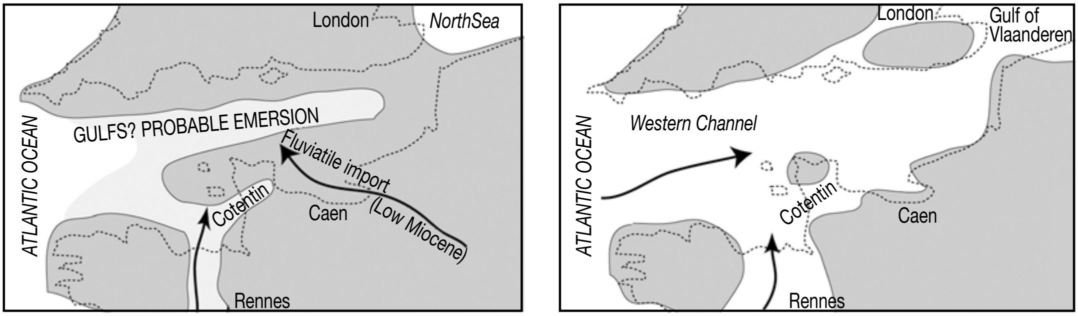
According to Brébion (1964, 1970), Redonian gasteropods of the Ligerian Gulf record three independant transgressions without marked faunal change within the upper Miocene: 1) the Redonian of Anjou is the oldest; 2) the Redonian of lower Loire river and Vendée; 3) the youngest one, the *Nassa* sp. marls of Redon. Fécamp Crag belong also to the upper Miocene. Still for this author, the "Redonian" of Normandy belongs to the Pliocene (Pareyn *et al.* 1984) (see Table 1).

According to Lauriat-Rage & Vergnaud-Grazzini (1977) and Lauriat-Rage (1981, 1982), Redonian bivalves belong already to Pliocene and are diachronous, controlled by the progressive lowering in temperature of marine waters since 3 Ma. In order to define more accurate palaeontological zones a new approach has been further developed on the base of isotopic analysis of oxygen and carbon on shells. On this base, warm (Anjou), mild (some of the Anjou and lower Loire beds) and cold (Anjou, Oléron and Brittany) assemblages were defined. The Redonian deposits represent a major transgression with secondary pulses (probably equivalent to Oxygen Isotopes Stages), without any important discontinuity and with transitional faunal associations. Equivalent associations were recognized in the basins of Carentan and St-Sauveur-le-Vicomte in Normandy (Lauriat-Rage 1986; Garcin *et al.* 1997).

Controversies about the interpretation of the malacofauna stem for the survival of a high percentage of middle Miocene species, also by an enhanced endemism for the gasteropods during the Redonian, and finally, by the occurrence from the Pliocene of Nordic bivalve species. Recently, $^{87}Sr/^{86}Sr$ isotopic dating, performed on bones and teeth (Barrat *et al.* 2000) and on shells of bivalves (Mercier *et al.* 1997) from the Redonian of the Ligerian Gulf, raised new attributions. Three groups are defined: upper Tortonian-lower Messinian (Anjou), lower Messinian (Anjou, Brittany), late Messinian-lower Pliocene (Anjou, Maine, lower Loire). The main result is the aging of the Redonian deposits and the correctness of the oldest attribution given by the malacofauna data. The deposits of Normandy still yield a Pliocene age.

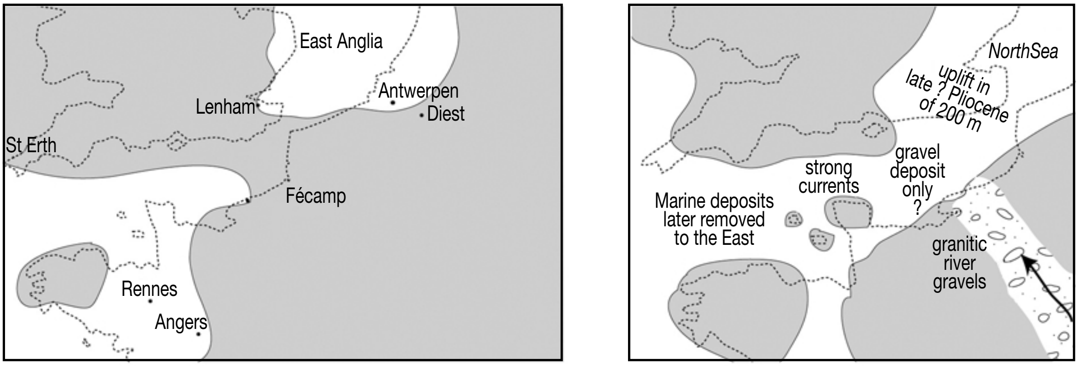
The Channel is characterized by an extended marine abrasion surface. Locally superimposed on much older surfaces of Mesozoic or Permian ages, Langhian-Serravallian Crag and older deposits have been eroded mainly in the Anglo-Norman gulf and the Hampshire basin. Lower Miocene is mostly lacking in this region but is preserved at the shelf break of the Western Approaches as the thin Jones Formation covering the late Oligocene to the middle Serravallian (Evans 1990).

For many authors the evolution of the Channel is still controversial. Margerel (1968) showed from foraminiferal assemblages that the Dover Strait was closed until the late Miocene but open during the Pliocene by comparison with Glibert's (1962) observations. Following this work, a first palaeogeographic reconstruction was proposed by Larsonneur in 1971 (Fig. 2, top), indicating a progressive eastward expansion of the Channel



Middle Miocene (Helvetian)

Pliocene (Redonian)



Pliocene

Mio-Pliocene

FIG. 2. — Palaeogeographic maps of the Channel and southern North Sea, historic of the concepts. Middle Miocene (Helvetian) and Pliocene (Redonian) after Larsonneur (1971); Pliocene after Pomerol (1973); and Mio-Pliocene after Curry (1993).

Gulf during the upper Miocene and a breaching during the Pliocene. He also suggests a western communication between the southern North Sea and the Channel in the region of Aldershot, West of the Weald. Re-using the same data and new analysis by Bassompierre *et al.* (1972) and Laga (1972), Pomerol (1973) proposed a slightly different map (Fig. 2) with a closed Dover Strait during the Pliocene, emphasizing the development of the Ligerian, Rennes and Cotentin basins. In agreement with the conclusion of Margerel (1968), Lauriat (1973) has observed common shell species between the North Sea and the Channel but supposed that the exchange was driven from the North around the British Isles. In 1975, Dingwall published the occurrence of Pliocene deposits in the northern prolongation of the Lobourg Deep which incised the Channel. In 1988, Gullentops also proposed that the Channel was open during the Diestian. More recently, Margerel (1989) showed by interregional comparison from the SW to the NE that the Dover Strait opened during the lower Pliocene, cold species migrating towards the Armorican regions from the NE (Table 2). Curry (1992) proposed a slightly different picture from Pomerol's: an open strait during the Pliocene (Fig. 2). The Dover Strait was thus clearly open during the Pliocene (Margerel 1968, 1989; Meijer & Preece 1995). This is probably related to an initial opening of the Dover Strait following the Messinian incision (Van Vliet-Lanoë *et al.* 1998a, b) (Fig. 3).

In Belgium, the Diest sands (Table 1) are as famous as the Red Sands of Brittany or the English Crag. Studies on Neogene have been

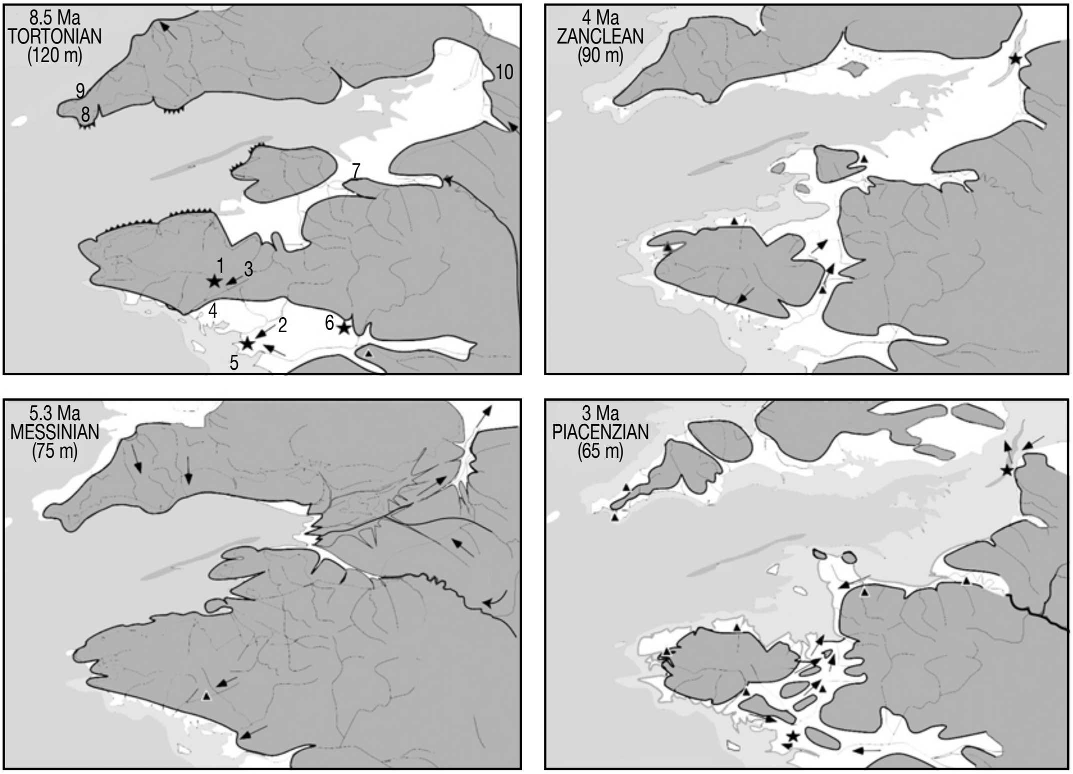


FIG. 3. — Palaeogeographic maps of the Channel, modified from Van Vliet-Lanoë *et al.* 1998a, b. Notice the restricted flooding during the late Messinian and the Zanclean.

undertaken fairly early in Belgium and based on bryozoa (Lagaaij 1952), molluscs (Heinzelin 1955) and invertebrates (Glibert 1962). All these authors claimed a late Miocene age rather than a Pliocene one for the Diest and Deurne sands. This attribution has been later confirmed by De Meuter & Laga (1976), Louwye & Laga (1998) and Louwye *et al.* (1999, 2000), allowing a better differentiation between a late Miocene Diest Formation and Pliocene deposits.

Despite these data, Smith (1985) and Gibbard (1995) among others have suggested that the initial "official" breaching of the Strait occurred during the Anglian (Oxygen Isotope Stage 12 or 0.42 Ma only, in their view), although the breach was prepared by a graben during Early Quaternary (Colbeaux *et al.* 1993), a hypothesis which is inconsistent with the geology of the strait zone and with malacological data of Meijer & Preece (1995). Kellaway *et al.* (1975) argued for a glacial tunnel valley, especially at the location of the "Fosses Dangeard". Alduc (1979) rejected this glacial hypothesis because of the absence of glacial evidence in the vicinity of the Dover Strait.

In 1996, in the light of recent dating of the Mio-Pliocene "Reds sands" of Brittany, a re-analysis of the Pointe-aux-Oies section in the Boulonnais (Van Vliet-Lanoë *et al.* 1998b) has led to a new interpretation of the Strait opening, confirming the palaeontological data. This event seems to have resulted from the regressive capture of the western Channel by the North Sea in relation to a temporary Messinian uplift of the Variscan front (Van Vliet-Lanoë *et al.* 1998b) by a Pliocene Lobourg river (Dingwall 1975), located on a fractured N-S lineation crossing the Variscan front (Fig. 3). The Zanclean and

TABLE 1. — Correlation between the new stratigraphical chart and the former formation nomenclature.

	Brittany-Ligerian	**East Anglia**	**Belgium**	**Netherlands**
Gelasian		Red Crag	Scaldisian	Tiglian
Piacenzian	Redonian cold	Coralline Crag		Reuverian
Zanclean				Brunsummian
Messinian	Redonian warm	Lehnam Beds	Diestian	
Tortonian				
Langhian-Serravallian	Helvetian			

Piacenzian highstands controlled this breaching (Van Vliet-Lanoë *et al.* 1998b). Murray Pit infilling (BGS Borehole 81/51) was considered to be early Pliocene in age, synchronous with the Belgian Luchtbal sands (Balson 1989), on account of the occurrence of phosphatic levels. The Luchtbal sands contain benthic molluscs, foraminifera and otholiths of middle and upper Zanclean age (Vandenberghe *et al.* 1998). The strait was closed once since from the Early Quaternary (Meijer & Preece 1995).

Concerning the Plio-Pleistocene times, the detailed work of Zagwijn (1989) and Funnel (1996) have brought complementary data, especially on the position of the successive shore lines in the southern North Sea basin. These data have been taken into account for the drawing of Fig. 4. Morphology of the basement of the Channel (Auffret *et al.* 1980) and a Digital Elevation Model of the whole region have been used to precise the palaeoshore-line boundaries.

According to malacofaunal evidences, the Strait remained closed until O.I.S. (Oxygen Isotopes Stage) 5e (Meijer & Preece 1995). From a geodynamical point of view, Herzeele deposits (O.I.S. 7 or 9) belong to the North Sea basin and not to the Dover Strait Flexure Zone such as seen in the Sangatte section (Van Vliet-Lanoë *et al.* 2000a).

During the Eemian (O.I.S. 5e: 130-110 ka), the Strait had its maximum width, probably from the re-excavation of the pre-existing Pliocene Lobourg Deep. Its opening was made easier by its soft sedimentary infilling (beach gravel ridge included), caused by potential stronger tidal currents (Scourse & Austin 1995) related to higher sea levels up to + 6 m above the Holocene levels. Normal faults along the Sangatte cliff argue for tensional subsidence of at least 5 m, North of the tunnel flexural zone at about 160-170 ka (pedostratigraphy) and after the Eemian, the deposits of which are now drowned at Sangatte (Van Vliet-Lanoë *et al.* 2000a).

REGIONAL STRATIGRAPHY

Brittany (Figs 4-6)

An important N-S shortening tectonic episode related to the alpine orogeny occurs at the end of Oligocene, stopping sedimentation in the basins (Ziegler 1992). At that time, valleys system preexists. From the onset of the Neogene, Brittany is uplifted and new valley incisions occur in relation with Chattian regression, mostly in the southwestern part of the region.

During the middle Miocene, western Brittany already emerges and the Langhian-Serravallian transgressions remain limited to the new valley system despite a relatively high sea level (*c.* 200 m, Hardenbol *et al.* 1998). The preserved deposits are mostly faluns (or "Crag") and clays, in a warm subtropical climate, the former "Helvetian" (Durand 1960, 1968; Pomerol 1973) (Table 1). The maximal inundation is reached since the Burdigalian with a maximum during the Langhian-Serravallian (Chasné-sur-Ilet clays, Margerel & Breheret 1984).

In the Armorican Massif, the widespread Red Sands facies has been formerly attributed to an upper Pliocene marine environment. They consist of prograding bodies of coarse azoïc quartz sands with some glauconite, usually preserved in a flooded palaeovalley system (ria).

In the upper valley of the Blavet river (Reguigny, central Brittany), fluviatile and estuarine deposits from an ENE-WSW shallow basin, crop out at the

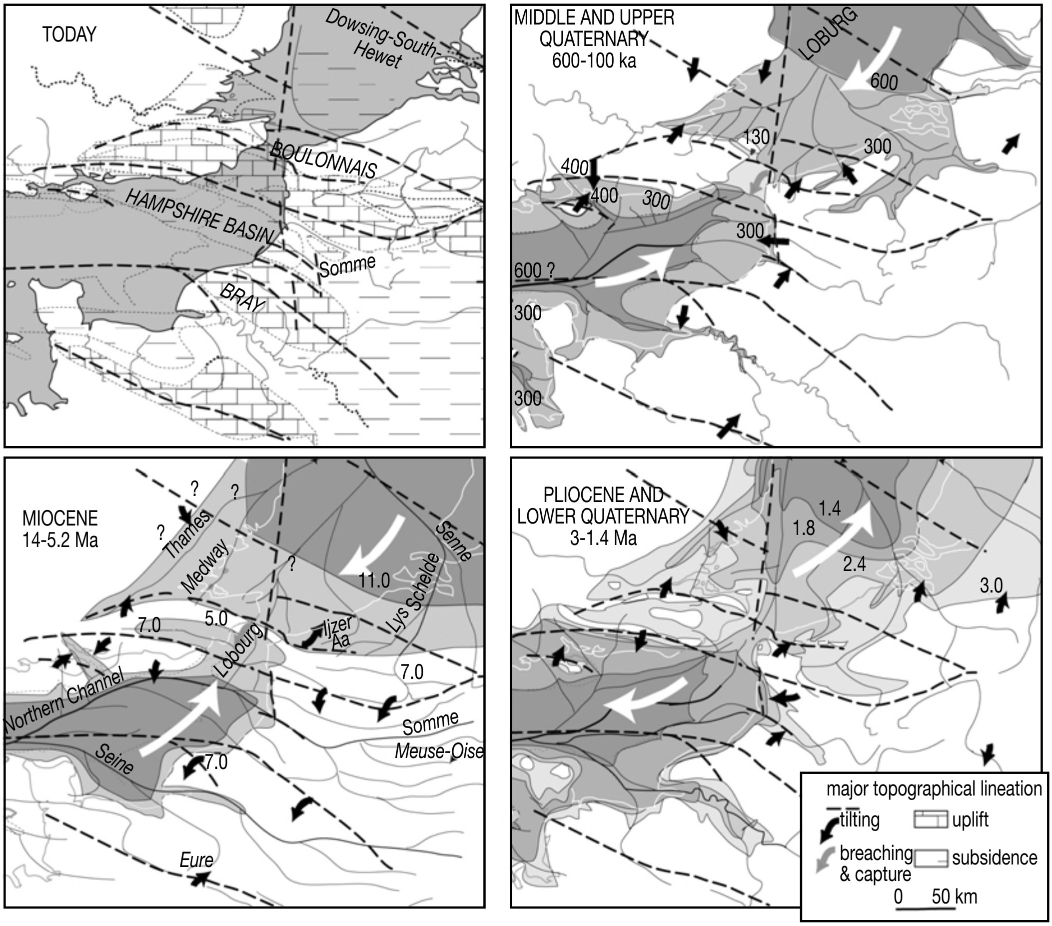


Fig. 4. — Palaeogeographic evolution of the Dover Strait region from late Miocene to Recent. Including data from Zagwijn 1989 and Funnel 1996 for the North Sea. Today structural boundaries (**-**, domains in subsidence; **bricks**, domains in uplift/inversion). Numbers indicate the age of the drawn shorelines in Ma and ka. Notice the evolution of the river pattern and the parallelism of the transgression pattern in the southern North Sea in relation the uplift-subsidence cycles.

quarry of Reguigny. ESR dating (exclusively performed on quartz), sedimentology, sequence stratigraphy, palaeopedology and microtectonic studies were performed on these sediments in which fauna and palynomorphs are usually lacking. Complementary analyses have been made on similar deposits from South Brittany (Rieux, Lauzach, Missilac) and from central Brittany (Rennes, Ploermel). Fluviatile sands originate mainly from local saprolites with some aeolian reworking and to the East from presumed Cretaceous sources. Several formations were defined at Reguigny (central Brittany) and recorded in most of the sites (Van Vliet-Lanoë *et al.* 1998a).

After the "Helvetian" period, the climate cools down and a large regression occurs at the base of the Tortonian, around 10.5 Ma. A new valley system is excavated further on the shelf break of the Western Approaches platform into the form of a large fan, the Cockburn Formation (Evans 1990). No channel incision is observed on the South Armorican shelf. At that time, Brittany is probably still residually uplifted: the Tortonian transgressions (*c.* 120 m, Hardenbol *et al.* 1998)

TABLE 2. — Faunal evolution in the Channel zone during the late Neogene. Modified from Margerel 1969, 1989; Pomerol 1973; Meijer & Preece 1995; Zagwijn 1989.

Global stages	Brittany Cornwall	Normandy	North Belgium
Eemian	**Malacofauna with dominant Lusitanian affinities**		
Lower to Middle Pleistocene (late Tiglian up to S.I.9)	Cold temperate Malacofauna with some Lusitanian affinities		Malacofauna with Nordic affinities (Celtic) Glacial / Interglacial
Early Pleistocene = **Eburonian + Tiglian C5** (mild on continent) cold temperate (Foraminifera)		Cold temperate	Cold temperate Glacial / interglacial
Gelasian (1.8-2.6 Ma) = **Tiglian A-C4 +** **Pre-Tiglian + Reuverian B-C** **Late Piacenzian**	FBN9 *Elphidium ottmanni* *Aubignyna mariei marei* **Temperate**	**Pliocene 3** *Elphidiella hannai* **Cool temperate**	
Piacenzian (3-3.6 Ma) Temperate	FBN7 *Pseudoeponides* *Pseudotepidus* *pseudot.*	**Pliocene 2** + *Buccella frigida* **FAUJASINA** *cooler*	BFN6 *Elphidiella hannai* *Cribrononion* *excavatum* **Cold temperate**
Zanclean (5.3-3.6 Ma) **warm**	foraminifera with Nordic affinities *Polymorphina frondiformis* *Elphidium paraskevaidisi*	**Pliocene 1**	BFN4 *Florilus boueanus* *Monspeliensina pseudotepida* With remains of Lusitanian Malacofauna *(Palm trees)*
Late Miocene warm to tepid Channel Foraminifera: Lusitanian affinities Malacofauna: Lusitanian			Foraminifera with Nordic affinities Lusitanian Malacofauna

invaded most of southern Brittany and the Ligerian Gulf. Here, the oldest transgression is recorded at Doué-La-Fontaine (Biagi *et al.* 1996) and the Sr isotopes results obtained on fish and marine vertebrates suggest an age of 11 Ma for the host sediment (Barrat *et al.* 2000). These transgressions are recorded in central Brittany by the setting of the Bolan Formation, a Tortonian fluviatile (with low sinuosity) to estuarine clayish sand body, showing at least two transgressive events, dated by ESR at 8.7 ± 1.5 and 7.0 ± 1.0 Ma (Fig. 5). Synsedimentary low-level seismicity is commonly recorded. It crops out mostly in South Brittany (Reguigny 80-90 m, Missillac 20-30 m, Lauzach 20-30 m, Riantec 15-20 m, St-Malo-de-Phily 30-40 m, Pénestin 5-8 m), west Brittany (Le Rhys 0-8 m), and contains generally authigenic glauconite and abundant pyrite ghosts. At Saint-Jouan-de-l'Isle and Reguigny, extension is synsedimentary recorded. It is at many places subsequently deformed by transpressional faulting. At Le Rhys, close to Douarnenez, the shore deposit rests immediately on the rock platform, in association with a palaeocliff covered by periglacial slope-deposits and associated with shore ice rafted blocks lying flat on the surface of the platform. The situation is similar at Pénestin, where the beaches-cobbles are lifted up by frost jacking. These two outcrops represent probably lowstands while Reguigny, St-Malo, St-Jouan represent sites of maximal inundation.

This Bolan Formation is truncated by a well-developed palaeosol (tropical podzol), consolidated by a goethitic iron pan underlining an incision (regression to *c.* 70 m, Hardenbol *et al.* 1998). This iron pan seals evidences of normal faulting at Reguigny. This palaeosurface is covered by coarse fluvial sands and gravels, the Reguigny Formation, attributed to the Messinian and dated of 6.5 ± 1.0 Ma at Reguigny. It consists generally of braided river deposits displaying a rather steep slope, alluvial cones and the reworking of fresh and weathered blocks of the substratum, implying strong floods. To the top, the stream dynamic evolves into sinuous system. It is associated with syngenetic neotectonic and palaeoseismicity record, the Post-Helvetian crisis of Durand (1960). Regional stress field is transpressional (150° N). Again Brittany is uplifted by comparison with the eustatic curve whereas the zone south of the SASZ is subsiding.

Because of this uplift, the Zanclean transgression, considered as the first one of the Pliocene, is only locally known in Brittany: in the Rennes basin at Apigné. The classical "Redonian" Crag of the Apigné Formation (Durand 1960, 1968; Margerel 1968, 1989; Pomerol 1973; Lauriat-Rage 1975, 1981) are the classical facies. A calcareous mudflat facies containing Zanclean foraminifera (La Frelonnière Formation that rests immediately above the crag, with a sandy tidal channel) yields an ESR age of 5.35 ± 0.8 Ma (Van Vliet-Lanoë *et al.* 1998). For this reason, the Apigné Crag may be attributed to the previous highstand. New observations in western Brittany, at Landerneau and Brest (Hallégouët *et al.* in press), show also the presence of Zanclean infra-tidal organic clays resting on an oyster crag in the Elorn valley, at Landerneau; this formation is overlain by another marine infra-tidal clay body of presumed Piacenzian age itself flooded by a Gelasian infratidal clay (Morzadec-Kerfourn 1997), before being incised by Early Quaternary fluvial braided deposits. Zanclean sandy facies are now also recognized in southern Brittany at Lauzach and Missilac on the main platform, South of the SASZ (southern shearing zone) and exists probably at Segré.

The Radenac Formation, the "classical Red sands", which also floods a ria system, truncates these three formations. The sands are normally light coloured and have a low content in reworked glauconite. They are mostly derived from earlier formations and saprolites and are devoid of authigenic glauconite, but contain a rather high proportion of aeolian grains. Traces of microfossils are observed in the Rennes basin in the last transgression. This formation corresponds to a Piacenzian fluviatile complex including three main estuarine transgressions, the setting of which is partially controlled by neotectonic and low level seismicity; it yields ESR ages at 3.8 ± 0.5, 3.5 ± 0.7, 3.3 ± 0.5 and 3.05 ± 0.5 Ma (Fig. 5). Deposits are first coarse braided fluviatile with a rather low sinuosity and sometimes subject to flood. They shift to internal estuarine conditions dominated by continental flow toward the end of the sequence. True shore face formations only exist at Missilac (30 m), from Quiberon to Riantec (30 m) in southern Brittany, at Pénestin in the form of meandering tidal channel and at St-Brolade (60 m) on the northern coast, forming a coastal bar. At Landerneau, in the Elorn valley and in Tregor, the platform is still invaded by deep water. Flooding surface is much more extensive than during the Zanclean (*c.* 120 m, Hardenbol *et al.* 1998). The Leon, which was emerged during the Zanclean, becomes submerged (Hallegouët 1971) despite a lower highstand (*c.* 100 m). Brittany is relaxing and watersheds extended further North, as visualized by Piacenzian sands progradation near Josselin (Oust river), in the Rennes basin and close to Bédé and St-Jouan.

Around 2.6-2.4 Ma age, the sedimentation changed abruptly, in relation to the global main cooling, but also with the onset of a new tectonic crisis associated with block tilting and local uplift. Similar stratified estuarine clays at Guindy (Tregor), at Landerneau (Elorn valley) and other sites of the western coast of France constitute a

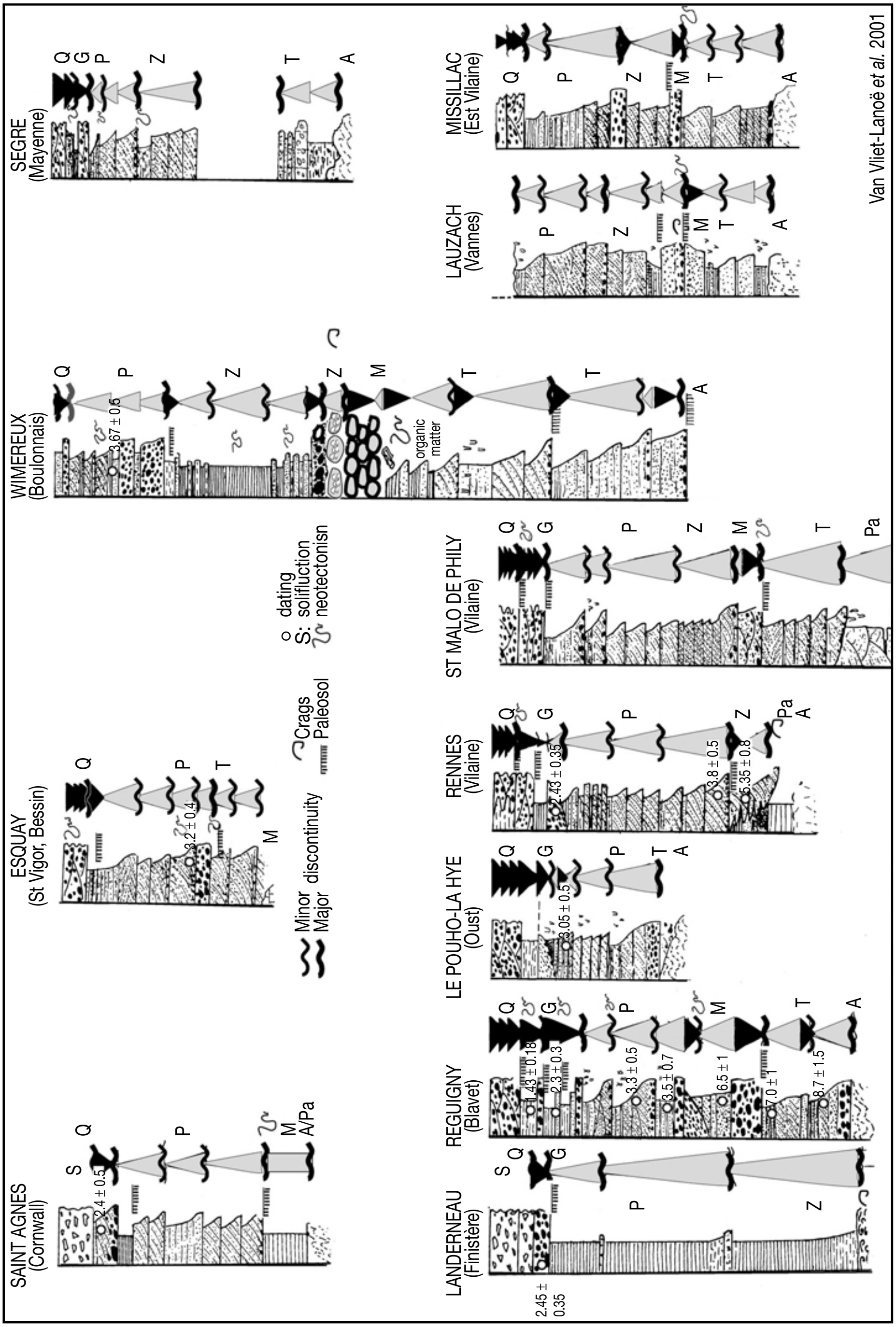


FIG. 5. — Stratigraphical logs for Brittany, Eastern Channel and Dover Strait region, with location of the ESR dating. Retrogradation events (transgressions) in gray; progradation events (regression) in black. Ages in Ma. Abbreviations: **A**, weather; **G**, Gelasian; **M**, Messinian; **P**, Piacenzian; **Pa**, Paleogene; **Q**, Quaternary; **T**, Tortonian; **Z**, Zanclean.

rather homogeneous assemblage (Morzadec-Kerfourn 1997), foraminifera excepted (Margerel 1968, 1989). ESR dating of equivalent clayey units (fluviatile) yield an age of *c.* 2.3 ± 0.3 Ma at Reguigny and 2.43 ± 0.35 at La Frelonnière (Rennes) (Fig. 5). Sands in the Elorn valley, East of Brest overlay continental deposits with scattered blocks, incising the lower clays of the Zanclean. These deposits include some granitic rocks from the submerged Léon Plateau. An ESR dating of the sands yielded also an age of 2.42 Ma ± 0.35. At Mesquer, a gravel level younger than the clay yielded an ESR age of 2.02 ± 0.3 Ma.

After this clayish formation, regional sedimentation is discrete. Even since Middle Quaternary, erosion prevails, Early and Middle Quaternary neotectonic is important, modifying drastically the valley system. Meandering valley alluvium is dated of the Pre-Tiglian, but the first periglacial braided rivers deposits, with ice rafted blocks, date back from the Waalian (ESR 1.43 ± 0.18 Ma, Reguigny).

Brittany evolution throughout Neogene can be summarised as follows (Figs 4; 6): a shift of prograding sedimentation on the shelf related to the successive uplifts (long waves deformations) of Brittany from late Paleogene up to now, with an acceleration from the base of the Tortonian (11 Ma). A slight tilting of Brittany towards the South is observed during the Tortonian/ Messinian. During the relaxation phase of the Piacenzian, the tilting is reduced by a lifted SASZ and so that some of the drainage of the Oust and upper Vilaine reached the northern coast before shifting back towards the South during the early Gelasian. This event is only recorded from flow directions but no incision seems to have occurred in the relaxing context.

Normandy and Picardy

Complex sedimentation is discretely preserved, usually W-E trending graben in the Cotentin Pass and in the Bessin (Bayeux). These have been known for a while and have provided the basis for the discussion of regional Neogene palaeogeography. Absence of accurate dating has led to geometric correlation problems, even though the sedimentology has been strongly improved by Baize (1998). Locally, old Crag (the Blehou Crag) are of Serravallian age (foraminifera, malacofauna: Hommeril 1967), rich in authigenic glauconite.

A first dating by ESR was performed at Esquay (Bayeux) (Figs 1; 3; 4; 6) in the St-Vigor sands believed to be of Early Quaternary age (Clet-Pellerin 1983; Pareyn 1987). The sequence records first the lower St-Vigor/Esquay Formation, a brownish clayey sand similar to the "Bolan" or the "Frelonnière Formations" of Brittany, resting immediately on the calcareous substratum weathered by a *terra fusca*. On a gravel lag, the upper St-Vigor/Esquay Formation consists in a whitish, fluviatile to estuarine sand, reworking Jurassic shells. It records at least two transgressive events with a flow direction towards the West and is similar to the Radenac Formation and the St-Eustache of the lower Seine. The ESR dating of the base of the upper St-Vigor Formation yielded an age of 3.2 ± 0.36 Ma, quite consistent with the age of the Radenac in Brittany. Both formations present synsedimentary water escape figures, an alluvial sheet (Early Quaternary) incised the St-Vigor Formation.

In the Cotentin, the coring at Marchésieux, within another graben (Santeny basin; Garcin *et al.* 1997), crossed two transgressive infra-tidal shelly sandy muds, of which the uppermost is of slightly cooler climatic condition. The lower one, the Marchésieux sandstone, is believed to represent the 3.1 Ma highstand (Baize 1998). A complex of marls (Boscq d'Aubigny and Pierrepont) followed, including a diachronic crag, the Bohon Crag, correlated with the St-Jean-La-Poterie *Nassa* sp. clay (Pareyn 1987). The Bohon ferruginous conglomerate (20 m thick!) was believed to be equivalent to the crag, as it reworks shells but *Nassa* sp. is absent (Vieillard & Dollfus 1875); it is probably a late Miocene unit.

It is covered unconformably by sand abusively called St-Vigor Formation (Baize 1998), including some shore ice rafted blocks. To avoid further confusion, we shall call this formation the "Lessay sands". This formation is incised by the

"Millières sands" of estuarine to fluvial origin, strongly deformed by local transtension tectonic. The record ends with a peat attributed to the Waalian (1.3 Ma) (Baize 1998). Subsidence is active during the setting of the Marchésieux and Millières formations, local flexural activity starts at the end of Millières sands deposition (prior to 1.3 Ma: Baize 1998).

The Seine and the Somme valleys represent two flexural gutters evolving into synforms during the Neogene and Quaternary. This is shown in the Seine by the preservation of clay with flints (Laignel *et al.* 1999) or by abrasion surface within the Somme valley (Van Vliet-Lanoë *et al.* 2000b). At Fécamp (Figs 1; 4), Crag are locally preserved in a karst at *c.* 55 m NGF and at *c.* 115 m at Valmont. They were attributed to late Miocene by Brébion (*in* Bassompierre *et al.* 1972). ^{87}Sr/^{86}Sr dating on shells from the collections (Brébion, Lauriat-Rage) of the Muséum national d'Histoire naturelle of Paris (Mercier *et al.* 1997) yield a lower Tortonian to late Tortonian/Messinian age estimation, but outcrop contamination is susceptible to age diagenetically the shells.

In the Pays de Caux and the Seine estuary, the records are very similar to these in Brittany: the first sedimentary prism is represented by the brownish Lozère sands (Normandy facies, younger than in the Paris Basin), a fluviatile to estuarine facies derived from a Massif Central source (Cavelier & Kuntz 1974; Kuntz & Lautridou 1974). These are probably contemporaneous with the Bolan Formation of Brittany. At Valmont, they are thought to rest on "Redonian Sands" (Cavelier & Kuntz 1974), but this stratigraphic attribution is not demonstrated. A second sand prism is constituted by the infra-tidal whitish sands of the St-Eustache Formations, at 95 m NGF (or Valmont: Cavelier & Kuntz 1974), which are overlain by the clays of "La Londe formation" attributed to the Gelasian owing to its pollen assemblages (Clet-Pellerin 1983). New observations at Petit Quevilly (downstream of Rouen) showed the superposition of the coarse Lozère sands and of two alluvial sheets of the St-Eustache Formation separated from the lower one by a strong pedogenesis, by transtension and by the reworking of large blocks of flint. Though the mapping of these formations is generally presented as a channel flowing across the Pays de Caux, the outcrops are clearly located in two zones: 1) an E-W one, related to the Bay of Seine; and 2) an independent one, related to the subsiding coastal basin of Fécamp. The Seine was forming a gulf shown by the occurrence of an estuarine facies of the Normandy Lozère Formation as far upstream as Oisel, in a middle terrace position below the Quaternary alluvial deposits. It is thus possible that the Seine valley, which is strongly meandering, was already excavated to the middle terrace at that time. Evidence of *in situ* silicification above Petit Quévilly should argue for an earlier setting of the valley at least form a late Oligocene age. This could explain the strong development of the cryptokarst at Cap d'Ailly, the preservation of the Fécamp-Valmont Crag in large karstic depressions and the thick preservation of "clay with flint" in this "gutter" (Laignel *et al.* 1999).

From the Somme valley up to the Canche river, discrete outcrops are preserved, at an altitude of 40 m at St-Valery, 43 m at Wailly-Beaucamp reaching 50 m at Montreuil. The record consists generally of whitish sands forming sandy bars or interstratified with ochreous marine pebbles, overlain by reddish stratified clays, similar to those of La Londe and re-incised by Early Quaternary gravel sheets, usually strongly frost shattered. The first dated fluvio-marine unit of Mt-Pillar yielded by ESR a very late Pliocene age (1.867 ± 0.27 Ma). The second level belongs to a commonly recorded highstand (1.4 Ma: 1.36 ± 0.117 Ma and 1.58 ± 0.28 Ma), correlated with the Waalian one. The palaeocliff is partly fossilised by these and is probably earlier. The large palaeotidal ridges belong to middle Pleistocene (Van Vliet-Lanoë *et al.* 2000a).

The Somme valley is overcalibrated in its upstream portion east of Amiens. A meandering train is locked at the entrance of the Boves-Argoeuvre subsiding zone, which is superimposed on a Permian pull-apart (Mégnien 1980). This depression is truncated by a surface and a palaeocliff at 60 m related with "avelanaire" gravels derived from the Eocene. They are likely the traces of a gulf of Neogene age intruding the valley as high as Boves. On shore, in front of the estuary, a very large splay of gravel is explained but with difficulty as Last Glacial (Auffret *et al.* 1980). It is more probably the trace of a more important Somme valley draining the upper Oise and the upper Meuse during the Neogene. A capture of the upper Oise by the Seine may be explained by a retrogressive capture on the crossing with the Bray fault (Van Vliet-Lanoë *et al.* 1998b). Ardenne derived gravels have been found in old drift on the NW side of Bray close to Persan.

Cornwall (Figs 5; 6)

Outcrops are sparse in this region. The first important one is now a very limited section of St-Agnes (Figs 1; 5; 6). A recent description by Walsh *et al.* (1987) defined as resting on the Beacon sands, a Chattian clayey sands (Oligocene), a lower member of the St-Agnes Formation, the Doble sands (102 m OD-130 m OD) seemingly tilted seawards. It consists of two sand bodies, including some pebbles, 1.50 m above the basal unconformity; it is believed to be aeolian by Walsh *et al.* (1987) though Coque-Delhuile (1987) considered it as marine. The other member, the New Down clays, contains lignite with a Miocene Mediterranean flora. A former description (Van Vliet 1981 *in* Coque-Delhuille 1987) and a recent sampling session for ESR (1996) and old pictures (1981) allow a re-interpretation of the sections. A first point is the unconformably overlaying of the New Down Clays by some of the Doble sands. A unit of the Doble sands reworks sand and goethitic ironstone with loading figures; two tidal sandy units overlay it with beach bars separated by a podzolic soil. The next unit was visible in 1996 and yielded an ESR age of 2.35 ± 0.37 Ma; it contains large drift blocks of goethitic iron pan and is lately covered by slope deposits.

In the light of the results obtained from other regions, it seems thus that the lower part of the Doble sands (with the pebble line) corresponds to a late Miocene shore-face sedimentation. The New Down clays correspond to a continental event probably associated with the goethitic iron pan. A tectonic activity responsible for the co-seismic loading thus takes place before the Piacenzian but after the late Miocene. The upper Doble sands seems to correspond to the classical Red Sand and is probably Piacenzian in age. According to the dating, the "fluviatile" unit belongs to the Gelasian while slope wash is Pleistocene.

The other important outcrop, the St-Erth formations (16-28 m OD), is known to be late Pliocene in age (Jenkins *et al.* 1989), according to its rich molluscan-foraminifer assemblages when compared with the malacofauna of the clays of the Elorn valley in Brittany (Figs 1; 5; 6). These infratidal clays are more precisely positioned from the dynocyst assemblages by Head (1993, 1998) between 2.09-2.00 Ma by comparison with the biozonations of the Haq *et al.* (1988) chart.

Rivers, probably of Chattian or/and Tortonian ages, see their watershed extended towards the North though the southern portions of the valley are flooded in the form of rias. This seems related with an important tilting to the South of Cornwall. Main cliff, West and North to Cornwall is associated with an important strandflat, though in the South, a palaeocliff is drowned to - 50 m OD (Fig. 3).

Boulonnais, Flanders and Ardennes

Until 1998, Neogene deposits were unknown in the Boulonnais area, apart from the Diestian shore face ironstone occuring at Sangatte on Les Noires Mottes (150 m NGF). De Heinzelin described in 1964 three marine abrasion surfaces as Quaternary. Now, the lowermost one, 90-100 m NGF, is attributed to late Serravallian or early Tortonian by Van Vliet-Lanoë *et al.* (1998b), on the base of the geomorphology and geometry of the deposits, ante-dating the "Diestian" goethitic pan, but not necessarily the marine formation. It is responsible for the shaping of the internal cliff and the setting of the Wimille Formation, already overlaying a well-developed oxisol on Jurassic marls (Figs 5; 6).

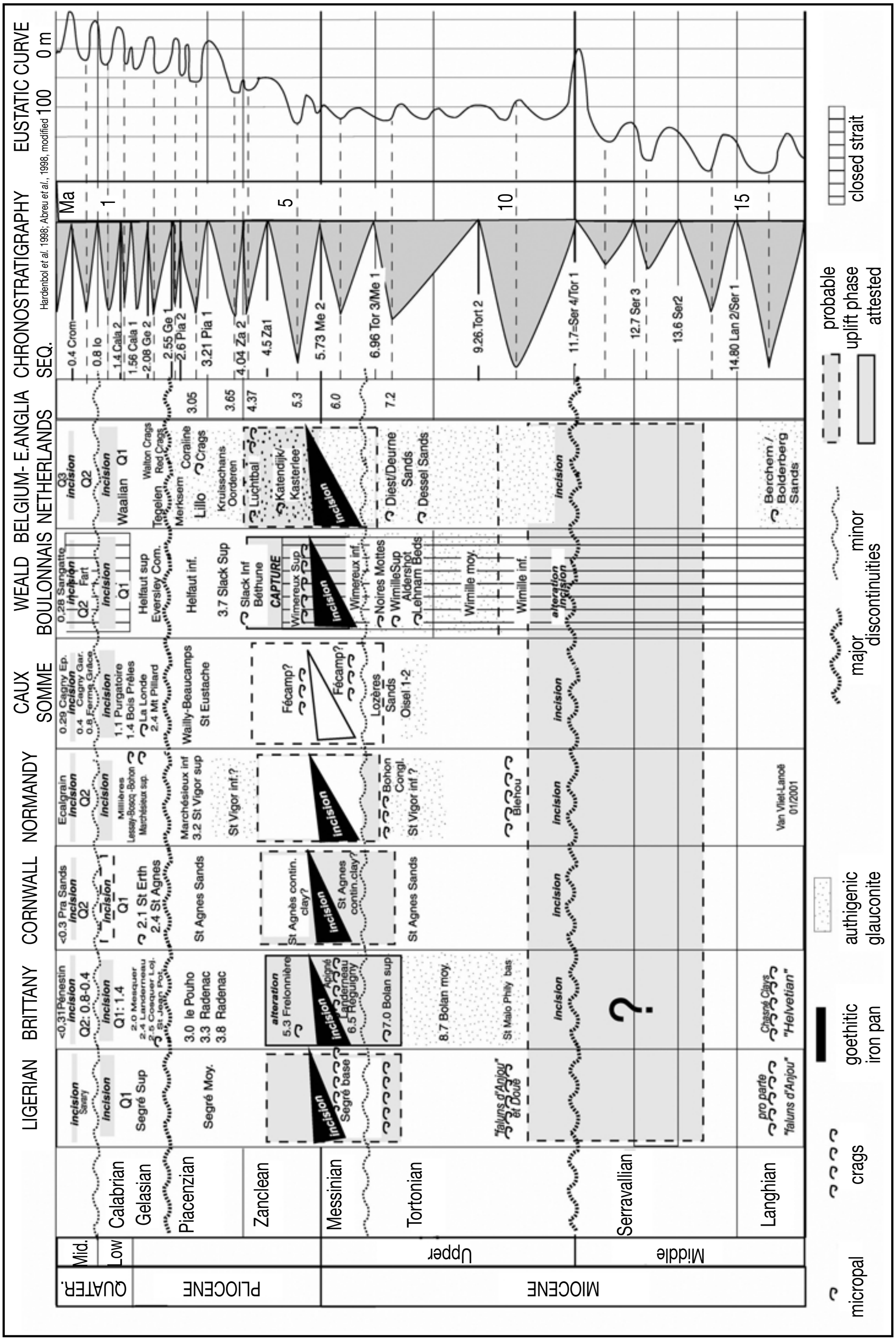


FIG. 6. — Correlation table from the stratigraphical units from the Ligerian region to the North Sea basin. Location and formation names are in the text. Ages are given in Ma.

The upper part of the Wimille Formations is a ria infill of sands rich in authigenic glauconite, with ghosts of shells and some lignitic clay at the top. This upper unit is faulted, affected by seismogenic deformations and sealed by an iron pan (goethite) associated with the development of an oxisol. The Wimille sands differ by their petrography from those of les Noires Mottes which belong to the Belgian basin, but are quite similar to the deposits, goethite cemented of the Pre-Meuse system, South of the Ardennes (volcanic quartz). After this tectonic movement, valleys are incised and a new marine surface is locally shaped and weathered. It leads at la Pointe-aux-Oies (Wimereux) to a tidal platform on a 8 m NGF with sandstones beach cobbles, both weathered with a 2 cm thick goethitic crusting and karstification, the lower Wimereux Formation. This surface crops out from Wimereux at 5 m NGF, up to Wissant (Butte Carlin) at about 60 m. It is overlapped by two transgressive units, the lower Slack Formation, a shore face with beach gravel to lagoonal deposit, with some synsedimentary slump activity and tilting, including by its summit the freshwater palynomorph *Pediastrum* sp. (S. Louwye pers. comm.). The upper Slack Formation resembles the Red sands of Brittany: it begins with a fluvial conglomerate with progressive intrusion of prograding sand bars with 3D dunes, organised in a braided river system and ends with a beach bar and pass system. An ESR dating has been obtained from the base of the sands and yielded an age of 3.672 ± 0.521 Ma (Van Vliet-Lanoë *et al.* 2000b). In the basal gravel, rounded blocks of a crag begin to crop out: these elements are believed to trace an highstand in the upper Wimereux Formation. Within the morphostructural frame of the Boulonnais, in the light of the first ESR dating, of the recorded succession of shore face deposits, and the eustatic chart (Hardenbol *et al.* 1998), our present-day interpretation of this sequence is as follows (Figs 5; 6): late Tortonian for the upper Wimille Formation, Messinian for the lower Wimereux Formation, Zanclean for the upper Wimereux Formation and the lower Slack Formation, Piacenzian for the upper Slack Formation. Gelasian seems to be lacking. The strait remains closed during the Tortonian and the Messinian, on mineralogical and micropalaeontologic grounds. In the St-Omer basin and close to Bethune complementary sections are visible on the northern flexure of the Boulonnais-Artois. The Helfaut terrace constitutes a large diachronic fan cropping out from about 150 m to the SW of Boulonnais to reach about 60 m at Clairmarais, North of St-Omer. It consists of a whitish prograding sand body, including some silty lenses, followed by a conglomerate of rounded cobbles, of often ochreous flints and rare iron stones, forming a beach ridge. It is lately covered by a thinely stratified silty sand interpreted as a mudflat. Frost shattered alluvial gravels truncate these units before being invaded by alluvial clayey silts. This record may be interpreted in the same way as the sequence of the Picardy coast: a late Piacenzian gravel ridge (upper Slack Formation), a Gelasian mudflat and Early Quaternary alluvial sheets. Local tectonism is recorded as tensional and may be confused with periglacial, though it is clearly related to the activity of the Landrethun flexure zone. The other outcrop is located close to Béthune (Bois des Dames, 60 m NGF); it consists in a massive greenish silty sand, with glauconite and some bioturbation resting on an erosion surface truncating the Eocene sands, covered by a fine layer of clay and re-incised by Plio-Quaternary gravels. This unit belongs to the same diachronic surface and probably records either a lower Piacenzian unit or an older Pliocene unit; Tortonian is excluded as it is located South of the Diestian shoreline (150 m, Mt Cassel).

Belgium (Fig. 6)

Except the Diest sands Formation, Neogene sediments in Belgium are only found in the Antwerp and the Campine areas. As in Brittany, the region seems to be emerged since the late Rupelian. The lack of early Aquitanian and the very shallow transgressions of the late Aquitanian (Edegem Sands) and of Burdigalian to Serravallian in NW Belgium is an argument in favour of a relative uplift of the southern part of the country at that time. Other Neogene outcrops are the topmost sequence of the "Monts des Flandres" where Heinzelin (1955) had already observed the trace of a marine highstand followed by fluviatile deposits.

In northern Belgium, the lithologies are dominated by medium to coarse grained sands, often rich in authigenic glauconite and intercalated with shelly beds. Decalcification and iron pans are frequent. De Meuter & Laga (1976) and the work of Doppert *et al.* (1979) allow a biostratigraphical correlation with the deeper marine Neogene of the Netherlands.

In northern Belgium, above the Berchem and Bolderberg formations of late Aquitanian to Serravallian age, a sand body well defined on seismic profiles is constituted by the Dessel and Diest-Deurne sands, the base of which records a clear erosive uncomformity (Gullentops 1954; Demyttenare 1989; De Batist & Versteeg 1999). Demyttenare (1989) demonstrated the importance of the tectonic control on the subsidence in the Roer valley graben and the location of a large W-E erosive channel, developed in parallel with the Flemish valley.

The base of the Diest Formation is a deeply incising erosive gully. Its base is filled with fine Dessel sands but the main body of Diest sands is formed by prograding sets of obliquely stratified tidal sand bars exposed in the Hageland and identified on seismic sections in the Campine subsurface. The Diest sand is the last marine deposit in southwestern and central Belgium; in northern Belgium, it is overlain by the marine abrasion basal surface of the early Zanclean Kattendijk and Kasterlee sands. This relationship suggests an important tectonic uplift event during the late Miocene. The erosion at the base of the Diest sand formations has locally eroded almost 100 m of Boom Clay; such massive erosion suggests a correlation with the Ser4/Tor1 sequence boundary low sea-level event, corresponding to the Msi-4 oxygen cooling event (Hardenbol *et al. in* Graciansky *et al.* 1998: chart 2).

Recent biostratigraphic analysis by Louwye *et al.* (1999) indicates that the Diest Formation (Dessel and Diest sands) was deposited sometime during the time interval (*c.* 6 Ma) from the early Tortonian to the Messinian. Reworked palynomorph assemblages in the base of the Dessel Sands are indicative of a lowstand systems wedge. However, small numbers of oceanic dinoflagellate cyst species are found throughout the Diest Formation and point to the influence of oceanic waters; most probably these species have been swept into this nearshore environment by oceanic currents. The Dessel and Diest sands are diachronous and display a marked younging towards the NW with a shifting in time of the depocentre to the shelf during late Tortonian-Messinian times. Louwye *et al.* (1999) correlate furthermore the boundary between Diest and Kattendijk/Kasterlee sands with the Me 2 sequence boundary of Hardenbol *et al.* (*in* Graciansky *et al.* 1998: chart 2). No major sedimentation break was found in and between the Dessel and Diest sands, as in homologue formations in Brittany (Fig. 6). The Diestian sands are channelized, rich in authigenic glauconite, strongly bioturbated, and testified to a river supplied shore face sedimentation. They are poorly rounded and derive probably from deep saprolites, like in Brittany. They represent apparently successive emersive events related to intraformational iron pan, of which the upper one is strongly consolidated by goethite. Upper gravels present a deep ochreous cortex (desilicification cortex stained by limonite).

The Kattendijk and Kasterlee formations have a basal gravel lag and belong to the lower part of the Zanclean. The thin Crag of Luchtbal (Lillo Formation) is associated with the second Zanclean sequence (Za1 of Hardenbol *et al.* 1998). This Lillo Formation was deposited in a shallow marine environment. Eastwards it becomes more estuarine and fluviatile as in Brittany and the lower Rhine area (Hilden *et al.* 1988).

Though the Dessel sands are irregular and often associated with an incised valley fill facies, the Diest sands form a spectacular prograding Tortonian wedge (Vandenberghe & Hardenbol 1998; Vandenberghe *et al.* 1998; Louwye *et al.* 1999), slightly truncated by the Kasterlee and Kattendijk sands attributed to the lower Pliocene.

From these data, the main tectonic activity took place in this region during the late Miocene/early Pliocene. Correlative to the tectonic movements controlling the transgression and sedimentation pattern, river network and karst confirm the succession of events.

The Neogene cryptokarst described by Dupuis (1992) in Condroz was at least initiated during the lower Oligocene and its reactivation was probably penecontemporaneous of the late middle Miocene and of the late upper Miocene tectonic uplift events. The sedimentary gap from upper Oligocene to middle Miocene records perhaps a more important event. Goethitic iron pan including gravels crops out commonly at the surface of Brusselian sandy fan South of Brussels, in apparent continuity with the palaeosurface of the Monts des Flandres, as already stressed by Gullentops (1957). Sedimentation of the Rhine river recorded since the middle Tortonian an increase of energy by dominant fluvial import and arrival of Alpine gravels in the basin (Hilden *et al.* 1988).

The evolution of the Meuse river is also complementary of the shore face data. Already in 1962, Heinzelin proposed a capture in middle or late Miocene of the "Ardennes" Meuse river (Charleville-Maizière to Givet) by the Sambre river. According to Voisin (1981), the Pre-Meuse was flowing westwards during the late Miocene. For Pissart (1974) and Bustamente (1976) the Meuse river drained the Vosges since the Pliocene across the Ardennes. Observations realised during geological survey of the Givet, Fumay and Renwez sheets (1/50.000) have shown the existence of North-South drainage recorded by fluvial deposits from Vireux and often consolidated by goethite. These deposits record a riverslope increase to the top (uplift). Petrography of the deposits records a mineralogical contribution solely from the Ardennes. These deposits crop out in a middle terrace position below the Quaternary alluvial sheets at Vireux and at Fumay. At Les Vieilles Forges, these goethitic conglomerate blocks are incorporated in light yellowish soft fluviatile sand, similar to the Pliocene one, prior to the deposition of the Quaternary alluvial sheets (Voisin 1981). Quaternary alluvial sheets are characterised by mineral assemblages derived both the Ardenne and Jurassic sources (in Lorraine).

Weald and Thames basin (Fig. 6)

In the Weald and the Thames basin, a controversial interpretation of outcrops exists.

The morphology of the Weald is similar to that of the Boulonnais, but for a stronger relief. Woolridge & Linton (1939) considered that the excavation of the Weald was post-Pliocene because of the age attributed to the Lehnam beds. In fact the system is not too different from the Boulonnais-Artois record.

The Lehnam beds, first described by Preswitch (1858), occur at 180 m OD, NW of Maidstone and also in the vicinity of Beachy Head. These deposits are preserved just above the northern flexure of the Weald. A complete description is available in Worssam (1963). They are often disturbed and preserved only in solution pipes. They usually consist of orange sand, micaceous, sandstone, dark brown clay and ochreous flints, sometimes included in dark brown ferruginous sandstone; iron stone is quite common buried by red residual clays. Their heavy mineral content is different from that of the Coralline Crag of East Anglia and seems to be equivalent to the middle Miocene of Netherlands (Elliott *in* Worssam 1963). The palaeontological content of the Lehnam beds is also different and has been attributed to the upper Miocene (Cooper 1980; Balson 1989) after several interpretation shifts between Pliocene and middle Miocene.

To the East, SW of Aldershot, the relict gravel of the Caesar Camp reaches 186 m OD in altitude. It is a complex of about 20 m of sands and gravel, sometimes more clayey, infilling an incision in the Paleocene Bracklesham beds (Clarke *et al.* 1979). These deposits are preserved just North of the Hog's Back flexure of the Weald. Flint gravels are well-rounded, orange yellow to brown in colour with nail impact traces or even not too much transformed compared to the flint nodules. Some quartz and sandstone cobbles exist in the sandy matrix. Prograding thin sand bodies are visible along one of the roads; direction is clearly eastward (pers. obs. Van Vliet-Lanoë 1999). Podsolisation is extreme and frost shattering is very limited. These gravels and sands correspond very clearly to a shoreline, in similar position as the Lenham beds. They may probably be interpreted now as an equivalent of the Diestian shoreline in Belgium. The Netley Heath gravel at about the same altitude is probably the continuation of the same system. These gravels are incised by the Wey and the Blackwater from which high terraces largely extended into the Weald from *c.* 120 m OD and pierced the sides of the Hog's Back flexure (Aldershot geological sheet, One Inch Scale, Ordonance Survey, 1928). To the East the Stour and Medway rivers behave similarly.

Lower and middle Pliocene deposits seem absent in the Thames basin. Recently, Head (1998) reattributed the Coralline Crag of East Anglia to the Piacenzian (< 3.3 Ma) owing to its dinoflagellate content with at least two transgressions. Laga (1972) and Doppert (1985) have already correlated them to the Luchtbal Formation in Belgium. Evidences of late Pliocene are still found along the North Downs.

A little further NE of Aldershot, the Eversley Common represents a "river" terrace at 90 m OD, cut into the Paleocene Barton Sands (Clarke *et al.* 1980). The basal gravel of the terrace is very well rounded, without evidence of frost shattering, sometimes intermingled with whitish sands and covered as in many sites to the West of the Channel by a reddish stratified clay before being covered by periglacial alluvial sheets. These can be interpreted as late Pliocene deposits, similar to those that outcrop in Brittany and Seine estuary and dated of the Gelasian by palynology. It seems thus that North of the Weald flexure (Hogs' Back) large diachronic pediments developed in a similar way as the St-Omer pediment. These relicts pediments develop along the Black Water course as perched alluvial ridges East of Camberley. The Eversley Common gravel and clay are probably equivalent to the Red and Norwich Crag of East Anglia, tilted eastward by Early Quaternary neotectonism (Sumbler 1996). On account on the dinoflagellate content (Heads 1998), the basal Red Crag (Waltonian) in East Anglia deposited *c.* 2.7 Ma. The Pre-Ludhamian from 2.6 to 2.4 Ma is already cool and corresponds roughly to the Praetiglian and Tiglian A-B. The upper Red Crag, or Ludhamian, corresponds to the marine highstand (2.3 Ma) and the Tiglian C, and is often clayey or silty clayey.

Concerning Plio-Pleistocene river evolution, the Wey, Blackwater, Medway and Stour rivers are believed to flow northwards in a parallel course across East Anglia and the western shelf of the southwestern North Sea, before shifting eastwards (Bridgeland 1988). It means thus with observation at Aldershot and at the Eversley Common that the river network was incised across the Hog's Back before the Pliocene. Jones (1999) dates this incision back to the Eocene. This answer is very similar to that of the northern side of the Boulonnais and is probably synchronous. The structural surface developed at 100-120 m through the Weald, though somewhat deformed by internal flexures has probably the same origin as in the Boulonnais and is apparently marked by a marine abrasion surface reworking partly the top of the "Upper Green Sands" especially from the Ashford to Lehnam districts.

DISCUSSION

EVOLUTION OF SHORELINES AND TECTONICS (FIGS 3; 4; 6)

Owing to all these data we are now able to reconstruct a series of shoreline maps through Brittany, the Channel and the southern North Sea. Data provided by Zagwijn (1989), Funnel (1996) and Louwye *et al.* (1999) have been used for the North Sea. For the Channel, the sources have been described for each region. The Mid-Miocene uncorformity recorded in North Sea basins is often attributed to prevailing climate and eustatism with a major ^{18}O decrease in stable regions (Husse & Clausen 2001). During late Paleogene, the "Channel gulf" was excavated to the East (Van Vliet-Lanoë *et al.* in press), probably re-reaching the Boulonnais during the Tortonian (lower Wimille Formation). This re-incision facilitated by enhanced thermal gradient on the Atlantic as mid-latitude still experienced mild temperature (Lécuyer *et al.* 1996), while northern Atlantic and especially Iceland experienced glaciations (Bleil 1989; Mangerud *et al.* 1996). This resulted in stronger storminess and higher erosion efficiency (Van Vliet-Lanoë *et al.* 1998b) as theoretically stressed by Posamentier *et al.* (1988). The platforms were excavated by marine abrasion as well as also the main palaeocliffs, mostly during lowstands, as recorded by the occurrence of an unconformity at the base of the Cockburn Formation in the Western Approaches (Evans 1990), which constitute a prograding submarine fan dated roughly of late Serravallian-Tortonian. At the same time, tectonism increased, progressively reactivating the Variscan front and the Armorican massif, in parallel with progressive rise of the sea level since the base of the Tortonian. Most of the sandy Neogene record started about 11 Ma ago in association with rising accommodation space. Tilting in the vicinity of the Variscan front is also proved, by the tilting of Cornwall, the drowning of the palaeocliff along the southern coast of England reaching - 80 m OD South of Portland, - 50 m OD South of Wight, - 30 m OD South of Dugness Point, by the absence of sediment into the northern valley of the Channel, whereas the southern one is filled. During late Tortonian/Messinian (upper Wimille Formation), the Strait corresponds to a shallow saddle with smooth relief guiding the river network. The strait remains closed and huge gravel or sand bars follow the coast from the Caesar Camp in Britain to the Diest region in Belgium, an area of large sand banks.

Along the Channel, the Pliocene transgression is correctly recorded especially at the West and along subsiding margins (South of SASZ or along the Western Approaches). A long wavelength deformation uplifts the whole sector (zones 1-2, Fig. 1A) during the Messinian. Subsidence starts by incipient flooding during the early Zanclean in the Rennes basin (5.2 Ma), apparently lately towards the Northeast related to delayed relaxation. The southern North Sea reacted similarly. Flooding is more important during the Piancenzian testifying to a main relaxation (Fig. 4), despite a lower highstand level than during the Zanclean, and also a delayed answer to the N-E. The uplift of the northern part of the Tunnel Flexure at Sangatte facilitated the retrogressive incision of a Lobourg river flowing towards the subsiding North Sea as stressed by Michelsen *et al.* (1995). Flooding of the strait occurred probably only during the late Zanclean, *c.* 4.4-4.3 Ma, as two highstands are recorded in the sequence of La Pointe-aux-Oies (lower Slack and upper Wimereux formations) and as the Murray Pit infilling is correlated with the Luchtbal sand. The lower Slack Formation recorded fault activity through tilting and slump. It is thus possible that the tectonic reactivation related to the 150°N shortening event, transmitted from the South of Europe since late Tortonian, reached at that time the Variscan front. A residual uplift delayed the flooding of the Strait. During late Pliocene, the global lowering of the sea level related to the building of the ice caps revealed North of the Strait the occurrence of a bulge, created by the reactional uplift of the Anglo-Brabant massif. This zone also controlled the pattern of the upper Miocene transgression as well as that of the Middle Quaternary. The modality of the marine transgression signs the uplift of the Variscan front and of the associated Anglo-Brabant block induced by the 150°N shortening on the Variscan front and older deep faults.

EVOLUTION OF THE RIVER NETWORK AND TECTONICS (FIGS 4; 7)

In Brittany the river network is mostly adapted to the structural pattern. Nevertheless three main systems are visible: an Oligocene system characterised by shallow valleys, a Miocene system already well incised, but sometimes disconnected from the present-day one and a Plio-Pleistocene system, incised stepwise. In subsiding zones, such as the regions of Rennes and Redon, the Vilaine valley preserves Pliocene deposits below the

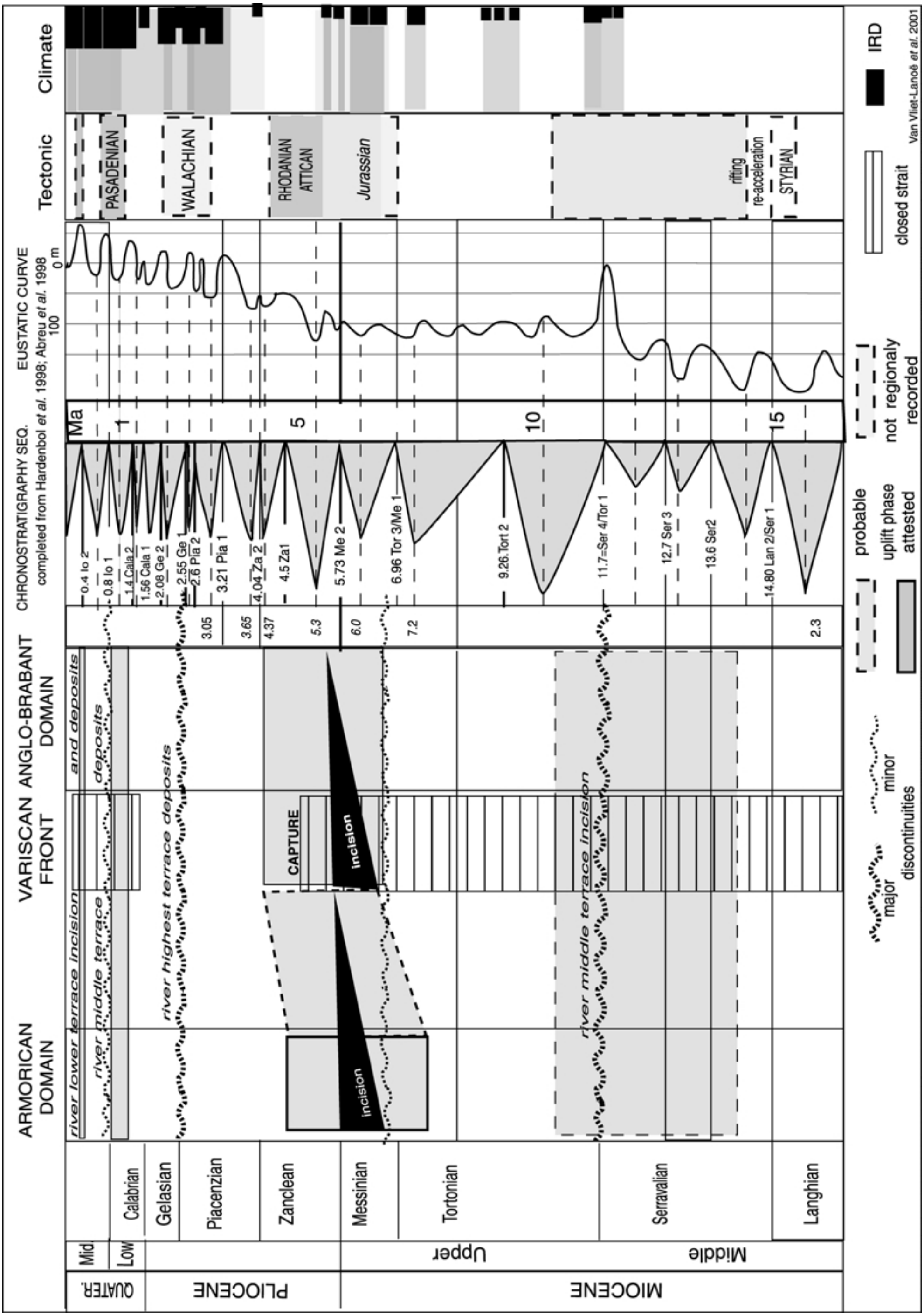


FIG. 7. — Tectonic correlations, eustatism and climate. Tectonic phases from Haq & Van Eysinga 1998. Darker shades in the climate column indicate cooler conditions (Ice Rafted Debris from Bleil 1989). Ages are given in Ma.

Pleistocene middle Terrace. The same situation is found for the Blavet, South to Lorient. The Seine valley presents a similar organisation at Oiselle, 10 km upstream of Rouen (probably late Miocene). In the Meuse valley at Vireux and at Fumay, local fluviatile sands, presumed late Miocene in age, were found by augering below the Pleistocene middle Terrace. Other outcrops also indicate the existence of a North-South drainage of the Meuse between the Famenne and the Lorraine.

Basal incision observed at the base of the Tortonian at St-Malo-de-Phily (Vilaine), at the base of the Dessel sands in Belgium, at the base of the Caesar Camp Gravel at Aldershot (Blackwater river, GB), argue for a generalized incision during the lowstand at the base of the Tortonian (11.7 Ma: Ser 4/Tor1). The evidences of higher and more recent (Piacenzian, Gelasian or Early Quaternary) alluvial deposits in the same sites, all included in the domains I-II of Fig. 1, demonstrate that long wave-length deformations inducing block tilting and/or continuation of basin inversion took place during the Attic tectonic phase controlled by an important relaxation mostly since the late Zanclean North of the Variscan front.

Tilting and flexural movements have induced a capture of the river Meuse by the North Sea basin due to the opening of the Givet-Hastière "canyon", precisely located on a N-S fault, a capture of the upper Oise-Sambre by the Meuse at Namur (Lacquement *et al.* in prep.), a capture of the Eastern Channel by the Lobourg and a capture of the Oise by the Seine in response of the uplift of the Bray and the Variscan front (Van Vliet-Lanoë *et al.* 1998b), and to the South, a capture of the "Lozère Loire" by the lower Loire. Karstification was potentially activated on faulted zones both during the early Tortonian lowstand and during the Attic phase. This could partly explain the Hurd Deep, already attributed by Boillot (1964) to a drowned karstic depression. Close to Argentré, the red sands fossilised the Mont Roux cave (Fleury *et al.* 1989). Similarly goethitic iron pans are common after the start of the global uplift at the end of the Messinian: they correspond mostly to water table alios developed in down slope position under subtropical climate. They are omnipresent in the whole region and sign an incision of the river network or a lowering of the water table (dominant uplift or regression). During the Piacenzian relaxation, the former valley incisions have been filled up to the level of the highest Pleistocene terraces before being reincised in relation with the global lowering of the sea level, and a temporary regional uplift related with the early Pleistocene tectonic phase. Various Pliocene deposits were inverted as at La Frelonnière in Brittany or within the Hurd Deep. Recent high resolution seismic indicate a bulging of the palaeovalleys infilling, interpreted as "periglacial pingo scars" by Lericolais (1997); they are in fact transgressive deformations (Riedel folding) of a fluviatile complex of Plio-Pleistocene age.

The same type of geodynamical evolution exists inland as in Bresse (Petit *et al.* 1996; Sissingh 1998), in the Somme, the Vilaine or the Meuse valleys. From the stratigraphic analysis (Kalin 1997; Sissingh 1998) the folding phase of the Jura is younger than 11 Ma. An uplift of the Vosges and the Black Forest is recorded from the middle Miocene, as in the Channel regions. Applying the same logic, we may read with a neotectonic key the evolution of the upper and lower Rhine (Sissingh 1998). During the middle Tortonian relaxation, the Dutch Rhine began to record, by the coarse sedimentation (Hilden *et al.* 1988), the early uplift of the Alps (Van Dijk & Scheepers 1995) and the Jura fold erosion. During the compressional Attic event, the Rhine graben was subsiding, capturing temporarily the Aar from the Zanclean. During the Pliocene relaxation, the Aar flooded temporarily the Bresse (Reuverian), in front of the Jura orogeny before its capture by a renewed subsidence of the Rhine graben during the Early Quaternary.

SEQUENTIAL INTERPRETATION

The sequence stratigraphy analysis of the Upper Cenozoic deposits of northwestern

Europe make now rather clear that the Tortonian 1-2 sequence is preserved in scattered places from a part of the Dessel/Diest sands to the lower Bolan Formation. Nevertheless the best preserved sequence is Tortonian 2-3 which represents the wider available accommodation space for sedimentation during the Tortonian relaxation event. It consists mostly of shore or estuarine formations (mostly along the Channel) of Dessel/Diest, Normandy Lozère, lower St-Vigor, lower St-Agnes, Bolan, and of the spectacular shoreline running from the Aldershot, through the Lehnam beds and the Dover Strait to the Monts des Flandres (Fig. 4).

The lower Messinian sequence (Tor3/Me1 to Me2) is mainly represented by continental formations, by soil weathering (tropical podzols, oxisols) and by water table goethitic pan, the upper sequence (M2-Za1) consists in Crag or fluvial deposits. This means that accommodation space is the most restricted and that greater part of the domain is emerged during the highstands, excepted the reaches of the North Sea or the South Armorican platform. High biogenic productivity responsible for the Crag is probably a response to enhanced soil erosion and consecutive fertility. Za1-Pia1 are recorded in subsiding areas only, South of the SASZ, in the Low Cotentin or in the southern North Sea domain, in parallel with an important weathering on the still emerged domains. Piacenzian Za2-Pia1-Pia2-Ge1 sequences are the best preserved ones, corresponding again to the larger accommodation space for sedimentation during the post Attic relaxation. The Attic event is slightly diachronous: based on outcrops record, it starts before 7.2 Ma in Brittany, probably 6.5 Ma in Belgium and relaxation started from about 5 Ma in Brittany to about 4.2 Ma in the Boulonnais. A similar sketch can be applied to the older transgression of Burdigalian-Aquitanian, in relation with the post compressional relaxation of Save phase, as well as in the Ligerian and in the southern North Sea domains, relaxation probably limited by the higher intensity of the long wavelength deformation compared with the Attic one. The limited record of the Langhian-Serravallian in the area is probably of same origin in combination with the Ser4/Tor1 regression responsible for drastic erosion (Fig. 7). Opposite to this, the Pleistocene tectonic events at 1-0.8 and 0.4 Ma seem to be rather synchronous and in agreement with events recorded to the South (Van Dijk & Scheepers 1995). This is not yet explained but may be interpreted as a reaction to a tectonic event related with the alpine orogeny by opposition to events controlled by the Atlantic rift.

CONCLUSION

In the Channel and southern North Sea areas, sandy facies, sometimes in the form of Crag cover mostly the Tortonian, the Messinian and the Piacenzian. In subsiding zones, Serravallian and Zanclean formations are also recorded. After the middle Miocene, an eastward re-expansion of the Channel occurred in relation to rising sea level following the Tortonian lowstand which correlates with the setting of the Cockburn Formation in the Western Approaches. A slightly diachronous global uplift of this Variscan region associated with a southward tilting of the northern front promoted a strong modification of the river system during the Messinian and the first part of the Zanclean. This led to a capture of the western Channel by the southern North Sea during the late Zanclean. A progressive cooling of marine microfauna (foraminifera, dinoflagellates) took place in relation to the flooding and excavation of the Dover Strait. This is not very perceptible in the composition of gastropods assemblages, characterised by a low frequency of the Nordic taxa, although bivalves show a similar population pattern to that of the foraminifera. The period of maximum accommodation space available for sedimentation is the upper Tortonian and the upper Piacenzian, both fitting in with tectonic relaxation events also signed by a synsedimentary microseismicity.

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