BASEMENT CONTROL ON SHAPING AND INFILLING OF VALLEYS INCISED AT THE SOUTHERN COAST OF BRITTANY, FRANCE

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Abstract: The shape and infilling of the submerged parts of valleys incised along the southern coast of Brittany (France) have been investigated using very high-resolution seismics and a small number of piston cores. The valley location and morphology are found to be controlled mostly by submarine topography, which is marked by a well-developed fault zone that lies between the modern coast and a prominent basement-cored island and shoal complex located 5–15 km offshore. The faults controlled the shape of the valley networks and the amount of incision along the valley profile. They were probably active until the end of incision, because the valley thalwegs show scarps up to 10 meters high where they are crossed by these faults. The valleys were incised during the Quaternary lowstands of sea level, and most of the fill was emplaced during the last postglacial sea-level rise. The valley fills form a transgressive succession, consisting mainly of fluvial deposits at the base (possibly amalgamated from older sequences) overlain by tide-dominated estuarine deposits and capped by offshore muds. The most prominent internal surfaces are the tidal- and wave-ravinement surfaces. The valley-fill architecture is strongly dependent on the valley morphology (depth of incision, width of the valleys, and extent of estuarine intertidal areas). Estuarine deposits inside narrow and linear valleys are mostly aggrading muds, whereas those inside large and dendritic valleys dominantly comprise sandier, tidal-channel and bar deposits.

INTRODUCTION

Late Quaternary shelf deposits provide the most precise records of environmental changes triggered by rapid and high-amplitude sea-level changes (e.g., Demarest and Kraft, 1987). The associated depositional sequences are best preserved within incised-valley systems (e.g., Ashley and Sheridan, 1994; Thomas and Anderson, 1994). In this paper we examine an incised-valley system that was filled during the last sea-level rise with mainly estuarine to marine deposits. Transgressive estuarine deposits dominate within the valleys incised offshore on passive-margin continental shelves. The eastern coast of the USA provides good examples of these valleys (e.g., Belknap et al., 1994; Guttierrez et al., 2003). The most prominent surfaces inside these deposits do not correspond to sequence-stratigraphic key surfaces but are controlled by hydrodynamic changes, namely wave and tide ravinements (e.g., Dalrymple et al. 1992, 1994 ; Foyle and Oertel, 1997; Allen and Posamentier, 1993, 1994; Reynaud et al. 1999) as synthesized by the Zaitlin et al. (1994) model. These changes reflect changes of the landscape of the valleys as they are flooded by the sea.

In what way and how much may the character and architecture of valley-fill deposits change as valley morphology changes? Few previously published studies demonstrated the influence of second-order factors on valley-fill architecture, such as sediment supply or coastal morphology (e.g., Ricketts, 1990; Lobo et al., 2001). One way to address the problem is to study adjacent valleys that have a similar history but differ strongly in shape from each other due to local basement controls. Valleys incised along the SE coast of Brittany (France) provide examples of this. The aim of this paper is to synthesize data showing the shapes of the submerged valleys and the architecture of their infill. It aims to interpret the observed differences and similarities in terms of the control that the valley morphology exerts on hydrodynamics during the last sea-level rise over the area. It is postulated that the incised-valley model of Zaitlin et al. (1994) is relevant for appreciating these differences and similarities (Proust et al., 2001; Menier, 2004). Because this study relies mostly on seismic evidence, there will be no attempt to question the model from a facies point of view.

The southern Brittany coast and related shelf have experienced a very low average rate of subsidence over the last 40 Ma, and, except over short periods, the shoreline might have remained at its present-day location (Guillocheau et al., 2003). The rivers in this area have a graded depositional profile (Bonnet et al., 2000). They supply a small amount of sediment to the shelf,

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which is mostly bypassed to the outer shelf margin and deep sea during lowstands. After a long continental planation stage which started 30 million years ago (Guillocheau et al., 2003), incised valleys developed during the Quaternary glacial cycles when sea level reached as low as -120 m (Fairbanks, 1989). Lowstand rivers did incise the shelf down to 70 m below present sea level (Boillot et al., 1971; Menier, 2004); beyond that depth, because of the very low gradient of the shelf (Vanney, 1977), they may have adapted to the changing base level by sinuosity adjustments (cf. Miall, 1991; Schumm and Ethridge, 1994; Thorne, 1994).

The four submerged valleys studied in this paper constitute the seaward extension of the main river valleys of southern Brittany (Fig. 1). They are developed and preserved mostly in the large bays located between the present-day coast and a basement shoal complex located 5–15 km offshore between a series of islands (Figs. 1, 2). Between the coast and across the basement shoal line, the average gradient of the shelf increases from 0.057° to 0.12° (Fig. 2). Farther seaward, the gradient of the shelf is 0.038°. This feature has a direct influence on the amount and location of incision and infilling of the submerged valleys. Several of these valleys have been studied in detail (Proust et al., 2001; Loget, 2001), and most of the results have been reported in Menier (2004). However, the present paper is the first attempt to (1) demonstrate the inheritance of faults in shaping the valleys and (2) examine the interaction between the topography of the transgressed estuaries and the tidal wave that enter them and the influence that this has on the nature of the valley-filling deposits.

STUDY AREA

The south Armorican shelf (NW–SE) is a morphological segment of the northern margin of Bay of Biscay (Montadert et al., 1971; Debyser at al., 1971; Derégnaucourt and Boillot, 1982; Thinon et al., 2001). This margin is located on the margin of the Armorican Massif, an old block of the Variscan chain (Le Corre et al., 1991), composed mainly of crystalline rocks (granites, gneisses, and mica schists). Old faults of this domain were reactivated several times since the rifting of the Bay of Biscay in the Cretaceous, and in response to the Pyrenean and Alpine collisional phases (Debyser et al., 1971; Montadert et al., 1979). In the studied

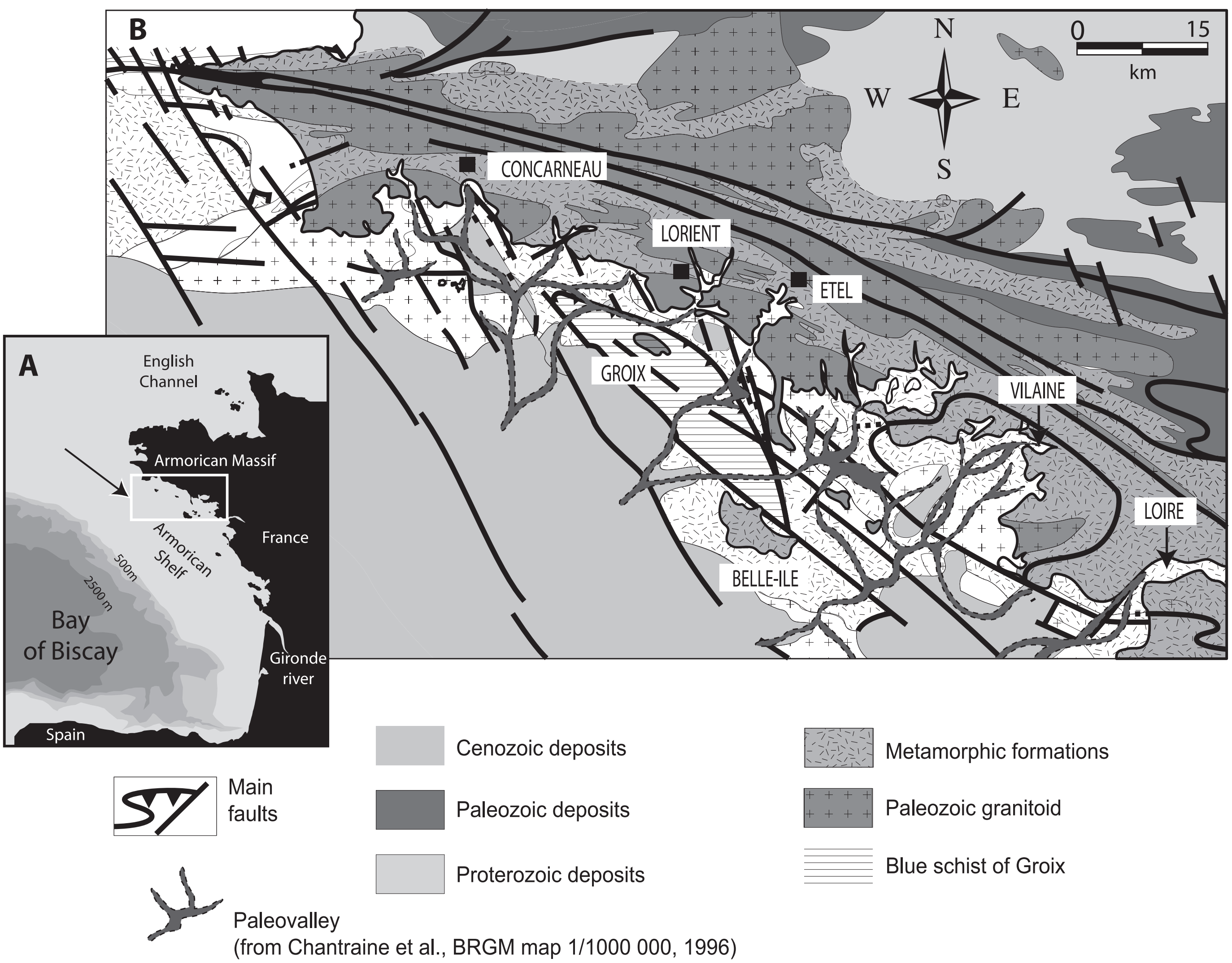


FIG 1.—**A)** Location of study area, northern Bay of Biscay. **B)** Geological map of southern Brittany (from BRGM, 1996).

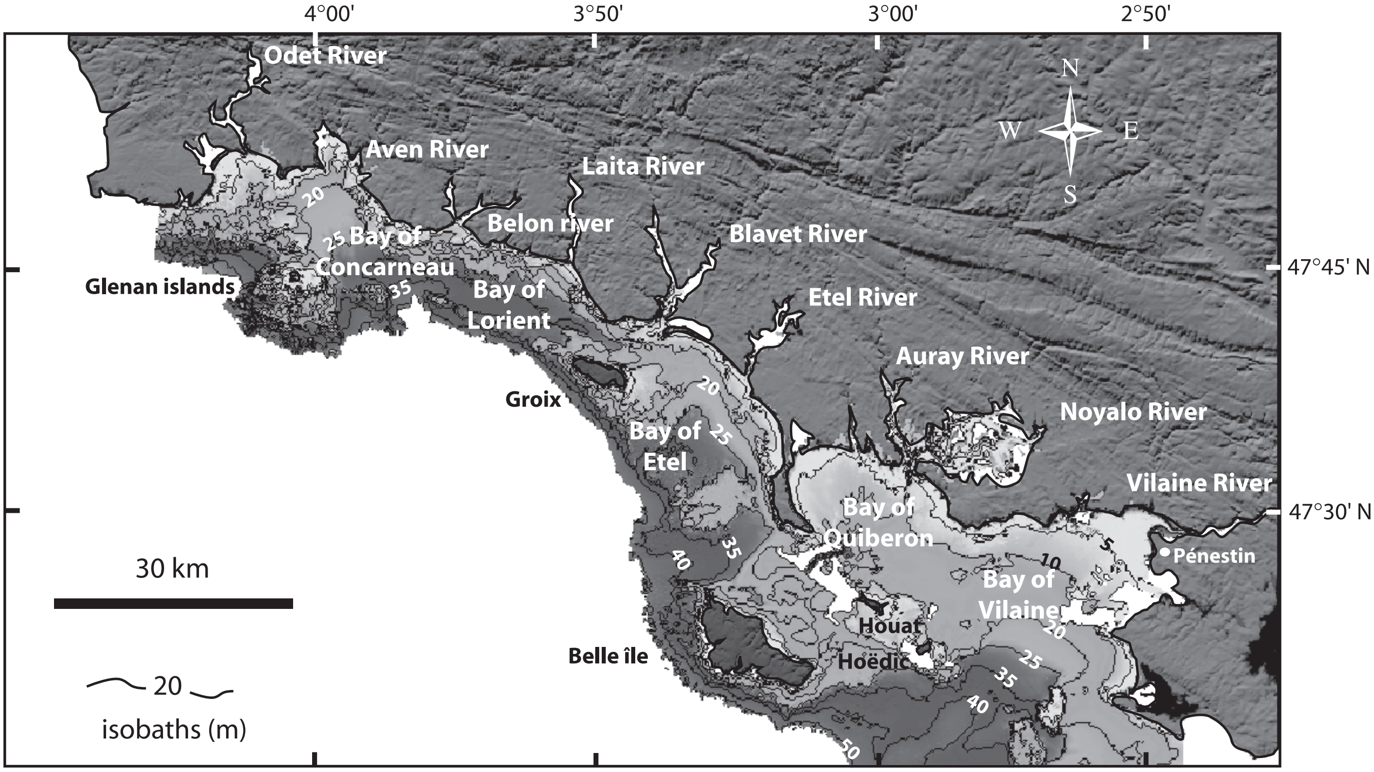


FIG 2.—Terrestrial and submarine topography of the study area. Topography data courtesy of IGN (National Institute of French Geography, and bathymetric data courtesy of the SHOM (Oceanographic Survey of the French Navy).

area (Fig. 1; Chantraine et al., 1996), preexisting Mesozoic deposits have been removed by uplift of the rift shoulder, and deposits lying on the crystalline basement are mostly Paleogene clastic limestones that were deposited and preserved mainly in areas seaward of the present-day coast (Fig. 1).

The morphology of the coast is complex, marked principally by the presence of a number of rivers flowing parallel to each other (Fig. 2). The most prominent feature of the morphology offshore is a series of islands and basement shoals oriented parallel to the regional major N120°-trending faults inherited from the Variscan orogeny (Figs. 1, 2). This feature is a series of horsts (Pinot, 1974) surrounded by a depth of water of 40–50 meters. This morphology creates five bays, each 20 meters deep (Fig. 2), that lie between the coast and the horst. The faults were reactivated episodically during the Cenozoic (Vigneresse, 1988; Le Corre et al., 1991; Caroff et al., 1995), mainly in response to NW–SE-oriented Alpine compression (Bevan and Hancock, 1986; Muller et al., 1992; Hibsch et al., 1993). Some of them still have a seismogenic signature (Lenôtre et al., 1999; Perrot et al., 2005). Other groups of basement faults control the morphology of the mainland and shallow offshore areas (Menier, 2004). These faults were probably active during the former stages of valley evolution, bringing about a general uplift of the area during the Pliocene that favored valley incision in the hinterland (Bonnet, 1998; Bonnet et al., 2000) and sediment delivery to the deep shelf (Hommeril et al., 1972; Guillocheau et al., 2003; Proust et al., 2001; Menier, 2004). In this paper, however, the faults are considered only as the cause of the major morphological differences between the valleys, not as active controls on sedimentation.

The submarine valleys are mostly incised in the bays located between the coast and the basement shoal complex, between the coast and 70 m below sea level (Fig. 2). The amount of sediment delivery to the bays by modern rivers is small, and is mostly trapped in estuarine and lagoonal basins such as the Morbihan Gulf (no data currently published). The present-day dynamics at the coast is wave-dominated, locally tide-influenced. Tides are semidiurnal with a mean range of 4–5 meters. Storm waves show an average height ranging between 1 and 3 meters and approach the coast from the NW (winter) and the SW (summer). As a result of the intense wave action, the Holocene mainland coast northwest of Quiberon consists of wave-dominated beaches subjected to strong littoral drift (Fig. 2). As a consequence, coastal deposits are dominantly high-energy lithic sands and gravels, with muddy facies only in sheltered bays or estuarine tidal flats (Vanney, 1977; Salomon and Lazure, 1988).

METHODS AND DATA

Seismic Data

The sea floor was mapped using high-resolution digital bathymetric data (Fig. 3). The paleovalley network and the deposits above the incision surface have been known for three decades (Horn et al., 1966; Bouysse and Horn, 1968; Boillot et al., 1971; Bouysse et al., 1974; Pinot, 1974; Lefort, 1975; Delanoë and Pinot, 1974, 1977; Vanney, 1977; Audren and Lefort, 1977). Due to the lack of resolution of earlier seismic data and to the wide spacing of seismic lines, these data could not provide detailed information on the morphology of the drowned valleys, or of the

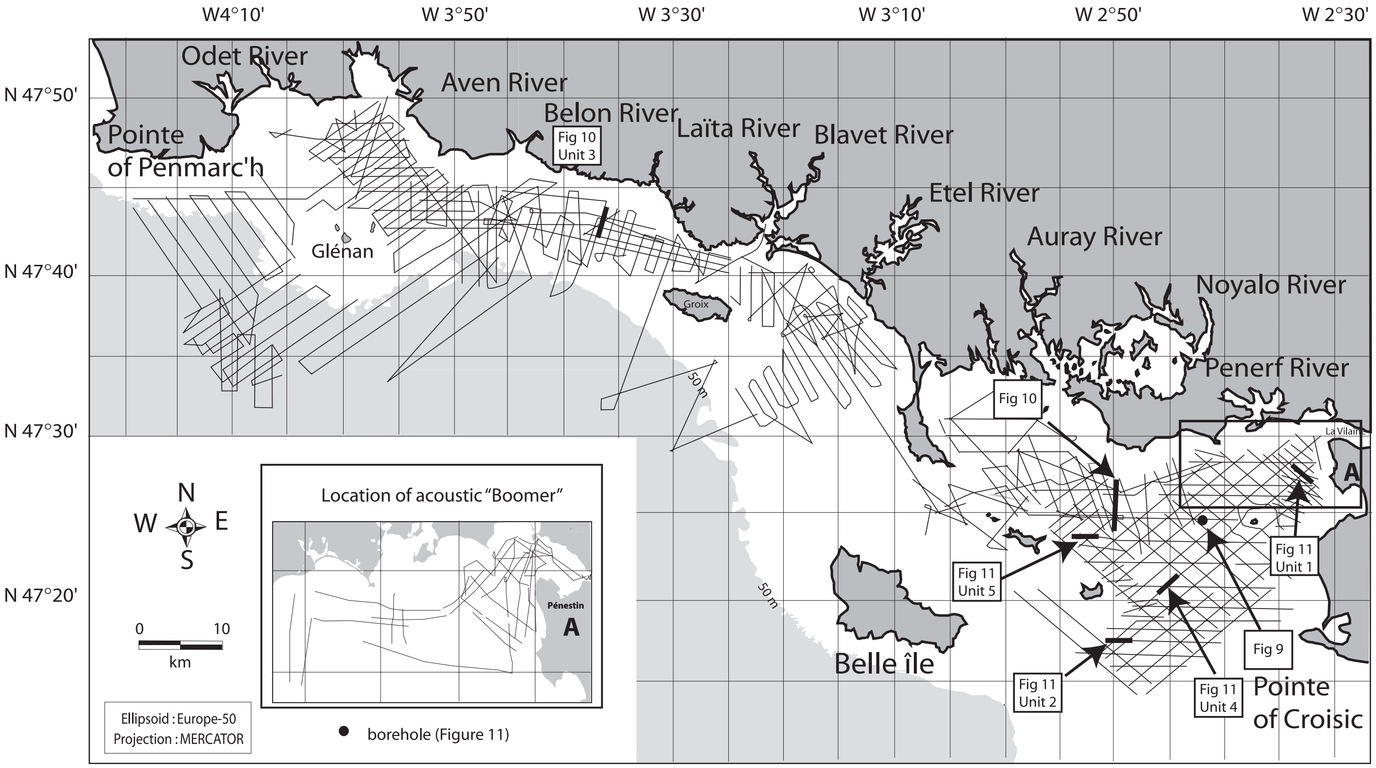


FIG 3.—Location of studied seismic profiles; **A)** sparker, and **B)** boomer. Bold line segments indicate locations of profiles presented in this paper.

facies and geometry of the infilling deposits. These details have become available by the acquisition of 3000 km of digital, very high-resolution reflection seismic data (sparker, 600–1000 Hz, 650 J; boomer, 0.5–10 KHz, 200 J) between the coast and the 50 m isobath. All tracklines were located using Digital Global Positioning System with a positional accuracy of ± 10 meters (Fig. 3). The area was surveyed over a square grid of profiles spaced 200 m to 500 m apart, and oriented parallel and transverse to the supposed main valley axes. Time–depth conversion was performed by assuming that the velocity of sound was 1500 m/s in water and 1800 m/s in the sediment. The relevant surfaces (sea-bottom and subsurface) have been digitized and interpolated on a computed GIS. The seismic profiles have been merged with the data available onshore to discuss the longitudinal evolution of the valley system in response to base-level fluctuations, the lithologic nature of bedrock, and eustatic sea-level variations.

Seismic profiles were interpreted following the guidelines of Mitchum et al. (1977) and Sangree and Widmier (1977). The most prominent erosional unconformity traceable over the area is the incision surface at the base of the valley network. This surface cuts down to Paleozoic crystalline rocks and Paleogene strata. Mapping this surface allowed us to identify five main valley systems and to extract for each valley the longitudinal profile of the thalweg and the adjacent interfluve area (Fig. 4). We have identified five unconformity-bounded units (as summarized in Table 1) above the valley incision surface and its correlative surface on the valley interfluves (Table 1). The seismic units are assumed to correspond to stratigraphic units. These units largely overlap to each other, but all of them can be observed on several seismic profiles.

Core Data

A series of gravity cores and vibracores provided lithologies of most of the substratum and deposits of the valleys described from the seismic profiles. Grain-size analysis, radiocarbon dating, and lithologic and fauna (foraminifera) descriptions were performed on core samples. The result led to a number of papers that highlight the new seismic data preserved in this paper (Andreieff et al., 1968a; Andreieff et al., 1968b; Bouysse et al., 1966; Bouysse and Le Calvez, 1967; Delanoë and Pinot, 1977; Lefort, 1975; Vanney, 1977; Visset et al., 1995; Visset et al., 1996).

ORIGIN OF THE VALLEYS

A Terraced Network of Incisions

The transverse section of the valleys exhibit either V-profiles with a narrow, rounded thalweg or U-profiles with a broad, flat thalweg, locally exhibiting up to one to four morphologic terraces on the valley flanks (Figs. 5, 6; Menier, 2004). There is no concordance between the terraces and the stratigraphical boundaries within the valley fill, so these terraces correspond to erosional surfaces cutting the underlying rocks called strath terraces (Bull, 1990, already noticed by Lefort 1975, 1978, and Jouet et al., 2003). They are exposed at the sea bed or buried by more recent deposits. The terrace elevations range from between 15 m to 50 m below sea level and between 5 m and 15 m above the elevation of the adjacent thalweg. Within each valley network, the terraces dip on average in the same (seaward) direction as the valley thalweg. Their width may change from nearly zero to a few hundreds of meters

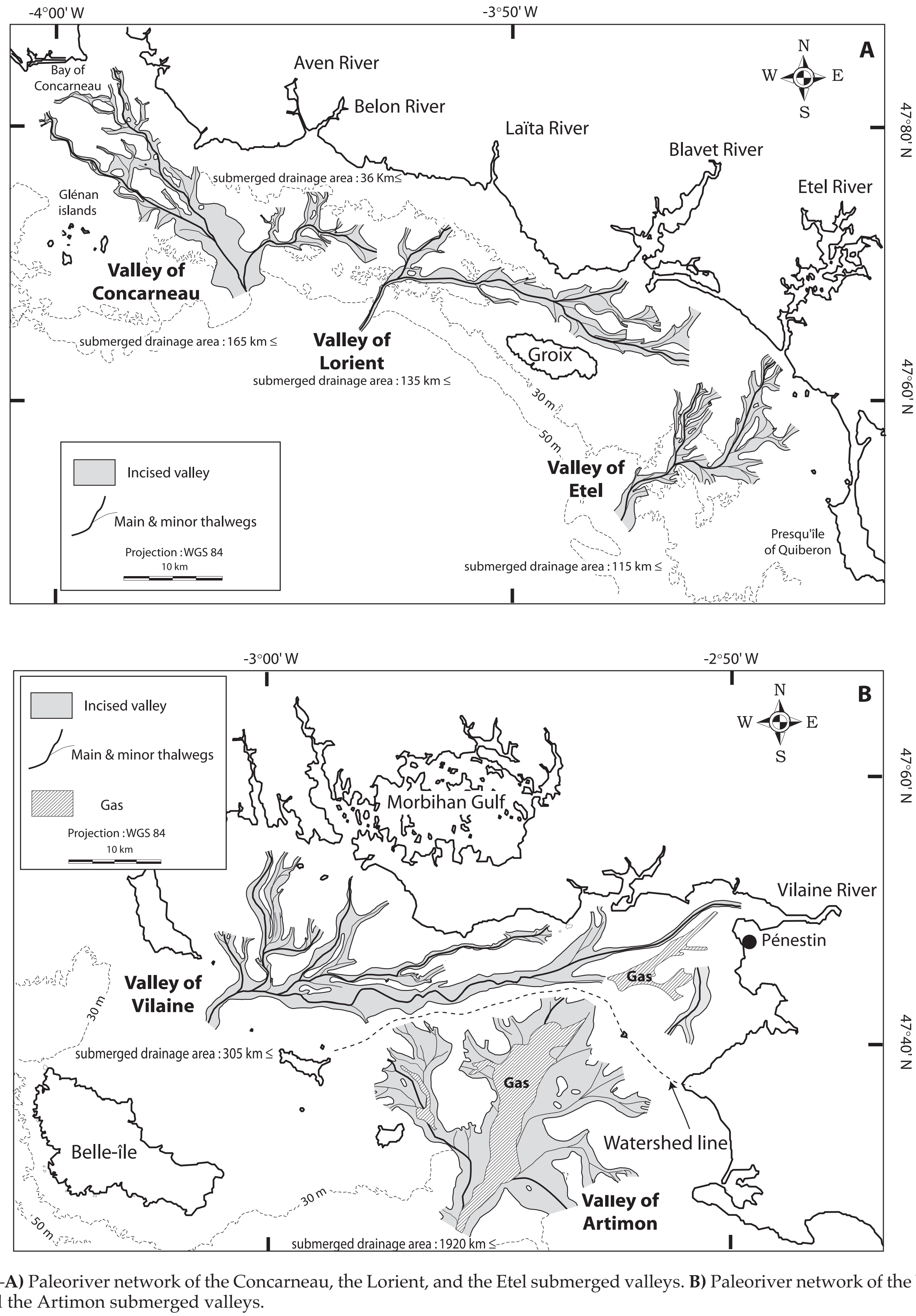


FIG 4.—**A)** Paleoriver network of the Concarneau, the Lorient, and the Etel submerged valleys. **B)** Paleoriver network of the Vilaine and the Artimon submerged valleys.

TABLE 1.—Characteristics of seismic units and their facies interpretation (F, frequency; A, amplitude; C, continuity; T, top; B, bottom) after Proust et al. (2001) and Menier (2004). See Figure 10 for illustration of terms defining reflector geometry.

Seismic units	Reflector characters	Principal internal reflector characteristics	Type of reflectors terminations	Max thickness (m)	Geological interpretation
U1	F : Middle A : High C : Middle	sigmoidal and oblique-parallel to subparallel	T : toplap B : downlap andonlap	15	Gravels and sands of fluvial deposits
U2	F : Middle A : High C : Middle to high	oblique-parallel, oblique-tangential to subparallel	T : toplap B : downlap	10	Gravels and sands of shoreface bars
U3	F : Low A : Low C : Low	oblique tangential to subparallel aggradational	T : truncation B : onlap and downlap	20	Sandy to clayey estuarine muds
U4	F : Middle A : Middle C : Middle to low	small oblique parallel to divergent local chaotic oblique-sigmoidal to oblique parallel	T : toplap and truncation B : onlap and downlap	10	Clayey channel muds Sandy to clayey point bar
U5	F : Middle to low A : Middle to low C : Middle	subparallel aggradational to divergent medium to high oblique-parallel	T : concordant and toplap B : downlap and onlap	30	Muds with turritelids Silt and clay Sand banks

(Figs. 4, 5). Where well expressed, they commonly occur at the same depths on both sides of the valley, forming paired terraces. In that case, however, the terrace on one side of the valley is much wider than that on the opposite side. More commonly, because the valley is sinuous in plan view, the terraces are found alternating on opposite sides of the valley (Fig. 6). Similar terrace systems are present within the coastal valleys onshore (Bonnet, 1998). The terraces observed offshore would likely be their counterparts.

Interpretation

We interpret the terraces as having been formed by the lateral planation of the rivers during one or more periods of sea-level fall and/or lowstand, as commonly described in many examples (see Thorne, 1994). This is supported by the fact that they can be traced over tens of kilometers along the valleys, and they do not depend on local erosion. Each terrace level may represent a level of fluvial entrenchment at a given time, the upper terraces being older than the lower ones. Similar terraces in the Seine and Somme valleys have been correlated to the high-amplitude 100 kyr sea-level cycles (Antoine, 1994). For the southern Brittany valleys, this interpretation is supported by their shape in plan view, which led to a drainage network of varying sinuosity (Figs. 5, 6). One could consider, as a hypothesis, that each terrace was formed by fluvial incision and lateral planation during the maximum lowstand of the last eustatic sea-level cycles of large amplitude (fourth order; 100 kyr and > 100 m sea-level fall). The last sea-level fall at 20 ka would be responsible for the last and lowermost incision, which corresponds to the thalwegs of the valleys. An earlier network of valleys that would have existed at that place would have been almost completely overprinted during this last phase of incision.

This interpretation has to be confirmed by a more detailed correlation of the terraces and their comparison from one valley to the other. The uplift of the hinterland has been demonstrated from an analysis of the base-level profiles of the valleys onshore (Bonnet, 1998; Bonnet et al., 2000). This uplift could have been active as far south as the present-day submerged proximal offshore area. Almost the total amount of sediment produced by fluvial erosion over the entire length of the incised valleys appears to have been exported farther seaward, nourishing the lowstand shelf-edge deltas or feeding the slope canyons, as demonstrated for the deep shelf of the Western Channel Approaches (Pantin and Evans, 1984; Reynaud et al., 1999; Zaragosi et al., 2000).

CONTROLS ON VALLEY MORPHOLOGY

Thalweg Profile of the Valleys

The longitudinal profiles of the thalwegs of the main valleys are represented in Figure 7. They correspond to the seaward extension of the present-day profile of the valley in the modern estuaries. They are composed of two segments. The entrenchment of the incised valley corresponds to the altitude difference between thalwegs and interfluve. The landward segment is relatively flat in the Concarneau and Vilaine valleys. The Lorient

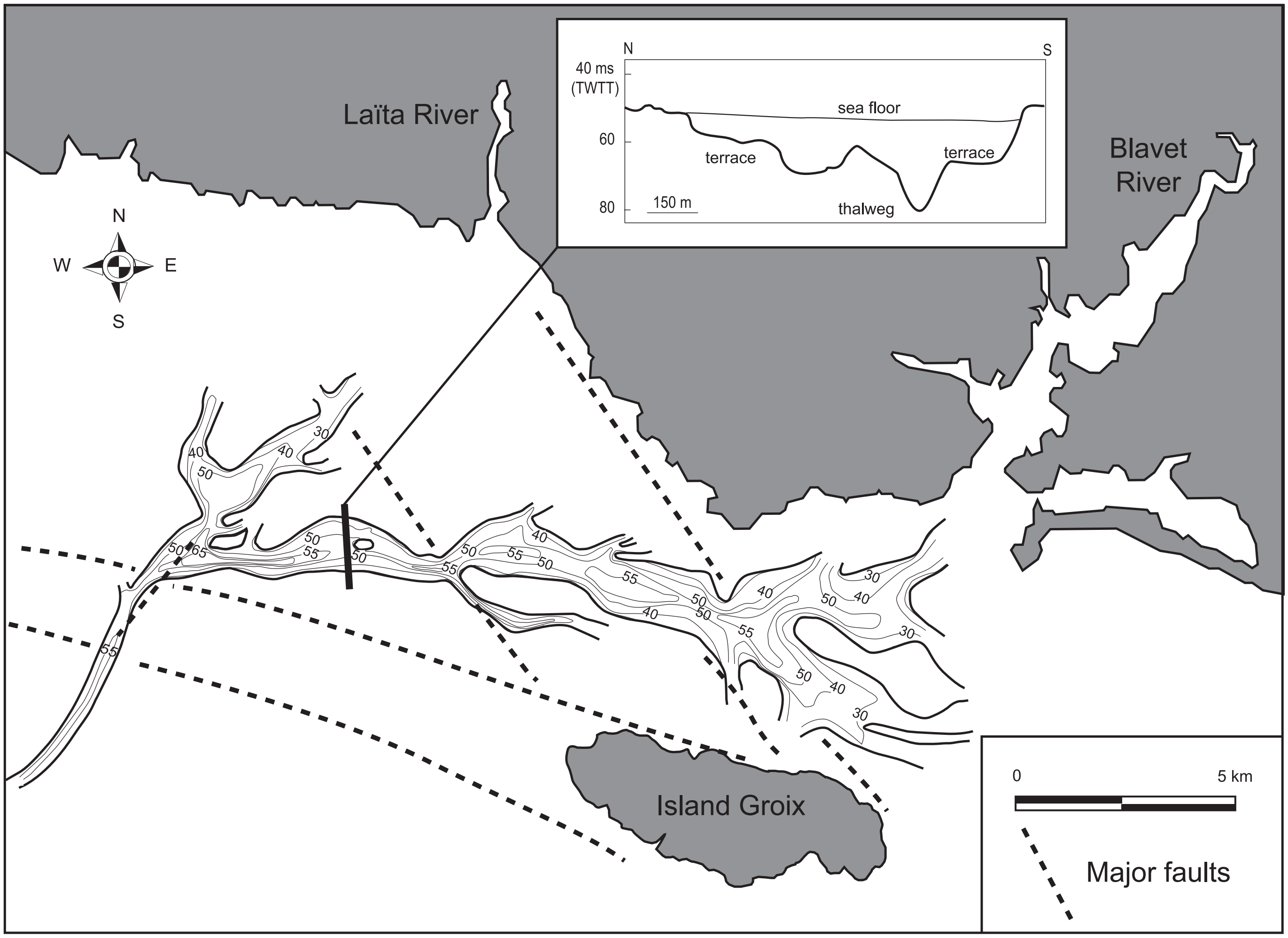


FIG 5.—Contour map of the surface at the base of the Lorient valley. Contours in meters, based on seismic data using a sediment velocity = 1800 m/s. Dashed lines indicate major faults as mapped from seismic data. Inset: cross section of the valley showing thalweg, interfluves, and terraces.

and Etel valleys are steeper in their landward part and flatten seaward. The lower, most seaward part of this segment is about 30 m deeper than the knickpoint (particularly in the Concarneau and Lorient valleys). The same profiles are observed from one valley to the other but with a vertical offset pointing to a general deepening from the east (knickpoint at 20 m below present sea level in the Vilaine valley) to the west (knickpoint at 35 m in the Concarneau valley). Locally, the profiles exhibit "saw tooth" sections, with an incision depth locally varying up to 80% of the total amount of incision (Fig. 7). The thalwegs may even dip landward over short distances. While substratum lithology has little if any influence on the longitudinal profile of the valleys, the "accidents" in slope profile are located (1) at the junction between two valley branches and (2) where N120° or N160° faults cross the valley (Figs. 5, 6, 7). The latter case seems to be the dominant control, inasmuch as the interfluves of the valleys follow the same offset pattern. Where they cross major faults, the valley thalwegs may show abrupt elevation changes of up to 10 m . Today, such dip inversions in the thalweg profiles are not demonstrated in the emerged sections of the valleys, perhaps because they are masked by sediments.

Major faults cross the study area, related to the southern part of a major Variscan shear zone (Fig. 1, faults N120°) that has been reactivated several times during the Neogene (Boillot et al., 1971; Pinot, 1974; Vanney, 1977). The basement shoal complex is bounded by such faults (Figs. 1, 5). It is worth noting that the knickpoint of the thalweg profiles coincides with this structural line. The longitudinal profiles of the thalweg of the valleys exhibit abrupt changes in slope that superpose to major regional faults (Figs. 6, 7). Relying on the idea that the thalwegs of the valleys were incised in response of the last glacial sea-level lowstand, the faulting may even have been active until short a time before final incision of the thalweg (so that the related scarps could not have been planated by fluvial processes) or even after incision ceased. This is supported by (1) the saw-tooth profile of the thalwegs and (2) the local offset of lowermost deposits inside the valley by the faults (Proust et al., 2001). However, there is no clear evidence of faulting that would have affected the valley fill above these lowermost deposits. The faults occur in the submarine topography as scarps, the landward part of which corresponds to bedrock outcrops. The same pattern applies onshore, where terraces are offset by the faults (Bonnet, 1998; Bonnet et al., 2000).

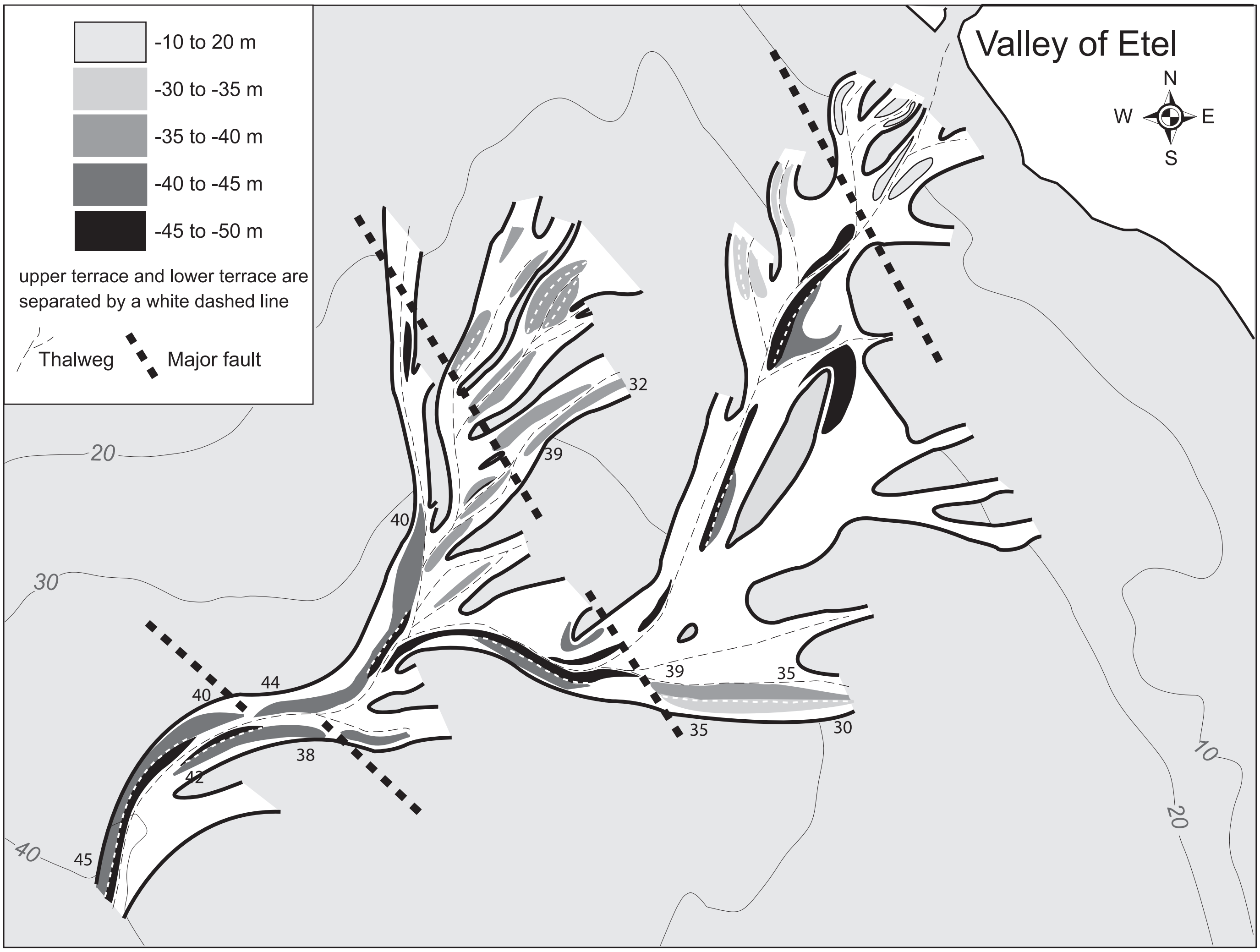


FIG 6.—Mapping of strath terraces of the Etel valley.

Fault Controls on the Valley Paths

Where faults do not affect the valley substrate and the seafloor morphology is planar, the rivers develop a dendritic drainage pattern (Small, 1978; e.g., Artimon Valley; Table 2, Figs. 4, 8). In this case, the trunk valley is wide because the substrate is softer and more easily eroded. Elsewhere, the paths of the incised valleys are controlled dominantly by faults (Fig. 8). Similar controls are reported in the literature (e.g., Dam and Sonderholm, 1998). Valleys either follow the main faults (Fig. 8) or are dissected by them. Where they run parallel to the faults, this may be due to substrate weakness along the fault zone or to the tilting of fault blocks, which captures the river and causes it to flow adjacent to the fault. As rivers follow the faults, they exhibit a linear longitudinal network (e.g., Lorient River, Figs. 4, 5, 8), whereas they may have a rectangular network with a typical zigzag path of the main thalweg where the valley is oblique to the faults and is locally dissected by them (e.g., Etel River, Figs. 4, 6, 8). Where the valleys follow the faults, the valleys run almost parallel to the coast for considerable distances (Figs. 4, 8). Because the faults do not extend farther seaward than the basement shoal complex, the valleys in this area exhibit a simple pattern and run downdip to the shelf margin (e.g., Etel Valley, Figs. 4, 6, 8).

The amount of entrenchment of the valleys matches what is expected from their pattern in plan view (Table 2). The more they are confined into faulted corridors, the higher the amount of incision. This amount of incision increases from E to W, correlatively with the elevation of the emerged mainland landward of the valley (Table 2, Fig. 7).

INFILLING OF THE VALLEYS

One of the cores recovered from the study area shows that the upper part of the valley fill is composed of the marine deposits (Fig. 9). The uppermost deposit is a homogeneous turritelid-rich mud interpreted as an offshore deposit. The lowest part of the core contains coarse glauconitic sand with mud drapes. The mud drapes in this lower facies express abrupt hydrodynamic changes that suggest a shallower depositional setting. Both deposits are bounded by a pebble lag with shell debris that have been dated at of 8110 yr BP. This lag is interpreted as a transgressive ravinement surface.

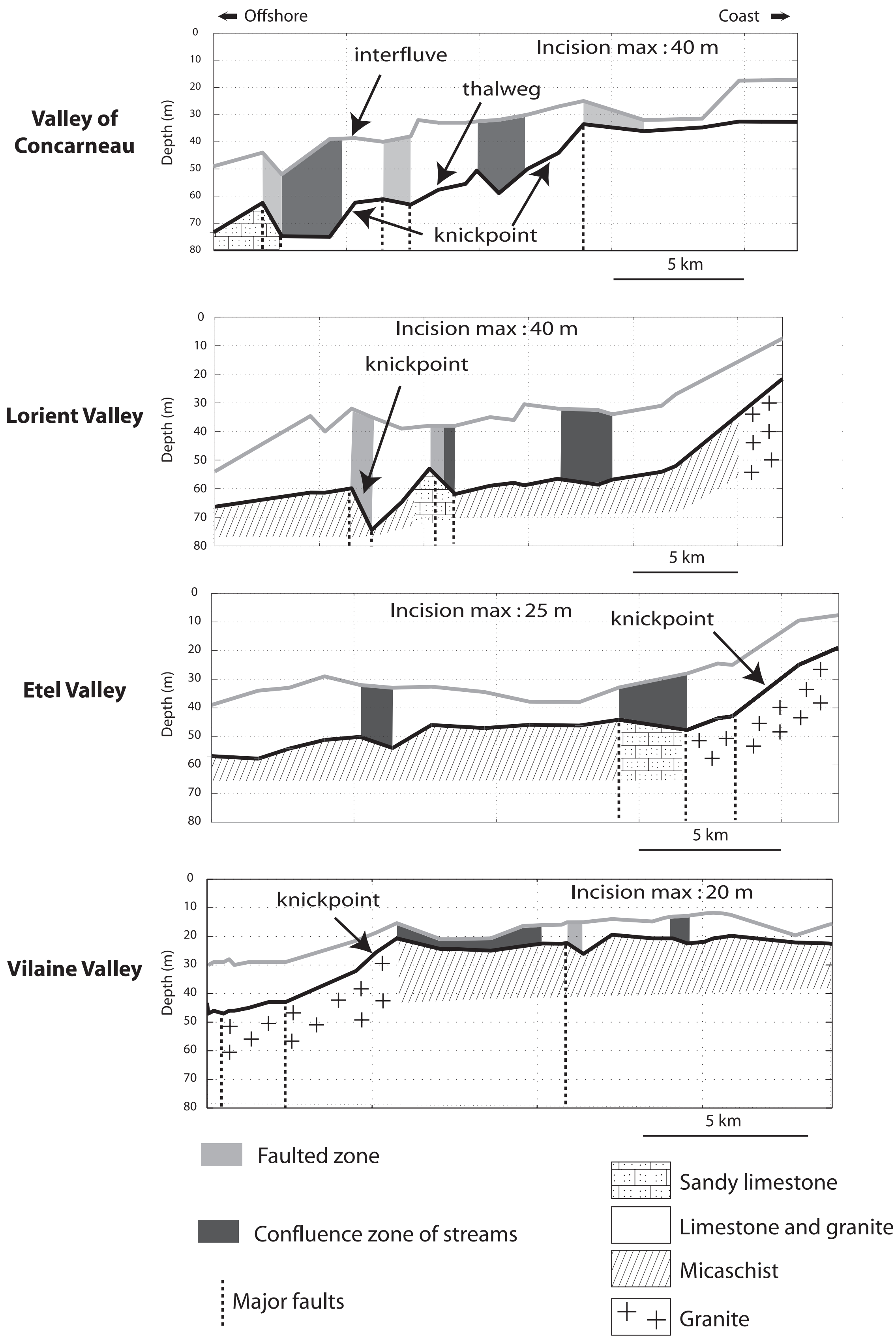


FIG 7.—Longitudinal profiles of the thalweg and adjacent interfluves of the main valleys (as located in Figure 4).

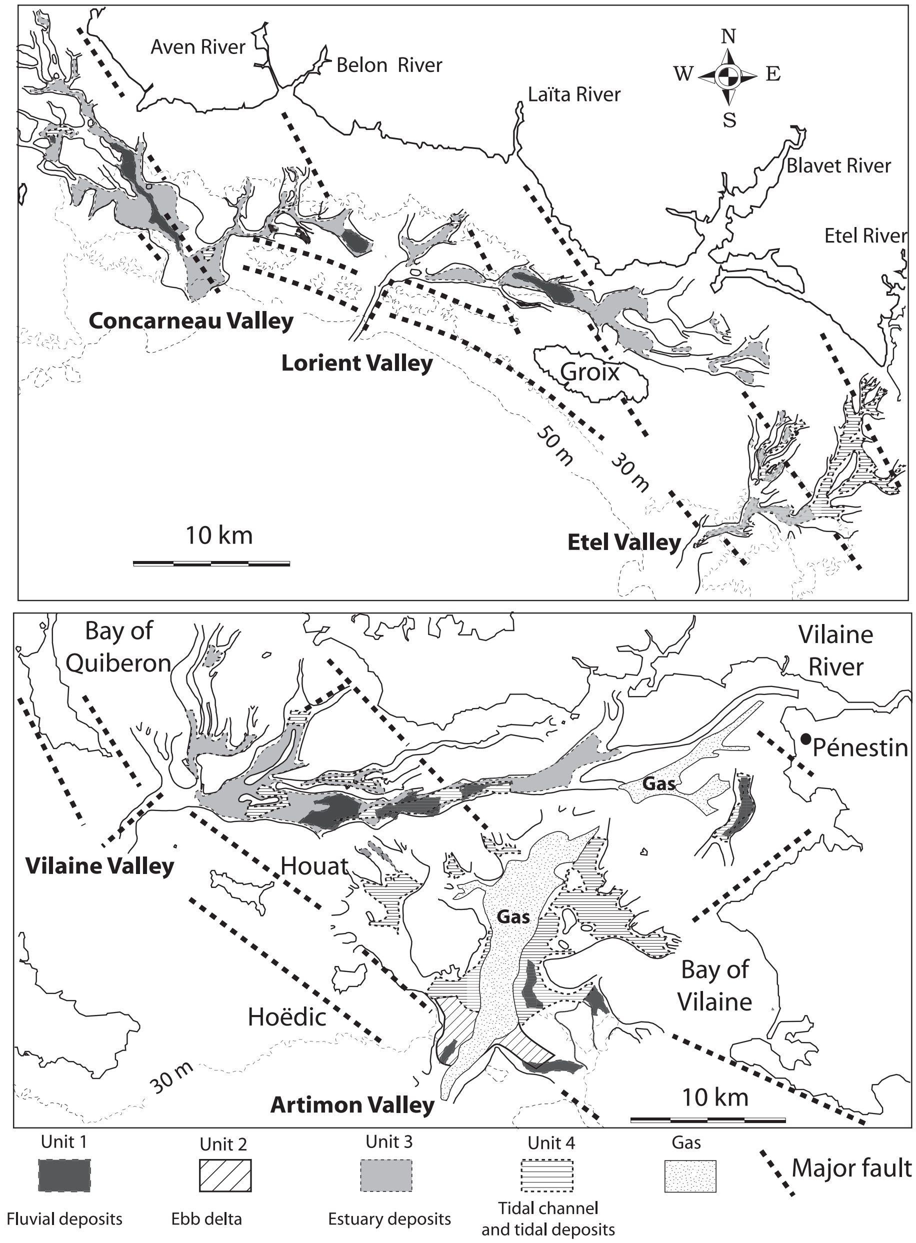


FIG 8.—Areal distribution of seismic units (1–4) in the valley fills between the coast and the 50 m isobath.

Depositional Units

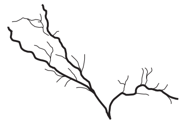
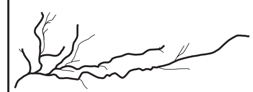
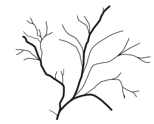
Five seismic units have been found within the valley-fill deposits. They comprise geographically extensive depositional units that are defined mainly on the basis of their seismic facies (amplitude, frequency, and continuity) and reflector shape and terminations (Table 1; Figs. 10, 11). These results are based on the study of the Vilaine and Artimon River valley fills (Proust et al., 2001) and have been extended to the whole surveyed area (Fig. 3). This permits an examination of how the infill varies from one valley to another and a discussion of the controls on origin of the observed variations.

Unit 1 is a relict unit preserved as patches only in the deeper parts of the valleys, bounded at the top by a truncation surface (Table 1; Figs. 8, 10, 11). This unit exhibits mostly flat, parallel reflectors of variable amplitude and frequency, dipping toward the thalweg of the valley at a very low angle. These dominant reflectors may cut low-relief trough-shaped reflectors of about 100 m lateral extent.

The deposits of unit 1 have been geometrically correlated to braided-river deposits that crop out onshore near Pénestin (Fig. 8). The general flat bedding of reflectors, the rare low-relief channels, and the low-angle sigmoids indicating some lateral accretion are consistent with the idea that unit 1 consists of braided-river deposits.

Unit 2 overlies either the valley basement or unit 1 and is restricted to the most seaward section of the valley of Artimon (Table 1; Figs. 4, 8, 10, 11). It is composed of stacked wedges of

Table 2.—Synthesis of geomorphic and facies information for the southern Brittany incised valleys.

	Concarneau Valley			Lorient Valley			Etel Valley			Vilaine Valley			Artimon Valley		
Drainage pattern	 Linear, parallel, rectangular			 Parallel, rectangular			 Parallel			 Parallel, trellis, rectangular			 Dendritic		
Setting	sheltered by basement highs			sheltered by basement highs			open shelf			sheltered by basement highs			open shelf		
Valley width (min, max; m)	200 to 4000			200 to 2500			300 to 2500			300 to 3000			300 to 7500		
Valley length (from the present-day coast) (km)	25			27			22			50			30		
Main thalweg dip (degrees)	0.05 to 1.19			0.03 to 0.42			0.01 to 0.33			0.02 to 0.93			0.02 to 0.30		
Max. entrenchment (m)*	40			35			25			20			< 20 (Gas)		
Max. thickness units (m)	landward		seaward	landward		seaward	landward		seaward	landward		seaward	landward		seaward
U1		15			10						8			5	10
U2															12
U3	8	18	10		18	10	12	10	12		15	8			
U4							5	3	3		10	10		10	12
U5	15	5	10	20	10	18	10	5	15	10	15	20	10	10	15
Relative reilef of U4 base (TRS)	low			low			high			medium			high		

* The maximum entrechment has been measured relative to the adjacent interfluve

oblique parallel, plane to sigmoid reflectors dipping < 8° toward the sea (southwest) or the valley thalweg over a > 160° span.

Unit 2 is truncated at the top by an erosional surface. It is not in connection with either unit 3 or unit 4.

Unpublished area data reveal that unit 2 is composed mostly of coarse-grained pebbly sand and gravel with abundant shell debris. The abundant shell remains suggest that this deposit is marine in origin. Rounded pebbles are related to high depositional energy. The coarse-grained sediments could be reworked from the underlying fluvial deposits in unit 1. The dip of unit 2 internal reflectors suggests that they are accretion surfaces of lobate sediment bodies that flank basement highs and tend to prograde toward the thalwegs of the valleys. These sediment bodies are therefore likely to correspond to shoreface banks at the seaward end of the valley (Figs. 4, 8, 11).

Unit 3 is present only in narrow valleys, where it forms the main part of the valley fill (Table 1; Figs. 8, 9, 10). It is bounded at the top by a sharp erosion surface and is composed of low-amplitude flat parallel reflectors onlapping the valley walls and locally downlapping to the base of the unit. The reflectors in unit 3 exhibit a very low-angle dip (about 0.1°) toward the thalweg of the valley. They express a centripetal infilling of the valley (from the edges to the thalweg), with dominant sigmoids backed to the valley walls, passing to concave-up surfaces in the valley axis near the top of the unit. In restricted areas, the seismic facies of unit 3 is chaotic or transparent (Figs. 8, 10, 11).

The reflector geometry suggest infilling of the valley by lateral accretion, under the action of currents following the valley paths. The low-amplitude seismic facies and the large-scale homogeneity of the sigmoids as well as their very low angle of dip suggest that these are formed by accretion of muddy sediments, under low to moderate currents that would not be able to channelize the deposit. This is supported by the very low dip of the accretion surfaces. The sigmoids can be traced over two-thirds of the thickness of the unit (about 7 m), suggesting that deposition occurred below such a water depth at least. These arguments point to deposition of unit 3 below sea level in a sheltered area. The weak to moderate currents that are recorded in this unit are likely to be tidal in origin. This suggest that unit 3 was deposited in an estuarine setting as the valleys were drowned by the sea (Figs. 8, 10, 11). The chaotic to transparent acoustic facies points to the presence of gas, which is very commonly reported in clayey organic-rich sediments from estuarine valleys (e.g., Mullins and Halfman, 2001; Garcia-Gil et al., 2002; Baltzer et al., 2005).

Unit 4 corresponds to the lower, sandy glauconitic deposit sampled by the core in Figure 9. This unit has an erosional basal surface that corresponds to the cored pebble lag (Fig. 9), is traceable over the entire area (Table 1; Figs. 8, 10, 11). It is relatively flat in the seaward part of the area, but it delineates a sinuous channel network up to 5 km wide in more landward areas. The amount of erosion at the bases of these channels decreases at the landward end of the unit. Unit 4 is made up of

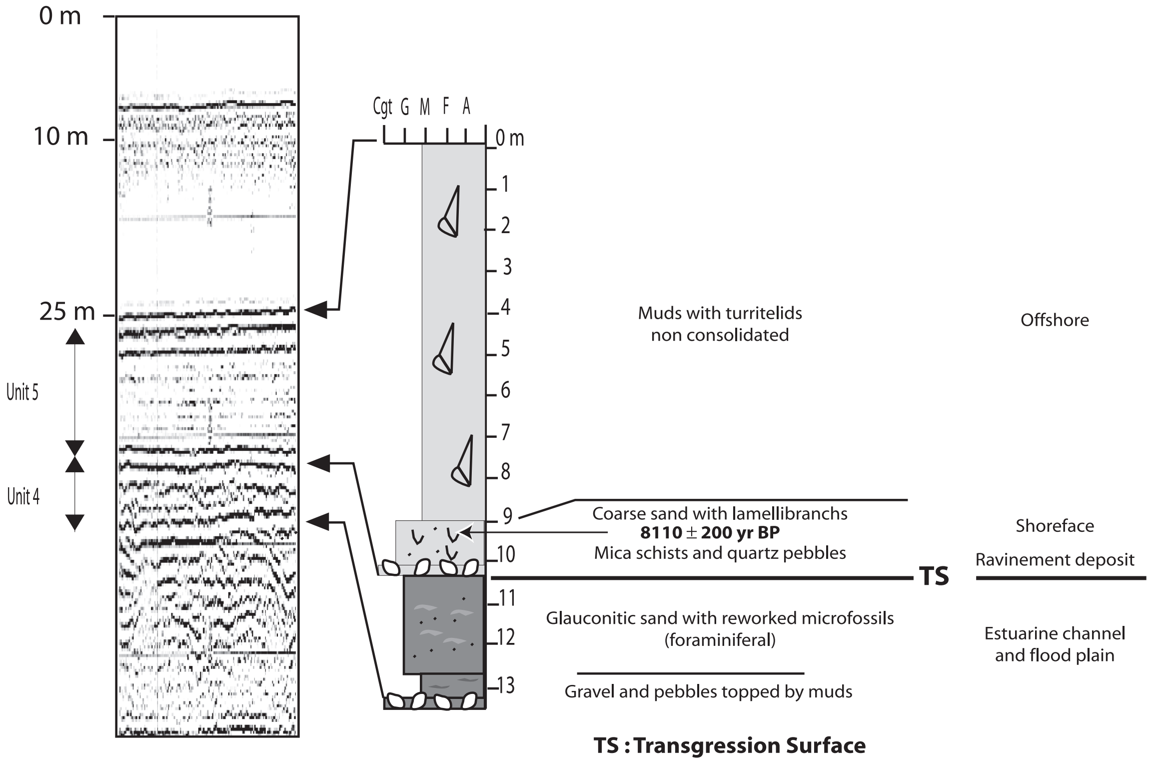


FIG 9.—Sedimentary log of the core located on Figure 3 (Bouysse et al., 1974, modified by Proust et al., 2001) and proposed correlation with a seismic profile obtained across the core location. Glauconitic grains in Unit 4 may have been reworked from older marine strata cropping out in the drainage basin of the Vilaine River (Guillocheau et al., 1998).

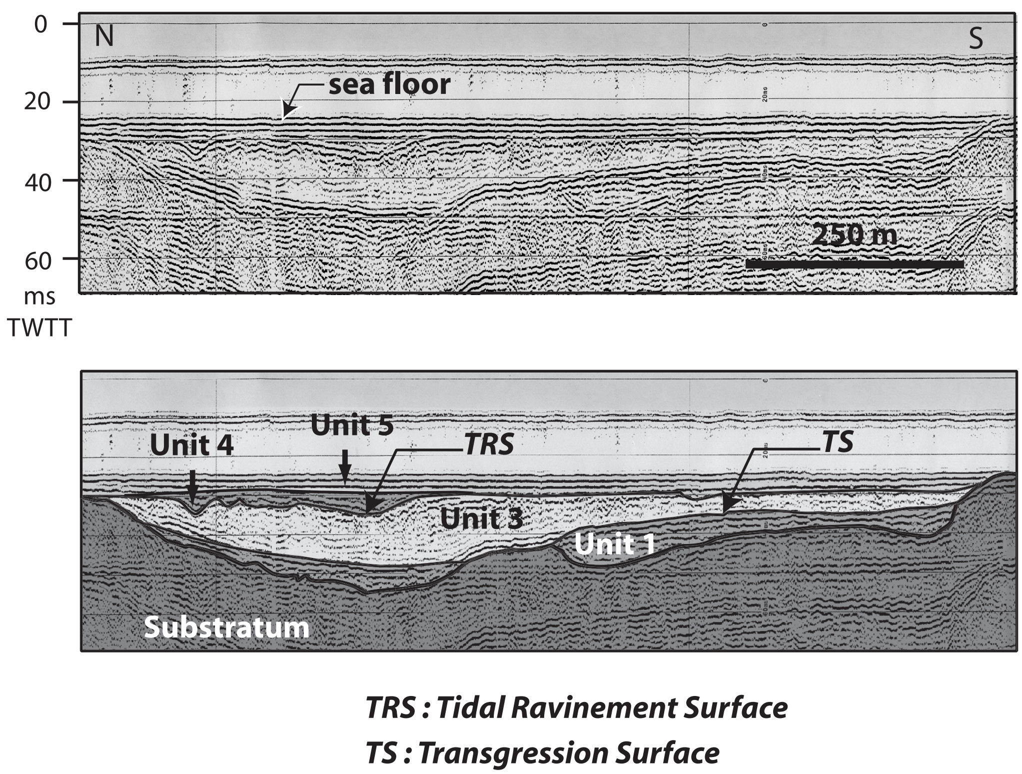


FIG 10.—Example of valley-fill architecture in transverse cross section. See Figure 3 for location of trackline.

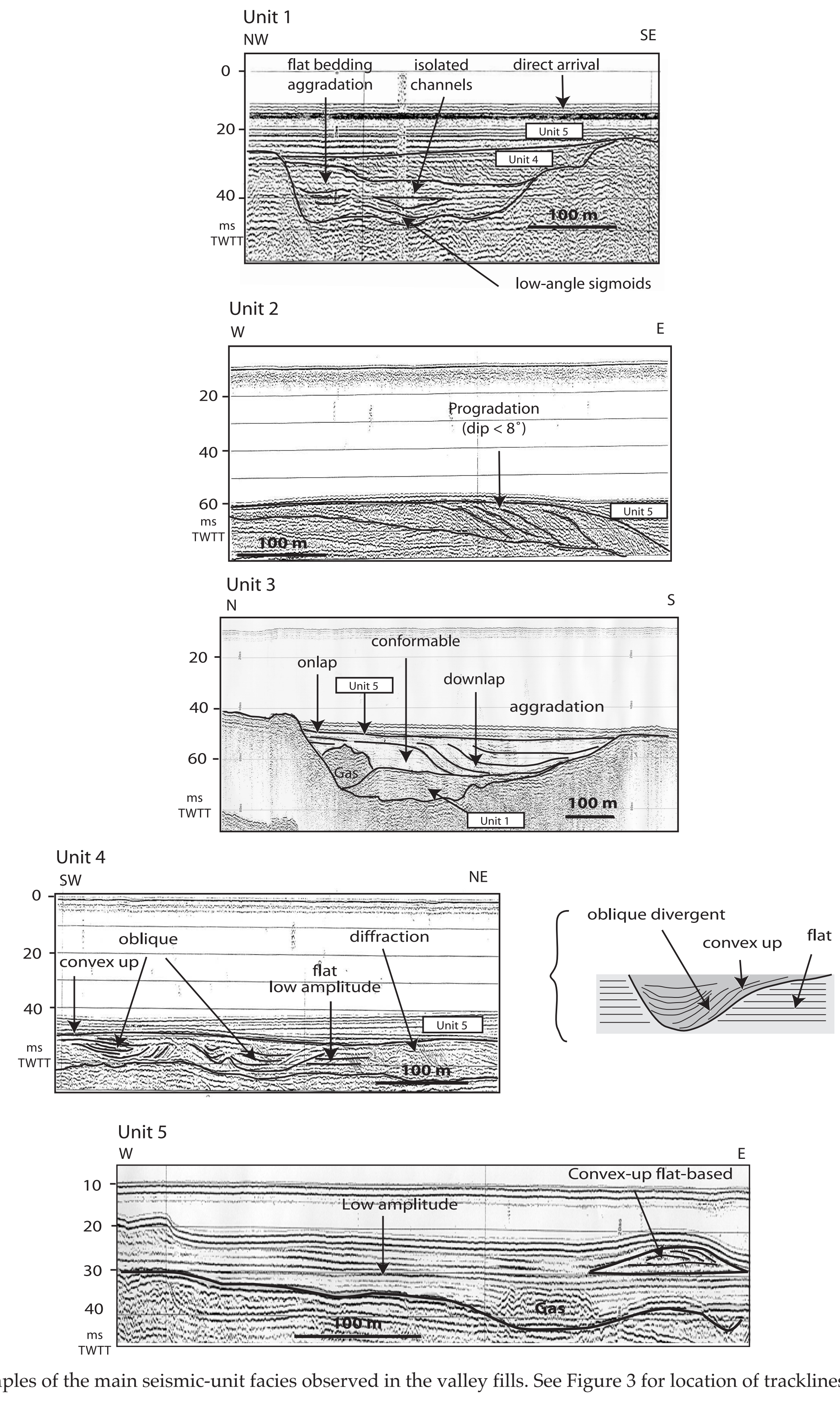


FIG 11.—Examples of the main seismic-unit facies observed in the valley fills. See Figure 3 for location of tracklines.

stacked subunits composed of sigmoid, oblique to flat reflectors delineating cut-and-fill channels up to 300 meters in cross section (Figs. 9, 10). The fills are commonly marked by reflectors of high amplitude, whereas the associated flat deposits may be nearly transparent (Fig. 9).

The deposit sampled in unit 4 contains numerous grains of glauconite and foraminifers (Fig. 9). Although glauconite may be reworked from older marine strata (Guillocheau et al., 1998), the mud drapes in the sands point to abrupt current changes, which are very typical of tidal-estuary environments (mud drapes deposited at slack water and reworked by peak currents). The cut-and-fill pattern of the deposit would likely be due to the onset of a strong tidal current regime in the estuarine environment, leading to dominantly channelized deposits. The higher-amplitude facies the channel fills would match the (muddy) pebble and gravel deposits that are common within tidal channels. In this respect, the base of this unit, marked by a pebble-rich clay layer (Fig. 9), would correspond to a tidal ravinement surface (Figs. 10, 11). The tidal ravinement would be forced by constriction of the tidal flow inside the valley. The decreasing amount of erosion at this surface toward the landward end of the valley is consistent with the landward decrease of tidal energy that has to occur inside the estuaries. Unit 5 corresponds to the upper, muddy deposit sampled by the core in Figure 9. Due to the presence of turritellids, this deposit is interpreted as a quiet offshore setting. Unit 5 extends over the entire valley system, overlapping valley interfluves (Table 1; Figs. 8, 10, 11). It is bounded at the base by a high-amplitude erosion surface that correlates to the top of unit 2. This surface matches the bedrock pebble and shelly lag sampled in the same core (Fig. 9). Most of this unit is composed of flat-bedded, aggrading deposits of low seismic amplitude that may even be transparent in the central parts of the valleys. Landward from Houat, Hoëdic, and Glénans islands (Fig. 1), the upper part of the unit contains flat-based and convex-up depositional shapes, up to 5 m high, that exhibit internal oblique bedding (Figs. 10, 11).

Unit 5, offshore deposits, was emplaced at full transgression over the area, inasmuch as it overlaps most of the valley interfluves. Because of that, tidal flows were no longer constrained within the valleys at that time and tidal currents were thus very low. This is also why this mud is almost flat bedded and aggrading, except for the convex-up depositional flows that locally occur near the seabed and that are interpreted as offshore banks. The flat and widespread lag at the base of the unit is interpreted as a wave ravinement surface. It is preserved at a depth of about -35 m below present sea level (Fig. 9), and dated at 8110 ± 200 yr B.P. From eustatic and local sea-level curves (Morzadec-Kerfourn, 1974; Fairbanks, 1989), this points to a water depth of about 8 m at the time of formation of the wave ravinement surface. This is shallow compared to the range of depth of storm wave base in oceanic borders but is to be related to the presence of the basement shoal that breaks the Atlantic swells seaward from the valleys. This sheltering effect is also recorded in the grain size of the Unit 5 deposit, which is finer landward from the basement shoal than seaward of it.

Depositional Story

While depositional units are not all equally preserved in all the valleys studied, a general stratigraphic scheme can be drawn that applies throughout the entire system (Fig. 12). The core and the major part of the valley fills is composed of estuarine deposits (Units 3 and 4), which are dominantly preserved landward from the basement shoal line. The vertical facies succession is transgressive. Because it has not been possible to sample the lower fluvial deposits (Unit 1), the chronology of the complete succession remains speculative.

The fluvial deposits in U1 are likely to be diachronous, possibly amalgamating several sequences. They are correlated onshore to the Pénestin beds (Fig. 2), which have been dated by EPR at 600 to 317 ka (Laurent, 1993; Van Vliet Lanoë et al., 1995; Van Vliet Lanoë et al., 1997; Brault et al., 2001; Brault et al., 2003). Farther south in the Loire Valley (Fig. 1), a sole of fluvial deposits has been dated from 60 ka to 13 ka (Barbaroux et al., 1980). Such a large time span reflects a sampling across terraces belonging to different stratigraphic cycles of the Upper Pleistocene. In our proposed model of valley morphogenesis, the thalwegs, which correspond to the last and deepest incision, would not be older than the last glacial sea-level fall, and therefore the fluvial deposits above could be younger than 20 ka. However, one (or more) sequence boundaries are likely to be present within these deposits (Fig. 12).

The truncation surface at the top of unit 1 separates the fluvial deposits from overlying estuarine to marine units. It is thus interpreted as a transgression surface throughout the area, which amalgamates to the sequence boundary at the base of the main valley fill (Fig. 12).

Unit 2 shoreface deposits are mostly resting on the bedrock but locally are seen on unit 1. They are truncated by the wave ravinement surface that correlates to the top of unit 4 (Fig. 12). Therefore, they can be considered as roughly time equivalents to Units 3 and 4. The pure marine setting of Unit 2, compared to the estuarine environments inferred for Units 3 and 4, is then explained by the more seaward location of Unit 2 (Fig. 12).

Units 3 and 4, which mainly rest above Unit 1, are interpreted as deposited in the course of the last sea-level rise over the area. As the fluvial valleys were flooded by the sea, sediment supply could not keep pace with the increase of accommodation space, and this would explain why Unit 3 is mostly aggrading. The very homogeneous structure of Unit 3, and the inferred fine-grained sedimentation, would point to a wave-dominated estuary (Dalrymple et al., 1992). No sandy barrier has been observed in the seismic record, but the basement shoal line could have acted to damp the tidal wave entering the valleys, favoring low-energy sedimentation comparable to that in estuarine central basins (lagoonal basins). Unit 4 is also aggrading, as suggested by the vertical stacking of channel belts (Fig. 11), and it was therefore emplaced during the sea-level rise as well. This unit corresponds to the onset of strong currents in the valleys, causing a grain-size coarsening and channelizing of deposits. The increase in tidal current could be related either to increased constriction of tidal flows inside the valley due to its progressive infilling by Unit 3 or to an increased tidal prism due to marine flooding of valley tributaries. While both hypotheses may work together, the second one is more likely to explain the sharp tidal ravinement at the base of Unit 4. Both Unit 3 and Unit 4 display a general aggrading pattern, while Unit 2 is clearly prograding, but all these deposits express the infill of valleys, and no direct relationship with sequence stratigraphic systems tracts can be drawn from this contrast in architecture. All of them are likely to belong to the same transgressive systems tract.

Unit 5 is interpreted as offshore muds aggrading above the estuarine valley fills. It rests above the wave ravinement surface that truncates all the units below (Fig. 12). The pebble lag is found at a depth of 35 m bpsl. It is dated at about 8100 yr BP, and was therefore emplaced when sea level was about 27 m below present, in a water depth of about 7 m. This is consistent with a shallow fair-weather wave base, which could be explained by the sheltering effect of the basement shoal line (Fig. 12). The wave ravinement surface is found very near to the elevation of valley interfluves. This suggests that the waves shaping this surface were not derived from

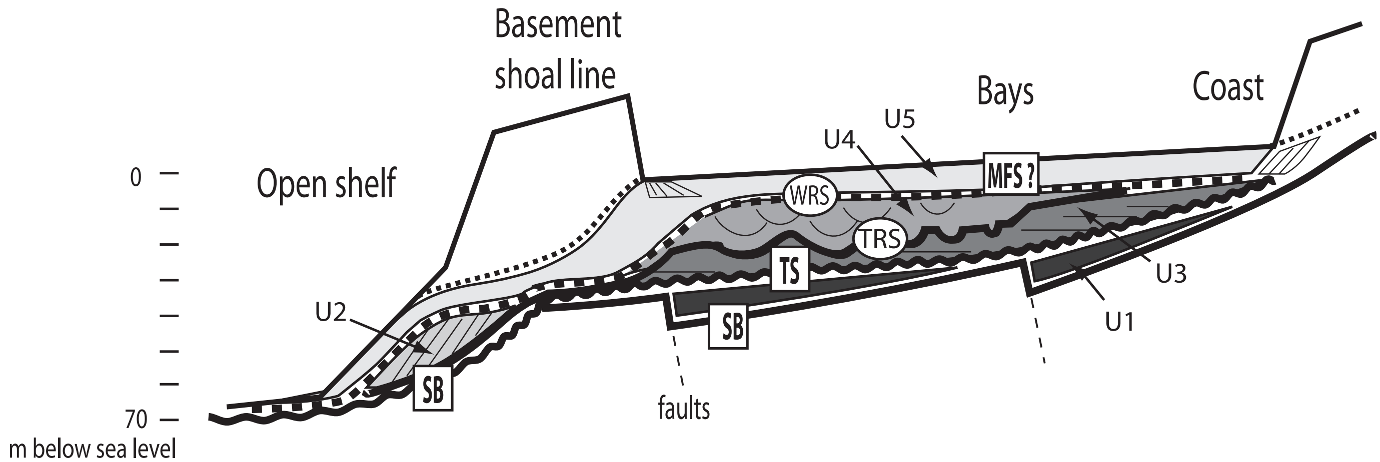


FIG 12.—Southern Brittany incised-valley-fill model. Sequence-stratigraphic surfaces: SB, sequence boundary; TS, transgressive surface; MFS, maximum flooding surface.

oceanic swells propagating inside the valleys (these would have been damped by the islands and the sinuous paths of the valleys), but rather originated inside the basin, located landward from the basement shoal line as soon as the valley interfluves were flooded by the sea in the course of the transgression (so that a sufficient fetch could permit waves to form there). In this respect, it is a *bay ravinement surface*, as defined by Allen and Posamentier (1993) and illustrated in the buried offshore incised valleys of western Brittany (Reynaud et al., 1999). The bedrock pebbles and abundant shell debris concentrated at this surface would have been reworked from the interfluves of the valleys, as these were rapidly swept by the retreating estuarine shorelines. Because large waves never could develop inside this coastal basin, wave ravinement rapidly ceased as water depth became greater than 7 to 10 m (8000 yr BP), allowing deposition of dominantly suspended sediments. Similar upper-valley fills have been found in incised valleys that evolved into larger embayments under transgression (Kindinger et al., 1994; Reynaud et al., 1999; among others). The wave ravinement surface is thus a good time line in this basin, as well as the best candidate for tracing the maximum flooding surface (Fig. 12), although no evidence of renewed progradation is seen in Unit 5. It is worth noting that although the area still experiences large tides, deposition has no longer been tide-dominated after valley interfluves were flooded. The rare bars preserved at the top of Unit 5 could represent the only hints of tidal influence in this offshore system (not taking into account the present-day estuaries, landward of it, which are still tide-dominated).

VALLEY FILL CONTROLLED BY VALLEY MORPHOLOGY

Shelf Slope and Valley Morphology

The interplay of several fault systems across the study area is responsible for the contrast between the morphologies of the valleys in plan view and along their longitudinal profile (Fig. 7). Although these faults might have been reactivated until very recently, they never affect the estuarine to marine valley fills, so that the same tectonic valley morphology as at present already existed as it was infilled by sediments.

The depth, width, and amount of incision of the valleys are influenced by the magnitude of base-level fall as incision proceeded (see Miall, 1991). This magnitude is directly controlled by the difference between the slope of the rivers and the slope of the shelf where rivers are forced to flow at lowstand. The shelf slope dip is about 0.23° in the western part of the study areas, where the valleys are linear, narrow, and deeply incised (Concarneau and Lorient Valleys; Fig. 2). By contrast, the average shelf slope is about 0.075° in the east, where the shallow submarine areas (from the coast to the basement shoal line; Fig. 2) are not steeper than the mainland. As expected, the valleys flowing through this domain are overall wider, dendritic in shape, and less incised (Artimon Valley; Fig. 4).

Size of the Tidal Prism

Valley morphology controls the size of the tidal prism, that is, the volume of water exchanged during one half tidal cycle through the valley section. The size of the tidal prism increases with the tidal range and the extent of intertidal areas inside the estuary. There is no hint that tidal range would have changed under regional oceanographic controls in the course of the last sea-level rise. By contrast, the hypsometry of the mapped coastal area shows that the extent of intertidal surfaces inside the valleys may have changed greatly through transgression (Fig. 4). Therefore, we consider that this is the main parameter driving the tidal prism. In this respect, there is a significant difference between the linear valleys that lack tributaries (small surface, small tidal prism), and the dendritic valley networks (large surface, large tidal prism) (Table 2). Tidal currents are scaled to the ratio of the tidal prism to the entire volume of the incised-valley system. They would increase with an increase of the tidal prism or a decrease in water depth in the flooded valley.

As a consequence, the currents in the linear, long trunk valleys such as Lorient Valley (Figs. 4, 5, 8) would have been much slower than in those with numerous lateral branches that were part of the same estuarine system, such as Artimon Valley (Figs. 4, 8). This could partly explain why narrow and linear valleys are mostly filled up by Unit 3 (muddy, aggrading deposits, indicating slow currents), while large, dendritic ones mostly comprise Unit 4 deposits (tidal channel and bars, indicating faster currents; Figs. 8, 9, 10).

This contrast may apply to the successive stages of infilling of one valley as well, because most of the valleys widen upward in cross section (Fig. 8). The increased channeling in unit 4 would be related to the increase in tidal prism that occurred when the broader parts of the tributary valleys were flooded (Fig. 8). When sea level was lower and only the narrow parts of the valleys cutting through the basement shoal line were affected by tides, tidal currents would have been slower because the tidal prism was smaller. This is supported by the fact that Unit 4 is best expressed in the valleys where the valley floor is incised not much deeper than 40 m (Etel and Vilaine Valleys; Figs. 4, 7, 8).

Friction and Constriction of Currents

Another reason for finding Unit 3 deposits mostly in narrow, linear valleys is that these valleys are isolated from the open sea by one or several narrow passes (Lorient and Vilaine Valleys; Figs. 4, 8). In such valleys, tidal flows are likely to be damped by friction on the valley walls and the tidal wave energy dissipated by diffraction through the tidal passes (a "hyposynchronous" estuary). Friction would have been increased even more by the longitudinal, locally zigzag pattern of these valleys (Etel and Vilaine; Figs. 4, 6). On the other hand, flow constriction could have been enhanced in the entrances of the narrow valleys, such as the Lorient and Vilaine valleys, bringing about increased tidal scour that would have removed all older deposits (Fig. 8). However, this effect would not have extended very far inward in those valleys, as opposed to friction and dissipation of the tidal wave. By contrast, where the valley entrances are broader, deposition of sediments supplied by tides at the mouth of the estuary could have been preserved. This is the way one can interpret the coarse-grained, prograding shorefaces of Unit 2 (Artimon valley; Fig. 8).

Magnitude of Incision

Most of the thalweg profiles of the valleys have a concave-up shape in their seaward part (Fig. 7), some ending landward at a knickpoint, from where the thalweg is almost flat farther landward. The actual knickpoint is articulated around the basement shoal line (Figs. 7, 12), where the magnitude of incision is maximum, and not in the landward part, as expected if the knickpoint were around the modern coast (see Miall, 1991; Schumm, 1993; Talling, 1998; Thomas and Anderson, 1994). As a consequence, there is more accommodation space in the seaward part of the incised valleys than in the general case. Also, the amount of accommodation space throughout the valleys is proportional to their length. At the time of valley incision, the basement shoal line acted as a barrier against the streams that had to find an oblique way between the islands to reach the outer shelf, so that the length of the valleys is much longer than it should be, taking into consideration only the regional slope.

Therefore, there is a net excess of accommodation at the end of the stratigraphic cycle, in contrast to the most common case (see Heap and Nichol, 1997). This would explain why estuarine deposits (Units 3–4) were not able to fill up the entire valley (Fig. 8). The amount of preserved estuarine deposition is scaled to the magnitude of valley incision but cannot keep pace with it. Therefore, Unit 5 marine deposits are thicker where the valley is more incised (Table 2, Figs. 7, 8). Also, the depressed areas of the sea floor indicate the location of the most incised parts of the valleys (underfilled segments of the valleys).

Preservation of Deposits

The transgressive deposits around and seaward from the basement shoal line are prograding (Unit 2), whereas they are aggrading landward of it (Units 3–4). The basement shoal line is thus also responsible for a major dynamic change, which is compared to that of a mobile sandy barrier in a wave-dominated estuary (Dalrymple et al., 1992). The similarity with a wave-dominated estuary is that the sills between the islands act as tidal passes, through which tidal currents are forced (Figs. 2, 4). The bedrock "barrier" is responsible for breaking wave energy propagating from the open ocean to the coast. However, the basement shoal line does not follow a continuous retreat of the shoreline throughout the sea-level rise, in contrast to the model of wave-dominated estuaries with mobile, sandy barriers (see Nichol et al., 1994). Therefore, the preservation potential of the sediments deposited behind this line is much higher than it would be in the wave-dominated estuarine model. The presence of wave erosion at the base of Unit 5 does not prevent offshore muds to aggrade over tens of meters above, a situation that is relatively uncommon in offshore settings at such latitudes. The bedrock high continues to act as a barrier, protecting the area behind from the full intensity of the waves.

CONCLUSIONS

The valleys incised along the southern Armorican coast were formed through several incision cycles, likely related to the last few 100 kyr eustatic cycles of the Quaternary, with the interplay of neotectonic movements (faults, uplift). Above a patchy of basal unit of older fluvial deposits, the valley filled during the last postglacial sea-level rise under the dominant action of tides. Most of the valley fill corresponds to the transgressive to early highstand depositional systems tracts. The main deposits are aggrading estuarine muds (Unit 3) and tidal-channel belts (Unit 4). Shoreface bars (Unit 2) may be preserved at the seaward end of one of the valleys. The estuarine deposits are capped by offshore muds (Unit 5). The most prominent boundaries within the valleys are the tidal ravinement surface at the base of the tidal channel belts and the wave ravinement surface at the base of the offshore muds.

This study illustrates the genetic linkage between valley morphology and the architecture of the valley fill. The shape of the south Armorican valleys is very complex, with each valley having a different shape, because of a strong local tectonic control on the river paths and depth of incision. The differences observed with the Zaitlin et al. (1994) incised-valley model are interpreted as caused mainly by the variations in the magnitude of incision that are forced by topographic heterogeneities, namely the presence of a submerged basement shoal line up to 30 km off the modern coast. The most important result is that linear, deep and narrow valleys are prone to fill mostly with aggrading estuarine muds (Unit 3), while tidal-channel belts (Unit 4) are best expressed in dendritic, shallow and wide valleys. The difference in current speed that produced these differences is proportional to the surface area of the related estuaries (i.e., the tidal prism). Tidal modeling would be the next step toward the assessment of the control by tides on the infilling of such incised valleys.

Due to the anomalously high accommodation space created offshore by incision across the irregular shelf topography, lowstand and highstand fluvial deposit are not volumetrically important in these valleys, and estuarine deposits were not able to fill the accommodation space completely. Therefore, the volumetrically most important component of the valley fills is offshore mud. However, the main reason there is marine mud here is the sheltering provided by the offshore bedrock shoal complex, which prevents dispersion of the mud by marine processes.

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