



Deep structure of the Armorican Basin (Bay of Biscay): a review of Norgasis seismic reflection and refraction data

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1 Deep structure of the Armorican Basin (Bay of Biscay): a review 2 of Norgasis seismic reflection and refraction data

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12 13 **Abstract**

14 The Bay of Biscay is bounded to the north by the North Biscay margin, which comprises the
15 Western Approaches and Armorican segments. In the 1970s and 1980s, most researchers
16 considered this margin typical of a non-volcanic passive margin: it is characterized by a striking
17 succession of tilted blocks beneath which occurs the S reflector and the continent–ocean boundary
18 is abrupt. This paper examines the Armorican segment and is based on a study of all early seismic
19 profiles together with new multichannel reflection and refraction seismic data (Norgasis cruise). An
20 important result is the discovery of a 80 km wide ocean–continent transition zone that coincides with
21 the Armorican Basin (a deep sedimentary basin). It is characterized by a high-velocity lower-crustal
22 layer (7.4–7.5 km s⁻¹) overlain by sediments. The other results are: (1) the main crustal thinning
23 occurs exclusively under the narrow continental slope; (2) the tilted blocks and the S reflector are
24 observed only at the base of the continental slope in the narrow domain called the ‘neck area’; (3)
25 the North Biscay Ridge is a large oceanic plateau present only off the NW Armorican margin rather
26 than a long ridge elongated off the whole North Biscay margin.

27 **Keywords:** Bay of Biscay, passive margins, transition zones, rifting, crustal thinning.

28 **Introduction**

29 From the West Iberia margin to the Goban Spur margin, the zone of transition between thinned
30 continental crust and typical oceanic crust was first proposed as a sharp transition, less than 10 km
31 wide (e.g. Montadert *et al.* 1979a; Avedik *et al.* 1982; Derégnaucourt & Boillot 1982; de Graciansky
32 *et al.* 1985; Ginzburg *et al.* 1985; Boillot *et al.* 1987a; Whitmarsh *et al.* 1990; Pinheiro *et al.* 1992;
33 Sibuet *et al.* 1992). More recently, off the West Iberia non-volcanic margin, seismic profiles and
34 deep boreholes have indicated that this contact may involve a wider (30–120 km) zone referred to

35 as the ocean–continent transition (e.g. Pickup *et al.* 1996; Whitmarsh & Sawyer 1996) or the zone of
36 exhumed continental mantle (Whitmarsh *et al.* 2001). The ocean–continent transition zone coincides
37 with a very quiet magnetic zone (Whitmarsh *et al.* 1990) whose magnetic anomaly amplitudes are
38 much lower than those of the well-known Cretaceous Magnetic Quiet Period.

39 Reflection profiles show that the upper basement rocks are very thin, with typically low velocities
40 (between 4.5 and 6.0 km s⁻¹), and are underlain by a lower basement layer characterized by high
41 velocities of 7.2–7.6 km s⁻¹ and by a complex reflectivity including both landward- and seaward-
42 dipping seismic reflectors (Pickup *et al.* 1996). This zone has a lack of Moho reflections (e.g. Chian
43 *et al.* 1995, 1999, Louden & Chian 1999). The deepest basement layer has seismic velocities
44 between 7.2 and 7.6 km s⁻¹ (e.g. Whitmarsh *et al.* 1990; Chian & Louden 1994; Chian *et al.* 1995,
45 1999), which are lower than that measured within normal mantle (8 km s⁻¹), and higher than that
46 measured within the oceanic layer 3 or within lower continental crust (6.5–7 km s⁻¹). This is a
47 common feature of this and other ocean–continent transition zones and is referred to as the high-
48 velocity lower-crustal layer (Louden & Chian 1999). Currently, ocean–continent transition zones are
49 interpreted variously as one of the following: (1) thinned continental crust intruded by melt from the
50 mantle (Whitmarsh & Miles 1995), which may represent transitional continental crust with more
51 basic composition; (2) thin, tectonized oceanic crust produced by ultraslow sea-floor spreading
52 (Srivastava & Roest 1995; Srivastava *et al.* 2000); (3) tectonized underplated gabbros, as previously
53 suggested for the Newfoundland Basin and Flemish Cap margins (Keen & De Voogd 1988); (4)
54 exhumed and variably serpentinized upper mantle, as suggested for the West Iberia margin (Boillot
55 *et al.* 1980, 1987b; Girardeau *et al.* 1988; Sawyer *et al.* 1994; Whitmarsh *et al.* 1998; Chian *et al.*
56 1999; Whitmarsh *et al.* 2001). This last interpretation is also valid for the Labrador and western
57 Greenland margins (Chian & Louden 1994; Chian *et al.* 1995; Chalmers 1997), the Newfoundland
58 Basin (Reid 1994), and the SW Australia margin (Nicholls *et al.* 1981). This last hypothesis has
59 been supported by geological data obtained by dredging, drilling and using submersible off the West
60 Iberia peninsula (e.g. Boillot *et al.* 1987b, 1988; Beslier *et al.* 1988; Girardeau *et al.* 1988; Sawyer *et al.*
61 1994; Whitmarsh *et al.* 1998) and by dredging off the Australian margin (Nicholls *et al.* 1981).

62 From these examples, there is general agreement for an ocean–continent transition zone formed
63 during prolonged periods of extension with little or no melt generated from the upper mantle.
64 Although the 7.2–7.6 km s⁻¹ velocities are characteristic of the ocean–continent transition zones of
65 some non-volcanic rifted margins, they are also observed beneath volcanic passive margins, where
66 they would represent underplating gabbros or a lower continental crust intruded by mafic intrusions
67 (Skogseid *et al.* 1992; White 1992a). One of the differences between volcanic and non-volcanic
68 margins presented by White (1992a, 1992b) concerns the adjacent oceanic crust, which is
69 considerably thicker than normal off a volcanic margin and is thinner than normal off a non-volcanic
70 margin.

71 This paper examines the Armorican segment of the North Biscay margin from the continental shelf
72 to the oceanic crust. The principal aims of the study reported here were to identify more accurately
73 the boundaries between oceanic and continental crusts, to determine the location of the zone of
74 crustal thinning and to determine the crustal structure of the whole margin. Such data and geometric
75 constraints are necessary for modelling the processes of continental break-up and of crustal
76 thinning, the initiation of oceanic accretion, and kinematic reconstructions. The main result of this
77 study is the discovery of a wide transitional domain between the typical continental and oceanic
78 domains.

79 **Regional setting**

80 The Bay of Biscay (see Fig. 1) is a triangular oceanic domain bounded by two conjugate margins,
81 the North Biscay margin and the North Iberia margin. The North Biscay margin comprises two
82 segments: the Western Approaches margin (oriented N110°) and the Armorican margin, with
83 northern (N110°) and southern (N140°) components. A steep linear continental slope dominates the
84 morphology of the North Biscay margin, disrupted only by the Meriadzek Terrace promontory in the
85 Western Approaches domain collinear with the NE–SW axis of the English Channel Rift. The SE
86 boundary of the Western Approaches margin with the Armorican margin is near the Black Mud
87 Canyon and linked by the abrupt eastern termination of Meriadzek Terrace. At the base of the
88 continental slope of the Armorican margin, we distinguish a deep basin, the Armorican Basin, limited
89 to the west by the Trevelyan–Meriadzek complex and to the east by the Dôme Gascogne structure.

90 The relative movements between the North American, European and Iberian plates (e.g. Olivet
91 1996) during Early Cretaceous time induced the formation of the Bay of Biscay, contemporaneously
92 with the opening of the North Atlantic Ocean. The absence of magnetic Chron M0 in the Bay of
93 Biscay and the age of the synrift sediments drilled on the Western Approaches margin (Montadert *et*
94 *al.* 1979b) mark the beginning of rifting during the Early Cretaceous (140–110 Ma, Neocomian to
95 Late Aptian) and the beginning of oceanic accretion in Late Aptian–earliest Albian time. Pre-rift
96 sediments consist of Jurassic (150–140 Ma, Kimmeridgian to Portlandian) platform carbonates. As
97 Chron 33 (80 Ma, Campanian) has not been recognized, the oceanic accretion of the Bay of Biscay
98 ceased after Chron 34 (Fig. 2; Williams 1975). The later post-rift structural evolution of the Bay of
99 Biscay is strongly linked to the Pyrenean orogenic phases induced by the Campanian–Oligocene
100 (80–35 Ma) convergence of the Iberia peninsula towards Europe. This convergent movement also
101 led to the partial closing of the Bay of Biscay and major deformation of the North Iberian margin.
102 The Trevelyan Seamount (see Fig. 1) and the Dôme Gascogne are large structural inversions
103 formed during the Pyrenean compressive phase (Debyser *et al.* 1971; Frappa & Vaillant 1972;
104 Thinon *et al.* 2001). Recent studies of the Pyrenean phase emphasize a particular characteristic: the
105 Armorican Basin is very weakly affected by the Pyrenean compressive deformation phase, which is

106 intense only at its oceanic and continental domain boundaries (fig. 5 of Thinon *et al.* 2001). Rift
107 structures are today preserved on the North Biscay margin and in the Parentis Basin (see Fig. 1).

108 **Background**

109 The Bay of Biscay was surveyed extensively in the 1970s and 1980s (Debyser *et al.* 1971;
110 Montadert *et al.* 1971, 1974, 1979; Derégnaucourt & Boillot 1982; Barbier *et al.* 1986; Le Pichon &
111 Barbier 1987). Oceanographic surveys began again in 1994 and 1997 (e.g. the Norgasis cruise;
112 Avedik *et al.* 1996; Thinon 1999).

113 The North Biscay margin is usually considered to be a typical example of a non-volcanic passive
114 margin (e.g. Montadert *et al.* 1974, 1979b; de Charpal *et al.* 1978; Avedik *et al.* 1982;
115 Derégnaucourt & Boillot 1982; Ginzburg *et al.* 1985; Barbier *et al.* 1986; Whitmarsh *et al.* 1986; Le
116 Pichon & Barbier 1987). From the few scattered seismic profiles shot off the Western Approaches
117 margin across the Meriadzek–Trevelyan complex (Montadert *et al.* 1971), de Charpal *et al.* (1978)
118 first observed a strong reflector below the tilted blocks, which they called the S reflector. Following
119 these observations, most workers (e.g. Avedik & Howard 1979; Montadert *et al.* 1979b) have shown
120 that the Western Approaches margin from the shelf to the true oceanic crust is characterized by a
121 striking succession of tilted fault blocks beneath which occurs the S reflector (Fig. 3a). The structural
122 pattern of the Armorican margin is controversial. First, on the basis of subsidence history, the
123 margin is characterized by significant crustal necking under the slope and by the lack of tilted blocks
124 (Fig. 3b; Le Pichon & Sibuet 1981). More recently, using the SNEAp seismic reflection dataset,
125 Barbier *et al.* (1986) applied the global structural pattern of Montadert *et al.* (1979b) to the
126 Armorican margin, and proposed that the Armorican margin was characterized, from the continental
127 slope to the oceanic domain, by a succession of tilted blocks over the S reflector (Fig. 3c). From our
128 new seismic dataset, we verify this crustal pattern.

129 The Armorican Basin is a thick sedimentary basin (5–7 km thick), discovered by Bacon *et al.* (1969).
130 Resting on the basement, Unit 3B, a seismically transparent unit characterized by velocities of 4.4
131 km s⁻¹ (Bacon *et al.* 1969) to 4.6 km s⁻¹ (Avedik & Howard 1979), is present (Fig. 4). Based on its
132 morphology and acoustic facies, Unit 3B was interpreted as: (1) evaporites (Sibuet *et al.* 1971; Grau
133 *et al.* 1973; Montadert *et al.* 1974; Derégnaucourt & Boillot 1982); (2) basaltic rocks (Montadert *et al.*
134 1971; Grau *et al.* 1973); (3) pre-rift Cretaceous sediments (Barbier *et al.* 1986); (4) equivalent to the
135 synrift formation of the Western Approaches margin (Montadert 1984). More recently, Unit 3B has
136 been interpreted as a sedimentary body emplaced by slumping at the end of rifting phase (Thinon *et al.*
137 *et al.* 2002). This unit rests on a subhorizontal layer with high-amplitude and continuous reflections
138 attributed to the top of the basement (Montadert *et al.* 1971, 1974; Sibuet *et al.* 1971; Grau *et al.*
139 1973) or to the S reflector (Barbier *et al.* 1986). Rather than being a typical oceanic crust (Limond *et al.*
140 *et al.* 1974), the substratum of the Armorican Basin is thought to consist of an extremely thinned (<4
141 km) continental crust (Fig. 3b and c; Grau *et al.* 1973; Avedik & Howard 1979; Montadert *et al.*

142 1979b; Le Pichon & Sibuet 1981; Derégnaucourt & Boillot 1982; Barbier *et al.* 1986). Uncertainties
143 about the nature and the age of Unit 3B have led to controversy about the age of the Armorican
144 Basin, the nature of its substratum and its origin. Different hypotheses postulated that the basin
145 was: (1) created along a strike-slip fault system of the Hercynian Orogeny (Ziegler 1982); (2) the
146 result of a late Triassic extensional phase (like the English Channel Rift and the Parentis Basin (Fig.
147 1) (Derégnaucourt & Boillot 1982; Olivet 1996); (3) formed during the Early Cretaceous rifting of the
148 Bay of Biscay (Debyser *et al.* 1971; de Charpal *et al.* 1978; Le Pichon & Sibuet 1981). Using new
149 seismic refraction and reflection data, we here describe the deep structure of the Armorican Basin
150 so as to understand its origin.

151 The northern continent–ocean boundary of the Bay of Biscay was previously assumed to be sharp,
152 less than 10 km in width (de Charpal *et al.* 1978; Avedik *et al.* 1982; Derégnaucourt & Boillot 1982;
153 Ginzburg *et al.* 1985; Whitmarsh *et al.* 1986). Most workers (e.g. Bacon *et al.* 1969; Debyser *et al.*
154 1971; Grau *et al.* 1973; Montadert *et al.* 1974, 1979b; Le Pichon & Sibuet 1981) have postulated
155 that this boundary is associated with a basement high called the North Biscay Ridge, which is
156 crossed only by the OC17 seismic profile (Figs 3b and 4). On the evidence of the continuity of a
157 strong, linear and negative magnetic anomaly observed on the total magnetic field map (Le Borgne
158 & Le Mouël 1970; Fig. 2a), this basement high was inferred to extend along the entire North Biscay
159 margin (see Fig. 2a; Montadert *et al.* 1979b; Derégnaucourt & Boillot 1982). In this paper, we will
160 show using our new seismic data and processed magnetic data that the North Biscay Ridge is
161 unlikely to extend along the entire North Biscay margin. We will also discuss the occurrence in the
162 Bay of Biscay of a wide ocean–continent transition zone.

163 **Seismic data acquisition and processing**

164 The structural analysis presented in this paper is mainly based on a reinterpretation of all the
165 seismic profiles acquired between 1969 and 1981 (Fig. 1; e.g. Debyser *et al.* 1971; Montadert *et al.*
166 1979b; Vaillant 1988). They include a rectangular net of 6500 km of industry acquired and
167 processed stacked multichannel seismic reflection profiles (with a 24-fold recording system) of the
168 Société Nationale ELF (Barbier *et al.* 1986). These last lines (SNEAp), which we have reinterpreted,
169 provide a shallow to deep crustal seismic image (10–12 s TWTT (two-way travel time) recording) of
170 the North Biscay margin and of the Armorican Basin. Unfortunately, the SNEAp profiles off the
171 Armorican margin do not reach the oceanic domain. These data are completed by new multichannel
172 seismic reflection and refraction data collected during the Norgasis cruise.

173 The Norgasis seismic reflection profiles were acquired with a 96-channel streamer (25 m interval)
174 using a single bubble array with 8–10 air guns and a generator capacity from 804 to 1230 in³
175 (Avedik *et al.* 1993, 1996). Twenty-four-fold coverage was available with a 50 m shot interval. The
176 17 s TWTT recording provides deep crustal seismic images of the Armorican margin including the
177 continental shelf, slope and the true oceanic areas. The Norgasis seismic refraction data (Fig. 1)

178 were acquired using an array of 20 ocean-bottom seismometers developed at Hokkaido University
179 by the LOBS Laboratory (Kanasawa & Shiobara 1994). The instruments recorded continuously,
180 providing a large set of long and short profiles together with many off-line sections. The seismic data
181 were digitized at a 100 Hz sampling rate. When necessary, the instrument location was corrected by
182 matching the observed and computed primary and first multiple water-wave arrivals. The initial
183 crustal structure of the in-line profiles was obtained using the joint inversion and ray-tracing
184 algorithm of Zelt & Smith (1992). The main horizons identified in the sedimentary cover on the
185 vertical incidence profiles were found to correspond to significant wide-angle reflectors and
186 refractors so that their location measured in TWTT was kept fixed during the inversion. Only the
187 layer velocities were allowed to vary. This restriction applied also to the MS and S horizons (see
188 later). The deeper layers are better constrained by the ocean-bottom seismometer data and their
189 positions were later compared with the vertical incidence data. The thickness, velocities and
190 gradients of the different layers were refined by the use of synthetic seismograms computed by the
191 method of Zelt & Ellis (1988), which is based on asymptotic ray theory. The seismic refraction
192 model, presented for the Norgasis 14 profile (see Fig. 7c, below) and discussed in the text, was built
193 mainly from record sections obtained from seven ocean-bottom seismometers. However, because
194 of the spacing of instruments (about 20 km), many features of the velocity model are clearest on the
195 strike profiles and the parallel profile near seismic line OC17 (see Fig. 6, below). In fact, the
196 Norgasis 14 model represents a synthesis of the interpretation of the whole dataset.

197 The details of the analysis of the seismic refraction data will be presented in a forthcoming paper but
198 some of the most relevant features are presented in Figure 5. The ray tracing that matches the
199 interpreted travel time curves (Fig. 5a) shows that the deep structure is well constrained in two
200 critical areas of the model. Travel time residuals are less than 0.1 s, the estimated picking
201 uncertainty. The layer densities were estimated from the P-wave velocity model using the curves
202 obtained by Ludwig *et al.* (1970) and a 2D gravity model was obtained (Fig. 5b). Near the coast the
203 seismic refraction model had to be slightly modified in the unconstrained part to obtain a good fit.
204 The density of the high-velocity lower-crustal layer was also reduced, corresponding to a difference
205 of 0.2 km s^{-1} in its original velocity (from 7.2 to 7.4 km s^{-1}). The r.m.s. misfit between observed and
206 computed free-air anomalies is 3.0 mGal . The velocity and thickness of the high-velocity lower-
207 crustal layer are constrained by refracted arrivals (phase PH) and reflected arrivals from its top
208 (phase PHP) and its base (phase PMP) (see Fig. 5d and e). The high velocity contrast between Unit
209 3B (5.2 km s^{-1}) and the high-velocity lower-crustal layer (7.4 km s^{-1}) produces high-amplitude
210 reflections, as illustrated by the observations and modelled by synthetic seismograms (Fig. 5e and
211 f). Refracted arrivals from the upper mantle (phase Pn) are also observed in some recordings (Fig.
212 5d). The structure of the oceanic crust is deduced from refracted arrivals from layer 2 (P2) and layer
213 3 (P3) (see Fig. 5c). Here the presence of the high-velocity lower-crustal layer is inferred by the
214 identification of both PHP and PMP arrivals that are conspicuous on several other record sections.

215 More recently, we made a multibeam survey of this zone during the Zee-Gascogne cruise in 1997
216 on board the R.V. *Atalante* (Fig. 1). During this cruise, seismic reflection profiles were acquired at
217 10 knots with a six-channel streamer, two Generator injector guns (105 in² each), and with 10 s
218 TWTT recording. A conventional seismic processing scheme (stack, migration) using the ProMax
219 software was applied to the Norgasis and Zee-Gascogne reflection data (Thinon 1999). This
220 complete and homogeneous dataset (18 000 km) provides new information on the deep structure
221 across this non-volcanic passive margin.

222 **Interpretation**

223 From our data, we have divided the Armorican margin into five domains (Fig. 6) from the continental
224 domains (I, shelf; II, slope) to the oceanic domain (V). They are separated by a transitional domain
225 (IV), which does not present any characteristics of the others and which passes to the slope (II)
226 through a zone we call the 'neck area' (III). The Norgasis14 profile extends from domain I to domain
227 V (Fig. 7).

228 ***The Armorican continental margin (domains I, II and III)***

229 The continental crust of the shelf comprises a non-reflective crust lying on a layered lower crust
230 (around 3 s TWTT thick) that displays horizontal, high-amplitude and low-frequency reflections (Fig.
231 8). At its base (at 10–11 s TWTT depth), there is a three-phase reflection with very high amplitude
232 and low frequency. We interpret it as the Moho reflection, following a comparison with other seismic
233 profiles across the continental shelf (Cazes *et al.* 1988; Dymant 1989). The continental crust is
234 about 30–32 km thick with velocities of 5.8–6.1 km s⁻¹ for the upper crust and 6.6–7.0 km s⁻¹ for the
235 lower crust (Avedik & Howard 1979).

236 The continental slope exhibits a variable morphology (Fig. 1). Its northern segment constitutes a
237 simple and steep slope (30 km wide, with a dip of 7°). The South Armorican slope is steep and
238 narrower in its lower part (15 km wide with an average dip of 7°), and wider and weakly dipping
239 (<4°) in its upper part. The seismic profiles show that the steep slope coincides with a major
240 escarpment, which is exposed or covered by thin sediments (Figs 8 and 9). This escarpment
241 constitutes the oceanward flank of a horst, which either bounds a hanging sedimentary basin in the
242 upper part of the slope (Fig. 8) or marks the edge of the shelf (Fig. 9). Beneath the continental
243 slope, no conspicuous tilted faulted blocks have been observed. Under the slope, the reflectivity of
244 the lower continental crust is strongly attenuated on all seismic lines (Fig. 8). This could be an
245 inherent feature of the acquisition and processing of seismic profiles in steeply dipping sea-floor
246 areas. It could also reflect the rifting process (tectonic and thermal events and consequences of
247 intra-crustal fluid circulation).

248 Faulted and tilted continental blocks are restricted to a deeper 30 km wide domain, at the base of
249 the continental slope, and corresponding to the 'neck area' (zone III, Fig. 6). This is the last domain

250 oceanward in which remnants of the continental crust have been observed. This domain contains
251 several horsts, which can create relief at the foot of the slope (Figs 1, 4 and 7a), as well as a few
252 tilted blocks (two successive blocks at most, Figs 10 and 11) and some shapeless faulted blocks
253 (Figs 8 and 9). These blocks are uncommon, relatively small in size, 2 s TWTT thick at most, and
254 less than 20 km wide. The tilted blocks include a very important pre-rift unit (1–2 s TWTT thick) lying
255 on a portion of the crust (Figs 9 and 10). The refraction model along the Norgasis 14 profile (Fig. 7b)
256 exhibits a horst with velocities from 5.4 to 6.2 km s⁻¹ that confirms the presence of upper continental
257 crust blocks in the 'neck area'. The blocks are tilted on a strong, continuous seismic horizon, which
258 is identified as the S reflector (Figs 7a and 10) and coincides with a sharp seismic velocity contrast
259 (6.2–6.8 km s⁻¹; Fig. 7b). We emphasize that on all seismic lines no direct seismic relation has been
260 observed between the S reflector and any structure or reflector within the continental slope. At the
261 foot of the continental slope, the Norgasis reflection data show a deep-layered unit (Figs 7, 8 and
262 11). The top of this deep-layered unit coincides with the S reflector. Its base is marked by a strong
263 reflection that deepens landward, evolving from a single reflection at 9 s TWTT to a triple reflection
264 at 10 s TWTT (Figs 8 and 11). The internal reflections characterizing this unit are truncated by or
265 wedged out above its base, which we have attributed to the crust–mantle boundary (Fig. 11). This
266 deep-layered unit is thus sandwiched between the S reflector and the shallow top of the mantle. The
267 vertical velocity distribution (Fig. 7b) shows that this deep-layered unit has velocities ranging from
268 6.8 to 6.9 km s⁻¹, similar to those of the lower continental crust. The unit vanishes toward the ocean
269 and seems to also have no direct seismic relation with the layered lower continental crust observed
270 under the shelf. The nature of this deep-layered unit remains open: it could be thinned lower
271 continental crust of the shelf, a new layered continental crust, or other material.

272 ***Transitional domain (domain IV)***

273 Along the Armorican margin, the continental and oceanic domains are separated by an 80 km wide
274 transitional domain (zone IV, Fig. 6), associated with a magnetic signature characterized by low-
275 amplitude magnetic anomalies, without linearity, similar to a very quiet magnetic zone (Fig. 2b). Its
276 eastern part exhibits some discrete, strong and positive magnetic anomalies. The acoustic facies of
277 its substratum has neither the characteristics of the continental crust nor those of the oceanic crust.

278 *Substratum of transitional domain.* Throughout the transitional domain, the top of the substratum is
279 an acoustic reflector, the MS reflector that gives rise to reflections with high amplitudes and low
280 frequencies that were in the past considered to represent the S reflector. The MS reflector
281 systematically truncates the dipping reflectors (DR) in the underlying material (Figs 7, 8 and 12). It is
282 situated at depths of 8–9.5 s TWTT (Figs 7 and 12). Its regional topography is characterized by
283 large wavelength bulges (Fig. 13), generally independent of the overlying structures. This indicates
284 that the bulges formed just before deposition of the first post-rift sediments (Figs 4, 7 and 12).

285 The MS reflection is generally continuous and smooth (Fig. 12a), except in the eastern landward
286 zone, under Unit 3C, where it is discontinuous with a low amplitude or is suggested to exist
287 indirectly by the upward terminations of the dipping underlying reflectors (DR) (Fig. 12b). On
288 Norgasis data, acquired in the western part, the MS reflector coincides with a clear seismic contrast
289 ($5.0\text{--}5.2\text{ km s}^{-1}$ v. 7.4 km s^{-1}), at the top of a 3–4 km thick layer with velocities of $7.4\text{--}7.5\text{ km s}^{-1}$ (Fig.
290 7b) called the high-velocity lower-crustal layer by Louden & Chian (1999). The base of the high-
291 velocity lower-crustal corresponds to a deeper and discontinuous reflector we call the Mn reflector. It
292 coincides with a velocity contrast (7.5 km s^{-1} v. 8.0 km s^{-1}) underlying the Moho. The high-velocity
293 lower-crustal is associated with the upper part of the substratum containing the DR reflections. The
294 DR reflections are high-amplitude, low–medium-frequency, dipping seismic reflections, interrupted
295 by the MS reflection. Their distribution and their orientation, which includes constant dips, seem to
296 be a function of the topography of the substratum, as they are more important on the side of the
297 bulges (Figs 7b and 13).

298 *The overlying materials.* The landward transitional domain is characterized by Units 3B and 3C
299 overlying the MS reflector (Figs 6 and 12). Unit 3B, observed in the western Armorican transitional
300 zone, is a seismic unit with low reflectivity (Figs 7 and 12a). It has velocities from 4.6 to 5.3 km s^{-1} ,
301 with a positive velocity gradient, and a thickness of 0–3300 m (Fig. 7b). Its top and bottom (MS
302 reflector) are continuous and smooth seismic horizons. On the seismic reflection profiles, Unit 3B
303 appears generally as a half-lens with a horizontal base (Fig. 12a), or a thick ‘incompetent’ body,
304 which pinches out towards the ocean (Fig. 9). From the seismic correlations, Thinon *et al.* (2002)
305 showed that Unit 3B is a sedimentary body emplaced by slumping at the end of the rifting phase,
306 because of its emplacement between the break-up unconformity and the synrift formation of the
307 Western Approaches margin. Unit 3C is observed exclusively in the landward area of the eastern
308 Armorican transitional domain (Fig. 6). It is bounded at its top by a strongly refractive and irregular
309 reflection (Fig. 12b) and at its base by the MS reflection, which is here discontinuous and of low
310 amplitude. On the Loire Maritime 2 profile (Fig. 12c), Unit 3B seems to onlap 3C. Our data do not
311 permit us to identify the nature of Unit 3C.

312 Along the oceanic domain, in the oceanward transitional zone, the MS reflector is generally covered
313 directly by a set of conformable seismic reflectors interpreted as Aptian–Cenomanian post-rift
314 sediments (e.g. Derégnaucourt & Boillot 1982; Thinon 1999). It is locally interrupted by a few
315 important bodies (2 s TWTT thick) that present conical to ridge shapes with flat tops that may
316 represent a pre-Pyrenean erosion surface (Figs 6 and 14). The conical structures may be volcanoes
317 and the basement ridges may be volcanic ridges, consistent with the presence of some discrete,
318 high and positive magnetic anomalies (Fig. 2). Alternatively, they could be peridotite ridges as seen
319 off the West Iberia margin (e.g. Boillot *et al.* 1980, 1987b, 1988; Beslier *et al.* 1988; Girardeau *et al.*
320 1988; Sawyer *et al.* 1994; Whitmarsh *et al.* 1998).

321 Other structures, called ST, are also observed on a disrupted MS reflector (Fig. 7b) around the
322 conical and ridge structures described above. They have a 'chaotic' acoustic facies and are 0.2 s
323 TWTT thick. The top is only suggested by the interruption of the reflectors belonging to the overlying
324 sedimentary unit (Unit 3). From acoustic facies, this feature may be interpreted as a volcano-
325 sedimentary series or volcanic extrusive rocks. Its velocity is too high (6 km s⁻¹, Fig. 7b) for
326 uppermost oceanic crust (White 1992b).

327 *Boundaries of the transitional domain.* The geometric relations between the continental and the
328 transitional domains are well imaged on the Norgasis lines (Fig. 11). The oceanward continental
329 block of the 'neck area' corresponds approximately to the junction of the S, M and MS reflectors
330 (Figs 7, 8 and 11). Figure 11c shows that the MS reflector lies on the prolongation of the junction
331 between the M reflector (continental Moho) and S reflector (base of tilted blocks and top of the
332 deep-layered unit).

333 The relationship between the transitional domain of the Western Armorican Basin and the true
334 oceanic domain is atypical, as locally the high-velocity lower-crustal penetrates the oceanic zone
335 (Fig. 7). Here, it underplates a large basement high (8 s TWTT deep, 60 km wide and 100 km long)
336 with seismic velocities in the range of 4.5–7.0 km s⁻¹ that we interpret as oceanic crust. This
337 basement high would coincide with the North Biscay Ridge defined in the 1970s (see the
338 Background section). Contrary to the initial hypothesis, our new seismic data show that this
339 basement high forms a large plateau, only off the north segment of the Armorican margin (Fig. 6).
340 Compared with the whole oceanic domain, this basement high has a reduced sedimentary cover
341 (Fig. 7), with no Aptian–Cenomanian sediments (Unit 3). This observation suggests that this relief
342 was initiated before the Paleocene to Oligocene Pyrenean compression, certainly during the early
343 spreading phase. This local oceanic structure can be compared with that described on the Vöring
344 margin where underplated low-density materials have induced the uplift of the oceanic crust and
345 significant erosion of its sedimentary cover (Skogseid *et al.* 1992). The underplating of the high-
346 velocity lower-crustal could have produced a relative and local uplift of the observed oceanic
347 plateau.

348 ***True oceanic domain (domain V)***

349 Apart from the large oceanic plateau described above, the acoustic basement of the oceanic
350 domain is 8.5–9 s TWTT deep, shallower than that of the ocean–continent transition zone. It is
351 strongly diffractive and irregular, similar to that of an oceanic crust. The vertical distribution of the
352 seismic velocities is also representative of typical oceanic crust with a 4.4–5.0 km s⁻¹ layer 2 and
353 6.2–7.0 km s⁻¹ layer 3. The depth of the base of the crust is 10 s TWTT. The well-controlled velocity
354 model indicates that the oceanic crust is 3–5 km thick, thinner than average normal oceanic crust
355 (6–8 km). This last observation agrees with other seismic measurements along the non-volcanic
356 North Atlantic continental margins (Ginzburg *et al.* 1985; Whitmarsh *et al.* 1990; Pinheiro *et al.* 1992;

357 White 1992a, 1992b), which show an abnormally thin oceanic crust immediately adjacent to the
358 continent–ocean transition. White (1992a, 1992b) suggested that the cause might be very slow sea-
359 floor spreading.

360 The magnetic anomalies of this domain have high amplitudes and a weakly marked linearity. They
361 are similar to those of the North Atlantic Ocean between the magnetic Chron 34 and the magnetic
362 Chron M0 (Macnab *et al.* 1995), which characterizes the Cretaceous Magnetic Quiet Period.

363 To a first approximation, our interpretation of the beginning of the oceanic domain agrees with the
364 sharp ocean–continent boundary of de Charpal *et al.* (1978) and Derégnaucourt & Boillot (1982)
365 (Fig. 1). Most workers (e.g. Bacon *et al.* 1969; Grau *et al.* 1973; Montadert *et al.* 1979b; Le Pichon &
366 Sibuet 1981; Derégnaucourt & Boillot 1982; Barbier *et al.* 1986) have postulated that the large
367 negative magnetic anomaly observed on the total magnetic field map (Fig. 2a) is the magnetic
368 signature of a basement ridge, the North Biscay Ridge, observed only on the OC17 profile (Figs 3b
369 and 4). On the map of magnetic anomalies reduced to the pole (Fig. 2b), this large negative
370 magnetic anomaly does not exist, but a strong magnetic gradient separates domain V from domain
371 IV (the very quiet magnetic zone). This observation agrees with the new seismic data, which show
372 that the North Biscay Ridge is not a basement ridge but a large and local plateau (Fig. 6).

373 Comparison of the seismic and magnetic data shows that the boundary between the oceanic
374 domain and the transitional domain identified on the seismic profiles coincides globally with a strong
375 magnetic gradient (Figs 2 and 6). However, off the eastern Armorican margin, a small shift exists
376 between the limits of the oceanic domain deduced from seismic data and magnetic data (Fig. 6).
377 This shift coincides with the presence of the ST bodies and some basement ridges, oriented WNW–
378 ESE and situated in the western prolongation of the Dôme Gascogne. These elements may have a
379 strong magnetic signature that influences the oceanic limit drawn from the magnetic data.

380 **Discussion**

381

382 ***The ocean–continent transition zone***

383 *The main characteristics of the Armorican ocean–continent transition zone.* The seismic data have
384 allowed us to image the ocean–continent transition zone along the Armorican margin. This zone,
385 previously considered as a sharp contact (Derégnaucourt & Boillot 1982), is in fact a wide zone (c.
386 80 km) that coincides with the Armorican Basin (Fig. 6). This transitional domain shares the principal
387 characteristics of other ocean–continent transition zones surveyed: a very ‘quiet’ magnetic
388 signature, occurrence of shallow high-velocity materials (high-velocity lower-crustal layer), and the
389 existence of a subhorizontal and smooth basement surface (MS reflector), which systematically
390 overlies a complex reflectivity zone including both landward- and seaward-dipping reflectors (DR
391 reflectors). However, the Armorican ocean–continent transition zone shows some distinctive

392 features in comparison with other ocean–continent transition zones. It is marked by the presence of
393 Moho reflections (Mn reflector) beneath the ocean–continent transition zone, by the occurrence of
394 volcanic bodies in proximity to the oceanic domain and by particularities of the high-velocity lower-
395 crustal layer, which is covered directly by sediments and overlies Moho reflections. Locally this high-
396 velocity lower-crustal layer penetrates the oceanic domain, where it underplates the thin Cretaceous
397 oceanic crust.

398 *Nature and origin of the high-velocity lower-crustal layer.* The high-velocity lower-crustal layer has
399 velocities ($7.4\text{--}7.5\text{ km s}^{-1}$) lower than those measured within the normal mantle (8 km s^{-1}), and
400 greater than those measured within oceanic layer 3 or the lower continental crust ($6.5\text{--}7\text{ km s}^{-1}$).
401 Various interpretations of the high-velocity lower-crustal layer have been formulated (see the first
402 section of this paper). It could be oceanic material, produced by ultraslow sea-floor spreading
403 (Srivastava *et al.* 2000). However, the smooth character of the MS reflector is not typical of the
404 surface of oceanic crust, the velocities of the basement are too high and this area has no linear
405 magnetic anomalies (Fig. 2b). Second, this material could represent an extremely thinned, and
406 possibly broken up, continental crust underplated and intruded by partial melt generated by
407 asthenospheric upwelling, as suggested by Whitmarsh *et al.* (1990) for the Tagus Abyssal Plain. In
408 this case the DR reflectors, which characterize the seismic facies of the high-velocity lower-crustal
409 layer, could represent traces of the volcanic intrusions. However, the existence of an extremely
410 thinned continental crust ($<4\text{ km}$ thick), with a smooth surface, without conspicuous structures,
411 throughout the Armorican Basin, seems to us unlikely. Third, the high-velocity lower-crustal layer
412 could be mafic magmas underplated during the rifting period, as shown by underplating beneath
413 other volcanic rifted margins (Skogseid *et al.* 1992; White 1992a), but also beneath the Iberia
414 Abyssal Plain non-volcanic margin (Cornen *et al.* 1996, 1999) and, further south, in the Gorringe
415 Bank area (Girardeau *et al.* 1998), where a 5 km thick by 80 km long gabbro body was described at
416 the top of the mantle. Our data cannot exclude such an origin. Another possibility is to consider the
417 high-velocity lower-crustal layer as an abnormal hydrated mantle that extends along the entire West
418 Iberia margin. In this case, the DR reflectors (Fig. 12) may represent faults, important for
419 serpentinization, and the bulges of the substratum (Fig. 13) may correspond to serpentinite diapirs
420 induced by a volume increase as a result of serpentinization (Recq *et al.* 1996).

421 This last interpretation of the high-velocity lower-crustal layer is supported by its seismic structure,
422 which presents some characteristics comparable with other ocean–continent transition zones: the
423 landward- and seaward-dipping reflectors (DR reflectors) under the MS reflector are comparable
424 with those described in the West Iberia ocean–continent transition zone by Pickup *et al.* (1996). Its
425 geographical position against the continental slope, its width and its magnetic signature are
426 comparable with those of the West Iberia margin. The overlying material can also be compared with
427 that found off the West Iberia margin: Unit 3B (Figs 9 and 12a) exhibits a similar seismic signature
428 to that of the Enigmatic Terrane described on Galicia Bank by Boillot *et al.* (1995). Also, the seismic

429 image of Unit 3C (Fig. 12b) exhibits numerous similarities with images from the Iberia Abyssal Plain
430 basement (see Fig. 16b, below; Pickup *et al.* 1996). If this last interpretation is correct, it implies that
431 the rifting in the Bay of Biscay took place, at least initially, with little or no magmatic activity.

432 ***The nature of the 'neck area' and its significance***

433 The 'neck area' (zone III, Fig. 6) is the last domain oceanward, at the foot of the continental slope,
434 where structural elements attributed to the continental crust have been observed. The rare tilted
435 blocks are restricted to this area. They comprise a low crustal thickness and a thick pre-rift unit.
436 They could also represent a few fragments of the more superficial part of the upper continental
437 crust. There are very few tilted blocks along the Armorican margin (five at most) compared with the
438 Western Approaches margin (e.g. Montadert *et al.* 1979b; Whitmarsh *et al.* 1986; Fig. 2a).

439 The blocks overlie variable crustal materials, such as the deep-layered crust or the high-velocity
440 lower-crustal layer (Figs 8 and 11). If we follow the conventional hypothesis these blocks reflect a
441 normal product of the stretching phase of the continental crust. However, according to Montadert *et al.*
442 (1979b) and Chenet *et al.* (1983), who worked on the Western Approaches margin, the extension
443 rate calculated from the tilted block geometry is too low to justify an overall crustal thinning. The few
444 blocks observed off the Armorican margin have a similar involvement. An alternative implication is to
445 consider that these blocks represent gravity slide structures related to the flexural subsidence of the
446 margin. In this case, the tilted blocks are uninvolved in the processes of crustal thinning. The 'neck
447 area' would represent a 'glide block area' and the true limit of the continental domain (the locus of
448 continental break-up) would be situated at the base of the slope. The continental slope would
449 therefore give, apart from rare exceptions, a good approximation of the precise extent of the
450 continent domain.

451 A similar deep-layered unit observed on the Norgasis profiles (Figs 7, 8 and 11) is observed also at
452 the foot of the West Iberia continental slope: the IAM 9 profile (see Fig. 16c, below) indeed exhibits
453 some small tilted blocks resting on a highly reflective material including landward-dipping reflectors.
454 This deep-layered unit could correspond to a part of the thinned and stretched layered lower
455 continental crust, as it displays similar seismic facies and velocities. However, we do not have direct
456 seismic data that allow us to confirm a genetic relationship between the deep-layered unit and the
457 layered lower continental crust of the shelf. Alternatively, this deep-layered crust could be new crust,
458 possibly underplated during the rifting phase. In the latter case, the base of the continental slope
459 would be the locus of the continental break-up.

460 Previous interpretations of the S reflector suggested that it corresponds to: (1) the brittle–ductile
461 transition within continental crust (de Charpal *et al.* 1978; Le Pichon & Barbier 1987), or as deduced
462 from the Galicia margin; (2) a detachment fault that penetrates the entire lithosphere (Boillot *et al.*
463 1989); (3) an intra-crustal detachment fault (Sibuet 1992); (4) a detachment between upper

464 continental crust material and serpentinized upper mantle (Boillot *et al.* 1989; Chian *et al.* 1995); (5)
465 a complex boundary with features (3) and (4) (Boillot *et al.* 1995; Reston 1996; Reston *et al.* 1996);
466 (6) an abrupt transition between the faulted upper continental crust and the top of the high-velocity
467 lower-crustal layer (Louden & Chian 1999). Along the Armorican margin, the S reflector is observed
468 only off the continental slope, in the 'neck area'. Its occurrence always coincides with the presence
469 of tilted fault blocks. It therefore corresponds most closely to a décollement surface, on which the
470 faulted blocks have moved and rotated. This interpretation is in agreement with that of Avedik *et al.*
471 (1982) and Barbier *et al.* (1986) on the Western Approaches margin.

472 The lack of relationship between these structures and the faults of the upper continental slope
473 excludes the hypothesis that the S reflector is the trace of a long, low-angle detachment, i.e.
474 continuation of a 'breakaway' as often suggested (Barbier *et al.* 1986; Boillot *et al.* 1987b, 1989;
475 Sibuet 1992; Reston 1996; Reston *et al.* 1996; Manatschal & Nievergelt 1997). On the Norgasis
476 reflection and refraction data, the S reflector appears to be the prolongation of the top of the high-
477 velocity lower-crustal layer that corresponds to the MS reflector (Fig. 11). On the basis of our whole
478 seismic reflection dataset, it appears that the S reflector corresponds to an intra-crustal detachment
479 surface of tilted blocks, and evolves towards the ocean into interfaces between faulted blocks and
480 either the deep-layered crust or the upper mantle (Boillot *et al.* 1987a, 1995; Reston 1996; Reston
481 *et al.* 1996; Manatschal & Nievergelt 1997; Louden & Chian 1999). Off the Armorican margin, the S
482 reflector has no direct seismic relationship with the location of crustal thinning. We think that the S
483 reflector has restricted influence within the zone of tilted blocks and seems to have had a minor role
484 in the crustal thinning processes.

485 ***The thinning of the continental crust***

486 Our data demonstrate that crustal thinning of the Armorican margin is restricted to a very narrow
487 area that corresponds to the continental slope, of 15–50 km width. The seismic profiles show that
488 the crustal thickness decreases under the continental slope from about 35 km at the shelf break to
489 less than 10 km at the foot of the slope (Fig. 8b). The crustal thinning is underlain by a major
490 escarpment, which breaks the upper crust above a striking Moho rise. The apparent expression of
491 the crustal thinning is thus only a very small amount of extension in the upper continental crust. The
492 behaviour of the layered lower continental crust beneath the slope is not well imaged by our data, as
493 a result of the reflectivity loss, but the steep rise of the mantle implies its disappearance (Fig. 8a).
494 The domain immediately at the foot of the continental slope corresponds to the 'neck area', which
495 displays some faulted blocks separated from a deep-layered unit by the S reflector. The tilted blocks
496 are considered here to be gravity slide structures. We consider thus that the continental slope
497 represents, as a first approximation, the locus of the initial continental break-up.

498 **Conclusions**

499 Integration of the new seismic reflection and refraction data, from the continental shelf to the
500 oceanic domain, highly constrains the shallow and deep structure of the Armorican margin.
501 Interpretation of the entire seismic dataset has resulted in a structural map of the Armorican margin
502 that defines five main domains: the unextended continental domain (the shelf), the thinned
503 continental domain (the slope), the oceanic domain, a wide transitional zone and a domain called
504 the 'neck area'. We know now the 3D crustal geometry of the Armorican margin. A structural sketch
505 of the Armorican margin that summarizes the observations made is given in Fig. 15.

506 The main conclusions are as follows.

507 (1) The ocean–continent boundary, previously proposed to be a sharp contact, is in fact an ocean–
508 continent transition zone of 80 km width. This zone coincides with a major part of the Armorican
509 Basin. It shares the main characteristics of other ocean–continent transition zones: very quiet
510 magnetic signature, occurrence of shallow high-velocity material (the high-velocity lower-crustal
511 layer, 7.4–7.5 km s⁻¹) and of a subhorizontal basement surface that systematically overlies a
512 complex reflectivity zone including both the landward- and seaward-dipping reflectors. However, the
513 Armorican ocean–continent transition zone does show some distinctive features compared with
514 other ocean–continent transition zones. It is marked by the presence of Moho reflections beneath
515 the ocean–continent transition zone, the existence of volcanic bodies close to the oceanic domain
516 and by particularities of the high-velocity lower-crustal layer, which is overlain by the sediments.
517 Compared with the other non-volcanic passive North Atlantic margins (White 1992a, 1992b), the
518 oceanic crust immediately adjacent to the Biscay continent–ocean transition is thinner than the
519 normal oceanic crust. The high-velocity lower-crustal layer of the Armorican ocean–continent
520 transition zone could represent an abnormal mantle constituted by serpentinized peridotites.

521 (2) The North Biscay Ridge, as defined in the 1970s, does not exist along the North Biscay margin,
522 but instead is part of a large plateau situated off the North Armorican margin. This plateau may be
523 due to local underplating of the high-velocity lower-crustal layer under the oceanic crust.

524 (3) Continental crustal thinning is restricted to the narrow continental slope (15–50 km wide) with the
525 corollary that the contribution of crustal stretching was very limited. The new seismic profiles show
526 that the crustal thickness decreases under the continental slope from about 35 km at the shelf break
527 to less than 10 km at the foot of the slope. Crustal thinning is expressed along a major escarpment
528 in the upper continental crust superimposed on a steep shallowing of the mantle. We consider that
529 the continental slope represents, as a first approximation, the geometry of the initial continental
530 break-up.

531 (4) The S reflector and extensional structures such as tilted blocks are observed exclusively at the
532 base of the continental slope in the narrow domain called the 'neck area'. The Norgasis seismic
533 data show the following features. (a) A deep-layered crust is squeezed between the S reflector and

534 the Moho. It could be explained as a part of the thinned and stretched layered lower continental
535 crust or as a crust underplated or intruded during the rifting phase. (b) The S reflector has influence
536 restricted to the zone of tilted blocks. It seems to correspond to an intra-crustal detachment surface
537 of tilted blocks that evolves towards the ocean into an interface between faulted blocks and either
538 the deep-layered crust or the upper mantle. (c) The tilted blocks are considered to be gravity slide
539 structures. We conclude that the 'neck area' represents a 'glide block area'.

540 In comparison with other divergent margins, we consider that the Armorican margin is not atypical.
541 The Labrador, West Greenland, Orphan Basin, Southern Grand Banks, Nova Scotia and Flemish
542 Cap margins present some very similar features: they all have narrow continental slopes, which are
543 the loci of the main crustal thinning (see Keen & Dehler 1997; Louden & Chian 1999, p. 745; Fig. 1).
544 The crustal geometry of the Armorican margin (Fig. 15) can also be compared with the West Iberia
545 margin, which also exhibits a narrow continental slope linked to a sharp escarpment, and a high
546 upper mantle uprising as exemplified by the IAM 9 profile (Fig. 16).

547 This paper has examined only one segment of the North Biscay margin, the Armorican margin. The
548 geometry of the Armorican margin seems very different from the published structural interpretations
549 of the Western Approaches margin (Avedik & Howard 1979; Montadert *et al.* 1979b). In contrast, the
550 Armorican margin has a wide ocean–continent transition zone and fewer tilted blocks.

551 Numerous questions remain: what are the geodynamic processes that permitted the creation of a
552 wide ocean–continent transition zone along the Armorican margin? Why are the two segments of
553 the North Biscay margin, the Western Approaches margin and the Armorican margin, so different?

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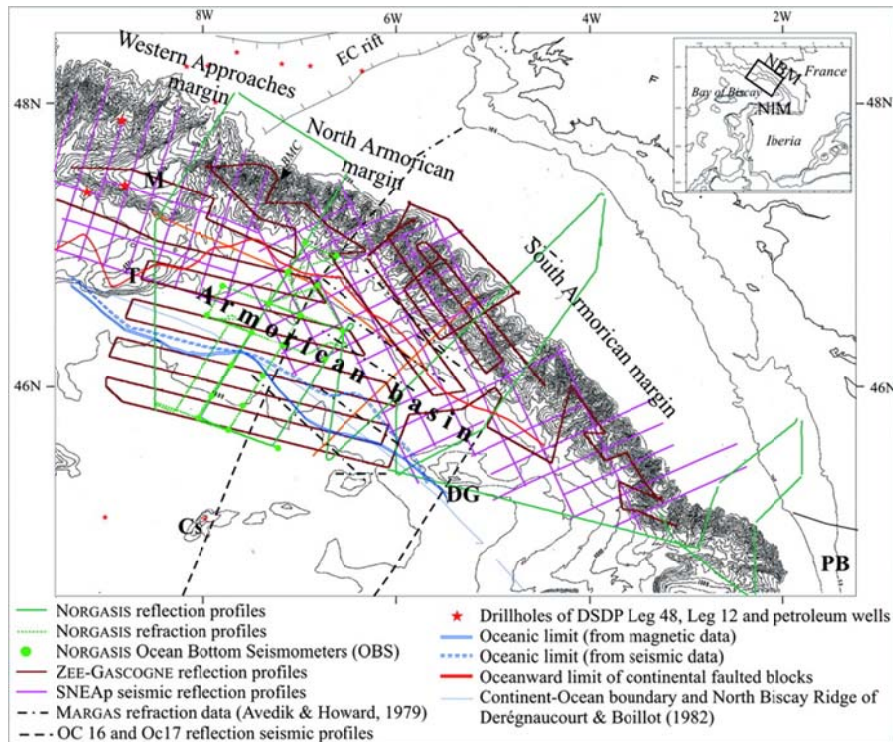
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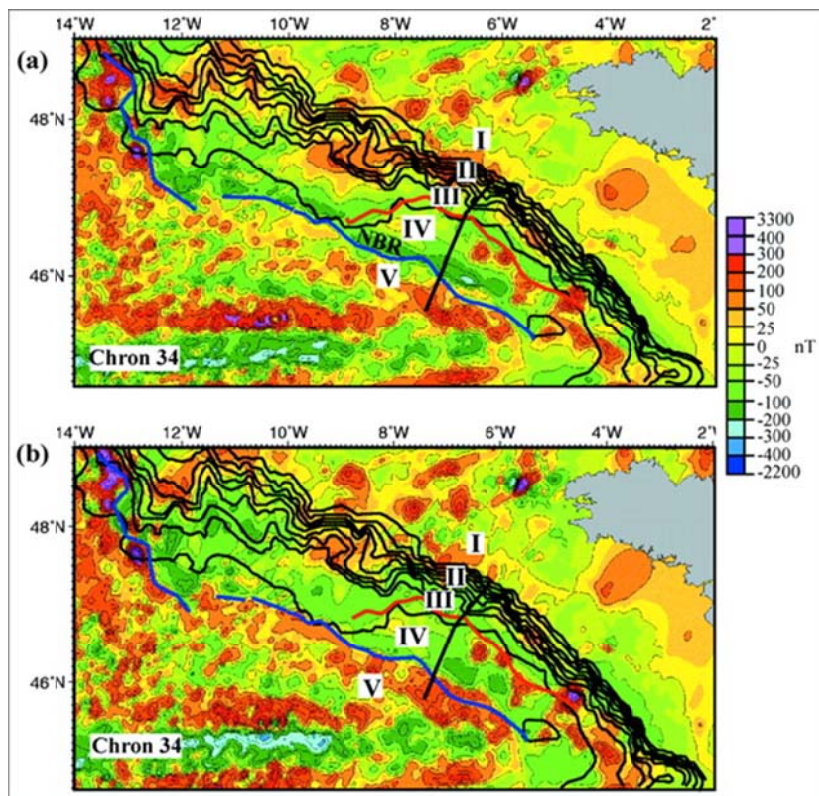
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Fig. 1. Bay of Biscay seismic survey location map. Isobath interval of the bathymetry (Sibuet et al. 1994) is 200 m. Mercator projection 1/2 400 000 at 41°N latitude, showing locations of the deep drillholes (Deep Sea Drilling Program (DSDP); petroleum). PB, Parentis Basin; M, Meriadzek Terrace; T, Trevelyan Seamount; BMC, Black Mud Canyon; Cs, Cantabria Seamount; DG, Dôme Gascogne; EC rift, English Channel Rift; NBM, North Biscay margin; NIM, North Iberia margin

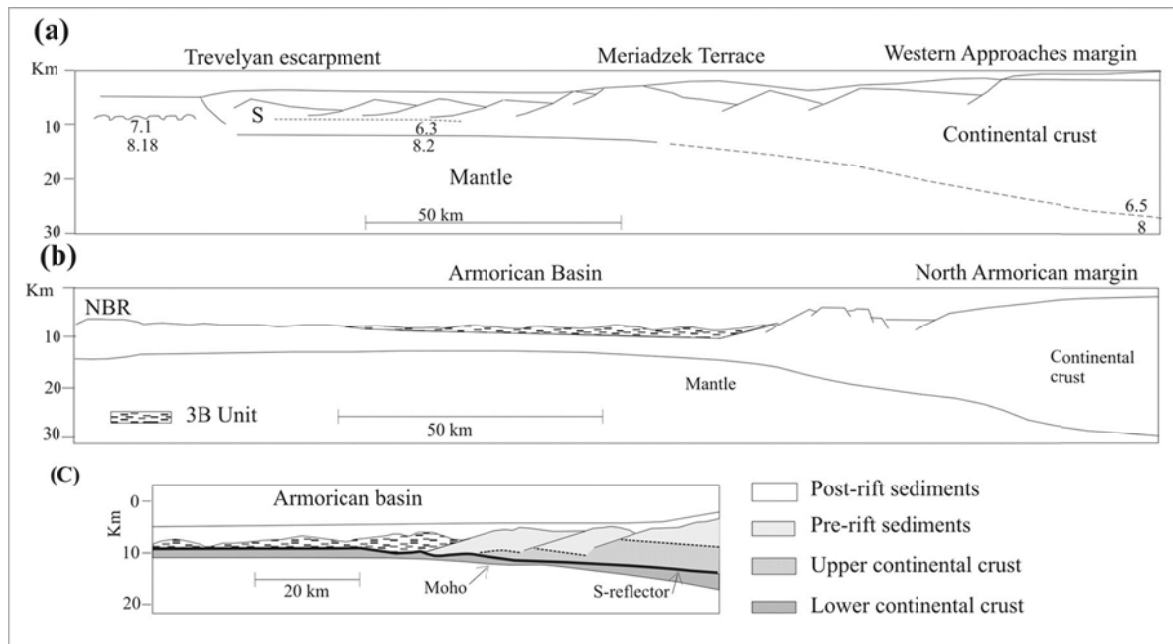


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Fig. 2. (a) Map of the total field magnetic anomalies (numerical magnetic data collected by Verhoef et al. 1996). (b) Map of the magnetic anomalies reduced to pole, from the total field magnetic data of Verhoef et al. (1996). The pole reduction of magnetic data has been made with surfer and gmipack software in collaboration with BRGM (Orléans, France). The structural zones (blue line, oceanic limit from seismic data; red line, limit of

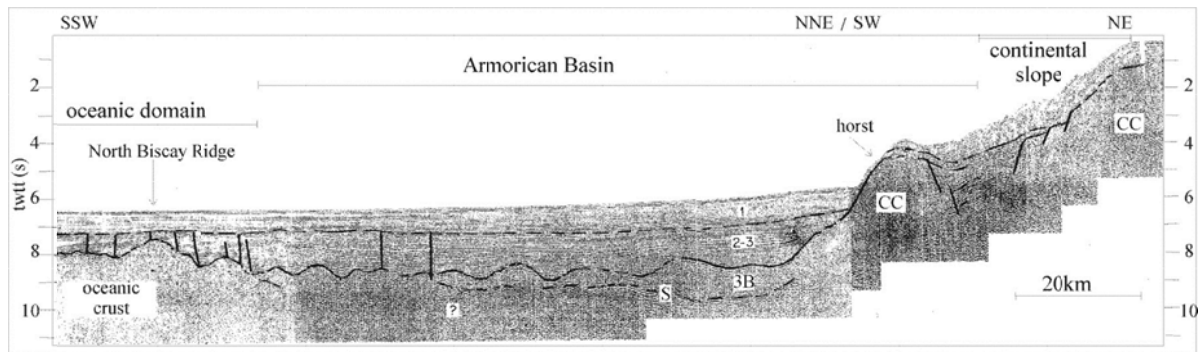
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the last continental blocks) and the OC17 seismic reflection profiles are reported here. I, continental shelf; II, continental slope; III, 'neck area'; IV, ocean-continent transition zone; V, true oceanic domain; NBR, North Biscay Ridge.



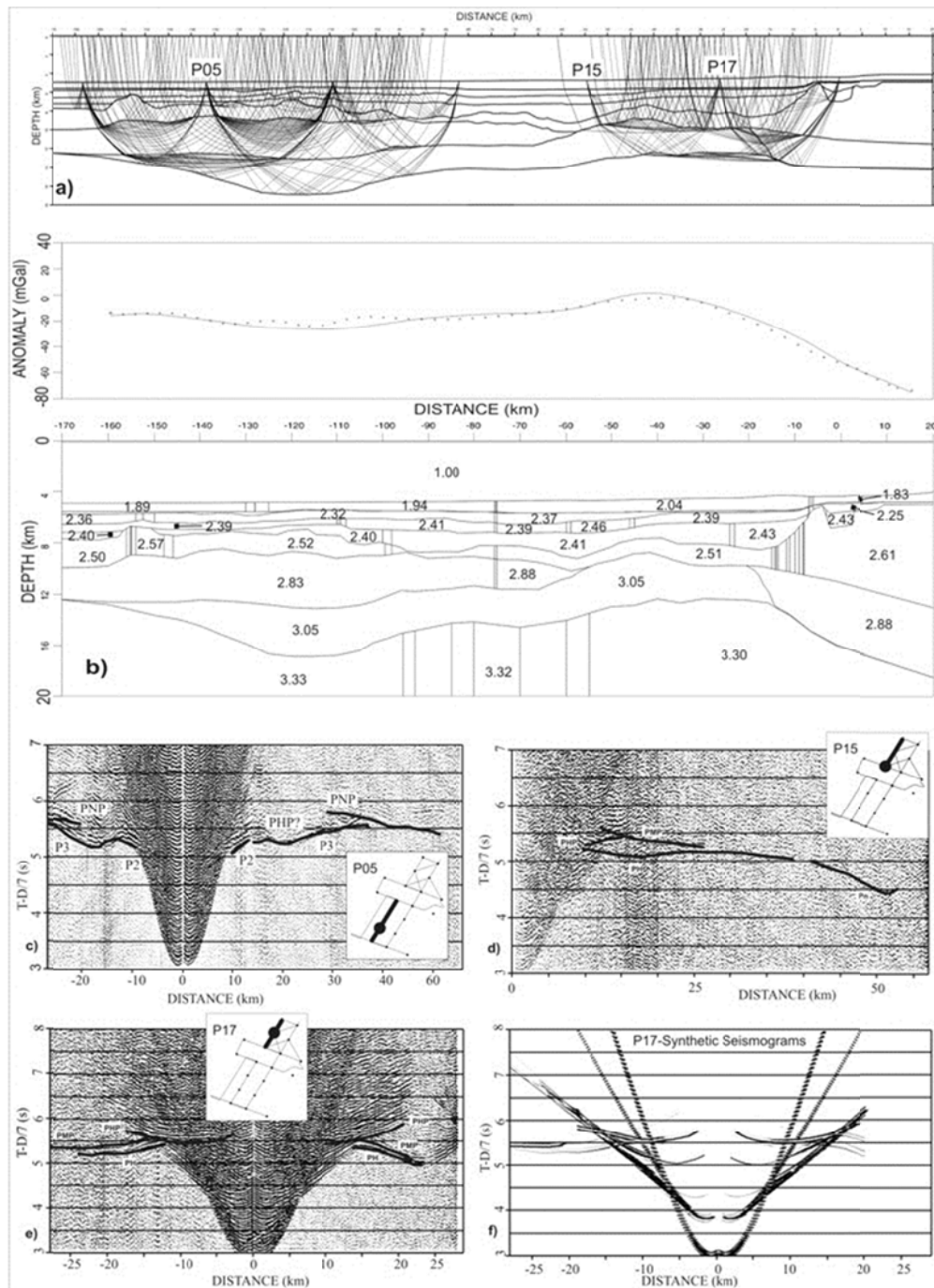
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Fig. 3. Published structural schema of the North Biscay margin. (a) The Western Approaches margin from Avedik & Howard (1979) and Montadert et al. (1979). (b) The Armorican margin from Le Pichon & Sibuet (1981). NBR, North Biscay Ridge. (c) The Armorican margin from Barbier et al. (1986) and Le Pichon & Barbier (1987).



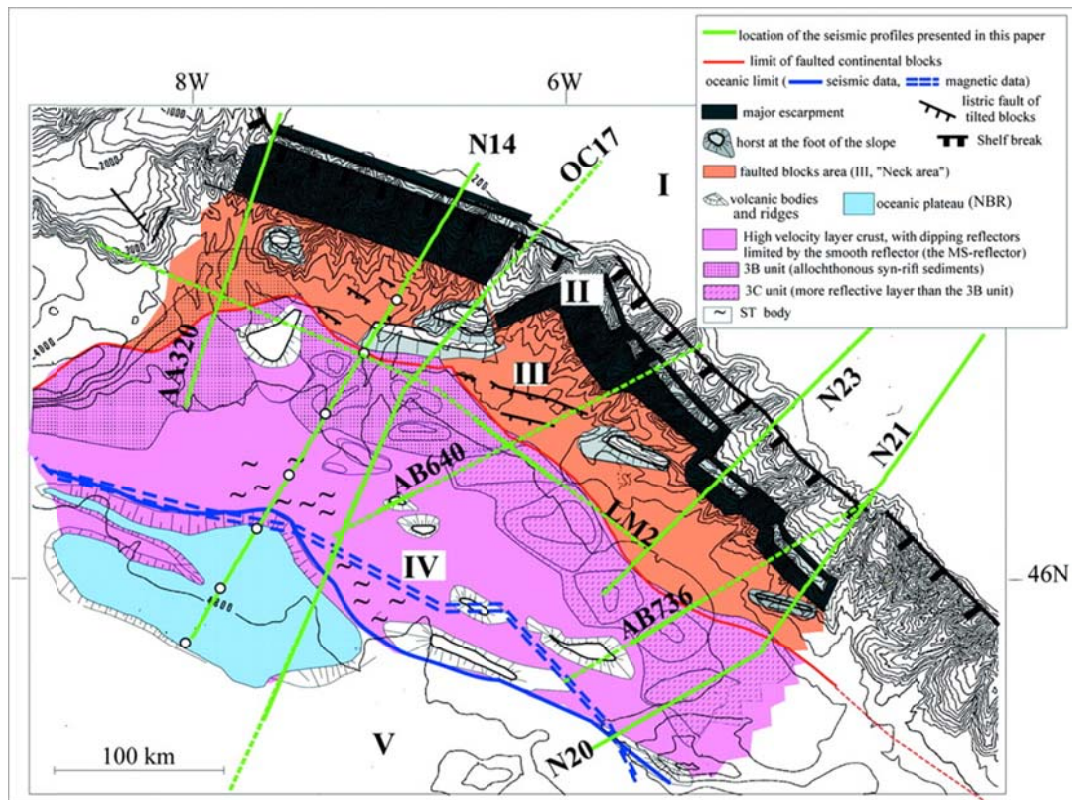
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Fig. 4. OC17 seismic profile (for location, see Fig. 6) from Debyser et al. (1971). CC, Continental crust; S, S reflector; 3B, enigmatic unit; 1, 2 and 3 are post-rift sediments.

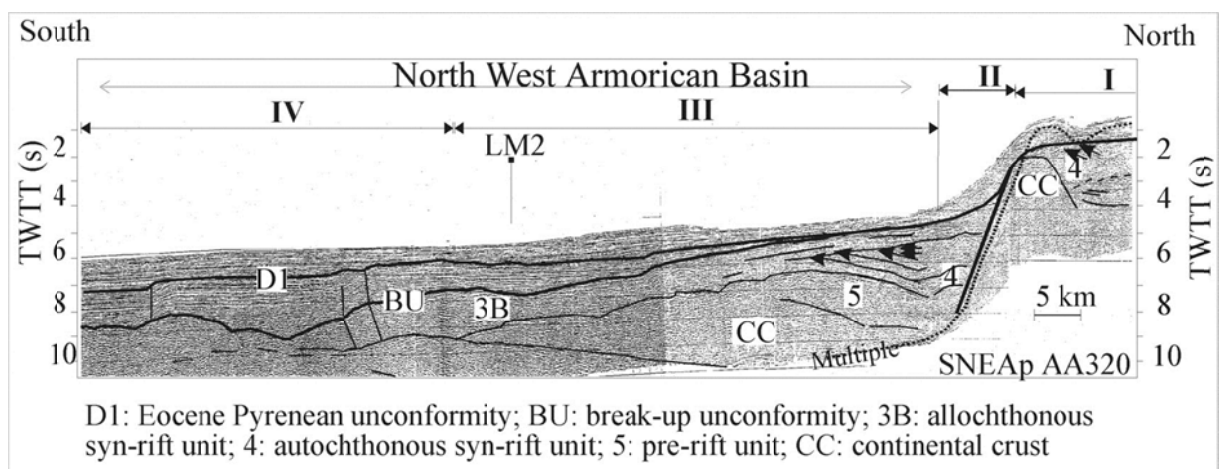


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Fig. 5. (a) Structural model for profile Norgasis 14 with the ray paths that were used to define the deeper layers (only one ray in every four is plotted). The velocities (see Fig. 7b) and depths are constrained by reflected and diving waves. (b) Observed () and computed free-air gravity for the Norgasis 14 profile after transformation of P-wave velocities to density using the data from Ludwig et al. (1970). To improve the fit the model geometry was slightly modified in the unconstrained part near the coast and the density of the high-velocity lower-crustal layer above normal mantle was reduced. To account for lateral velocity or density variations within the same layer, the model is divided into blocks separated by vertical boundaries. The large number of these blocks is the consequence of strong heterogeneity. (c), (d) and (e) show examples of seismic refraction data for ocean-bottom seismometers P05, P15 and P17, respectively, plotted with 7 km s reduction velocity. An offset-dependent gain and a bandpass filter (6–18 Hz) have been applied. The computed travel times for the interpreted phases are overlain. Interpreted phases: P2, refracted arrival from oceanic layer 2; P3, refracted arrival from oceanic layer 3; PHP, refracted arrival from the top of the high-velocity lower-crustal layer; PH, refracted arrival from the high-velocity lower-crustal layer; PMP, reflected arrival from the Moho; Pn, refracted arrival from the top of normal mantle. (f) Synthetic seismograms computed by asymptotic ray theory for the P17 ocean-bottom seismometer, using a Ricker wavelet as source function. Same gain and plotting parameters as for the data section were used.



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Fig. 6. Structural map of the Armorican margin. The location of cross-section (green line) included as figures is shown. ○, ocean bottom seismometers used for the refraction models along the Norgasis 14 profile.



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Fig. 9. Stacked SNEAp reflection profile (for location see Fig. 6) across the North Armorican margin. A faulted block, composed of the upper continental crust (CC) and a thick pre-rift unit (5), is tilted at the base of a major escarpment that defines the continental slope. Unit 3B covers the autochthonous synrift unit (4), the continental blocks and the basement of ocean-continent transition zone. The intersection with the Loire Maritime 2 profile (LM2) is reported here. 3B, Allochthonous synrift unit; BU, break-up unconformity; D1, Pyrenean unconformity.

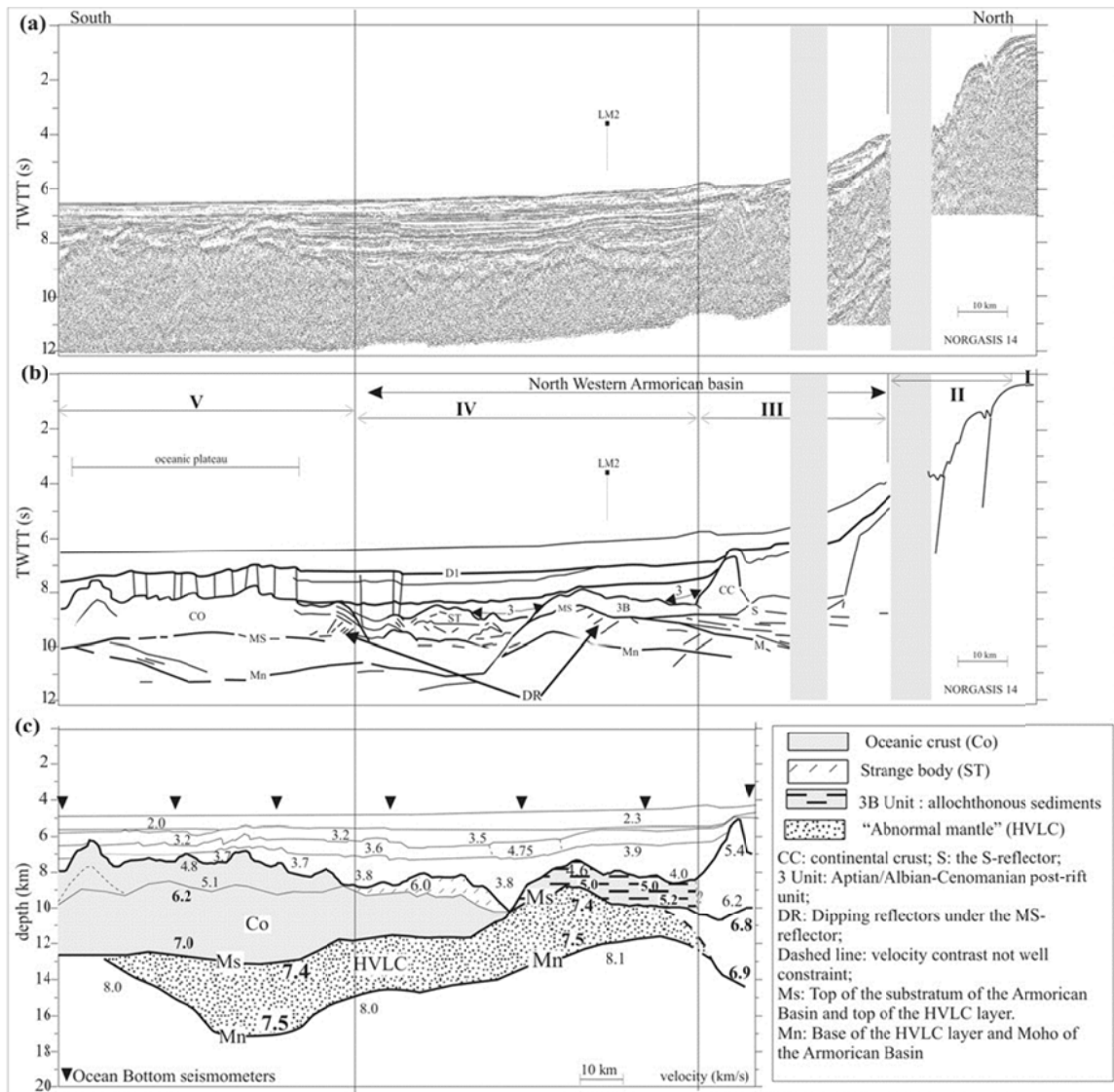
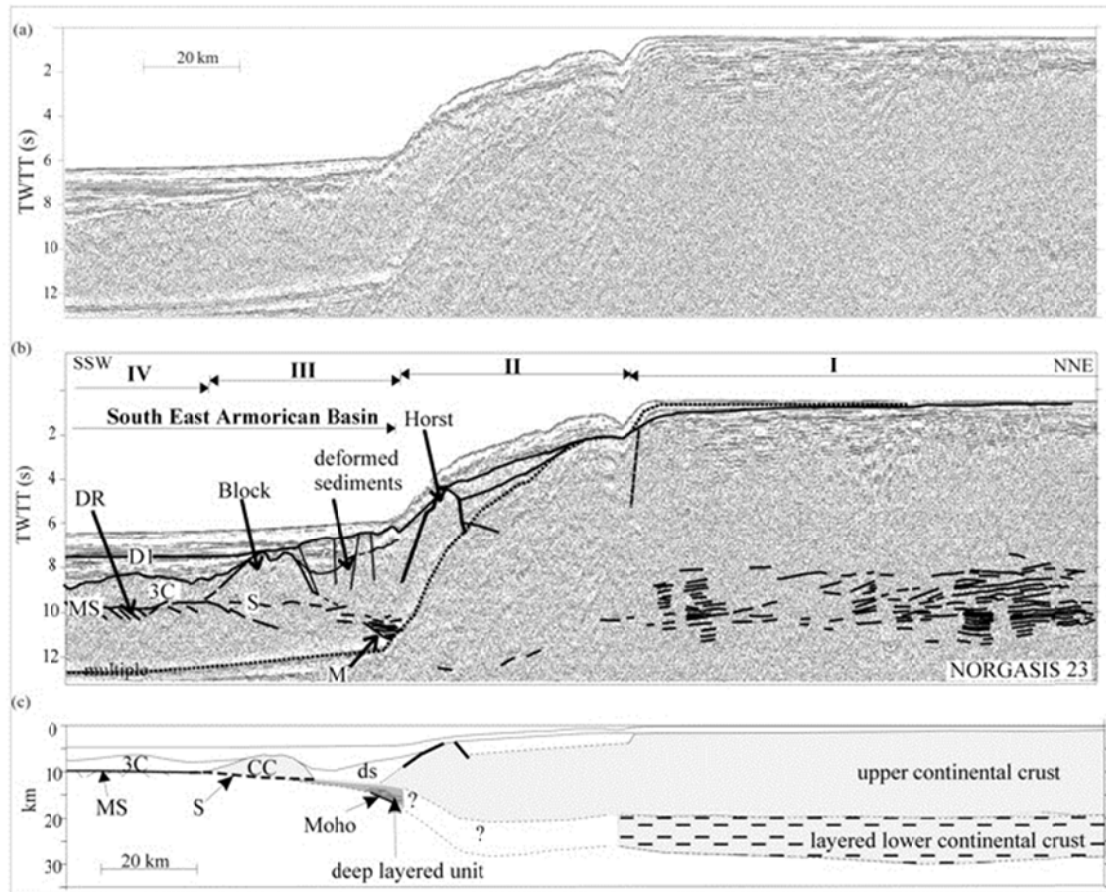


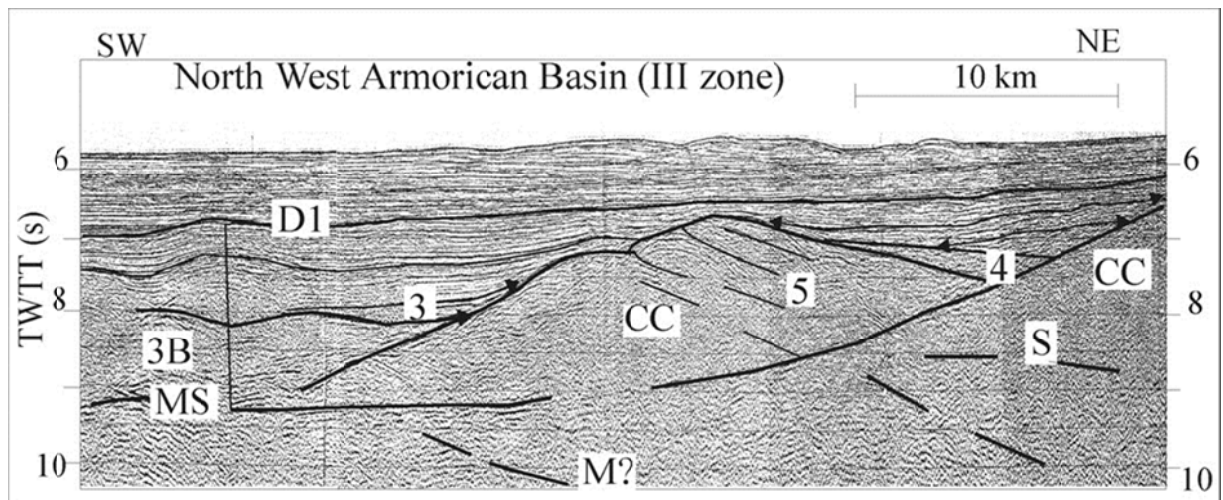
Fig. 7. (a) Time-migrated reflection profile (Norgasis 14) across the North Armorican margin from the shelf to the oceanic domain across the ocean-continent transition zone; (b) line drawing of this profile; (c) refraction model computed by L. Matias and A. Hirn from Norgasis ocean-bottom seismometer data along the Norgasis 14 reflection profile (for location see Fig. 6). The intersection with the Loire Maritime 2 (LM2) profile is reported here. D1, Pyrenean unconformity (Eocene time); M, Moho reflections.

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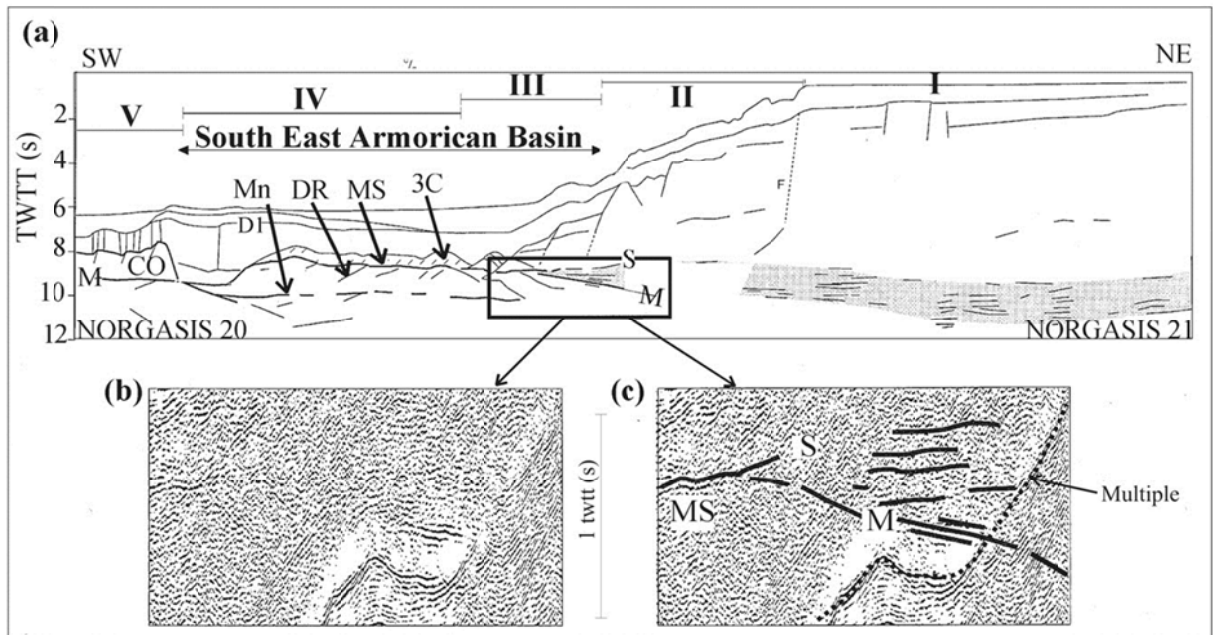
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Fig. 8. (a) Time-migrated reflection profile (Norgasis 23; for location see Fig. 6) across the South Armorican margin; (b) interpreted profile; (c) synthetic depth section without vertical exaggeration. D1, Pyrenean unconformity; MS, basement of the OCT zone; DR, dipping reflectors; S, S reflector; M, Moho reflections; 3C, enigmatic seismic unit; CC, continental crust; ds, deformed sediments.



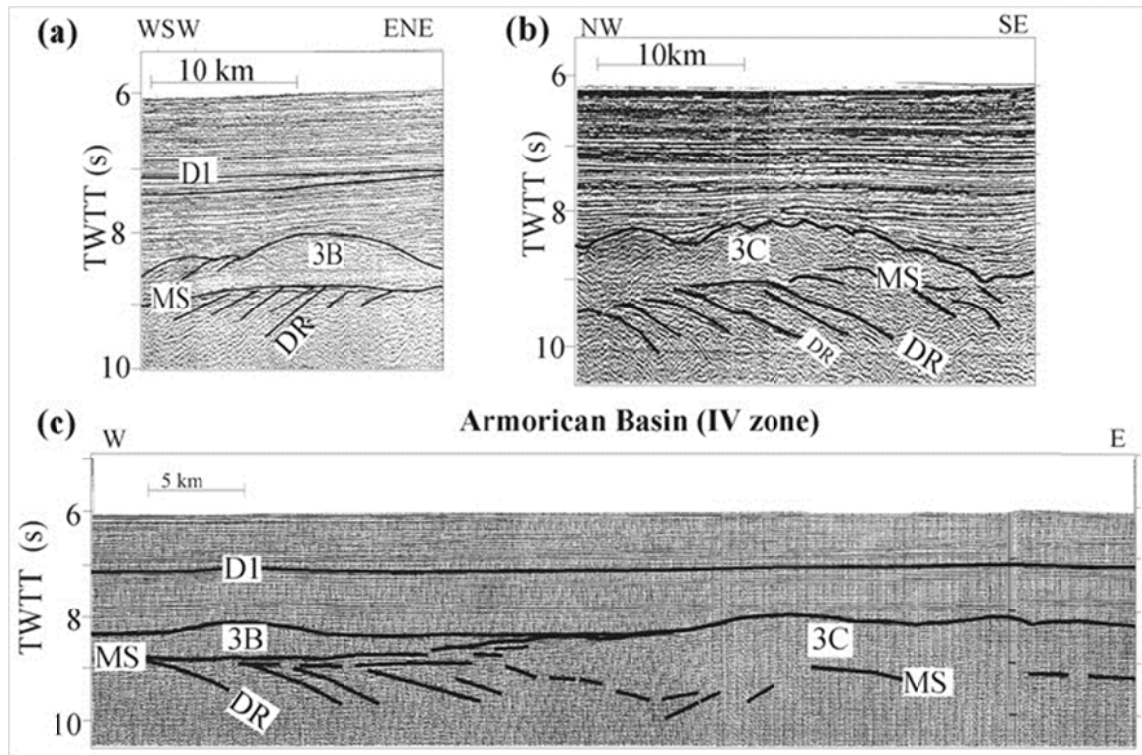
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Fig. 10. Section of stacked AB640 SNEAp reflection profile (for location see Fig. 6) across the Armorican 'neck area'. 3B, Allochthonous synrift unit; D1, Pyrenean unconformity; 3, post-rift sediments; 4, synrift sediments; 5, pre-rift sediments; MS, basement of the ocean-continent transition zone; S, S reflector; M, Moho reflections; CC, continental crust



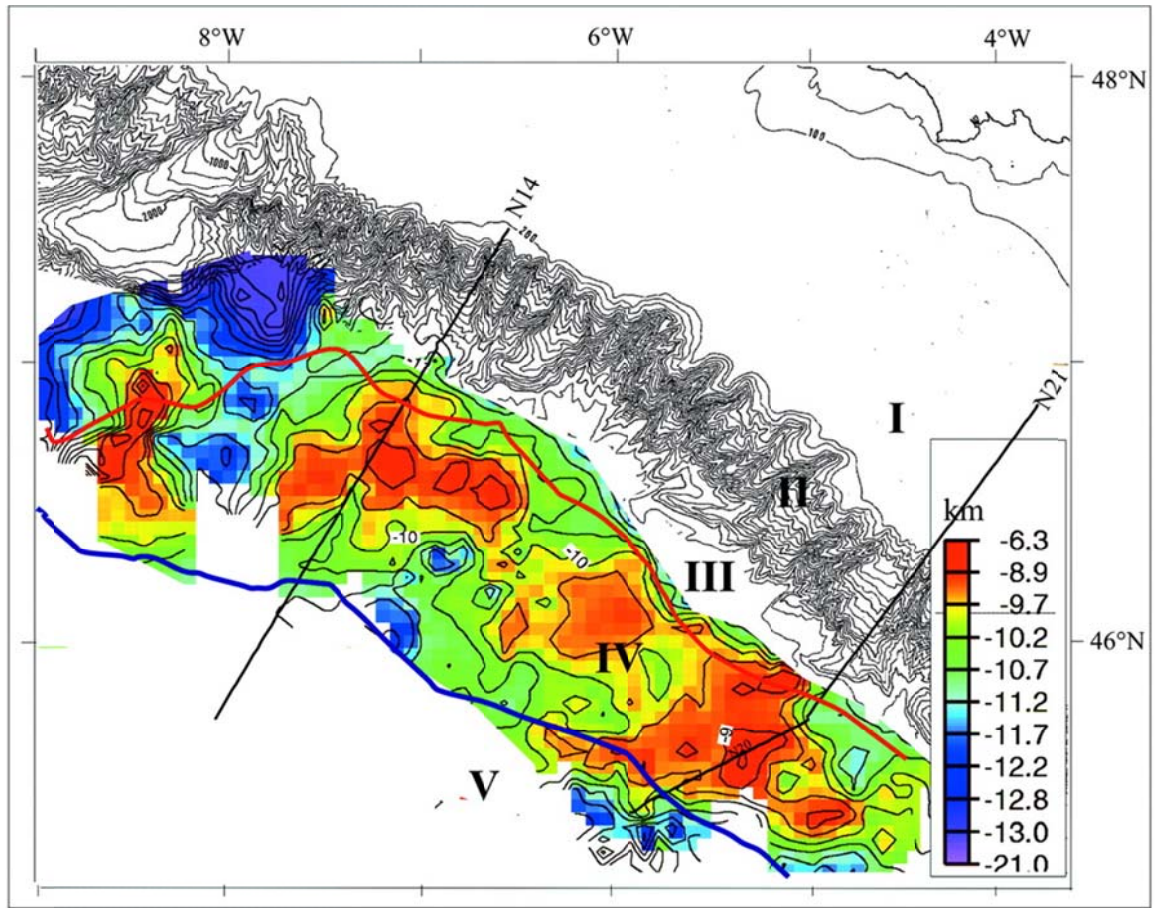
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Fig. 11. (a) Line drawing of time-migrated reflection profiles (Norgasis N20 and N21; for location see Fig. 6) across the South Armorican margin; (b) enlargement showing the deep layered unit and the relationships between the S, MS and M reflectors; (c) interpretation of the enlargement. D1, Pyrenean unconformity; MS, basement of the ocean–continent transition zone; S, S reflector; M, Moho reflections; DR, dipping reflectors; Mn, top of the normal mantle; 3C, enigmatic seismic unit; CC, continental crust; CO, oceanic crust.



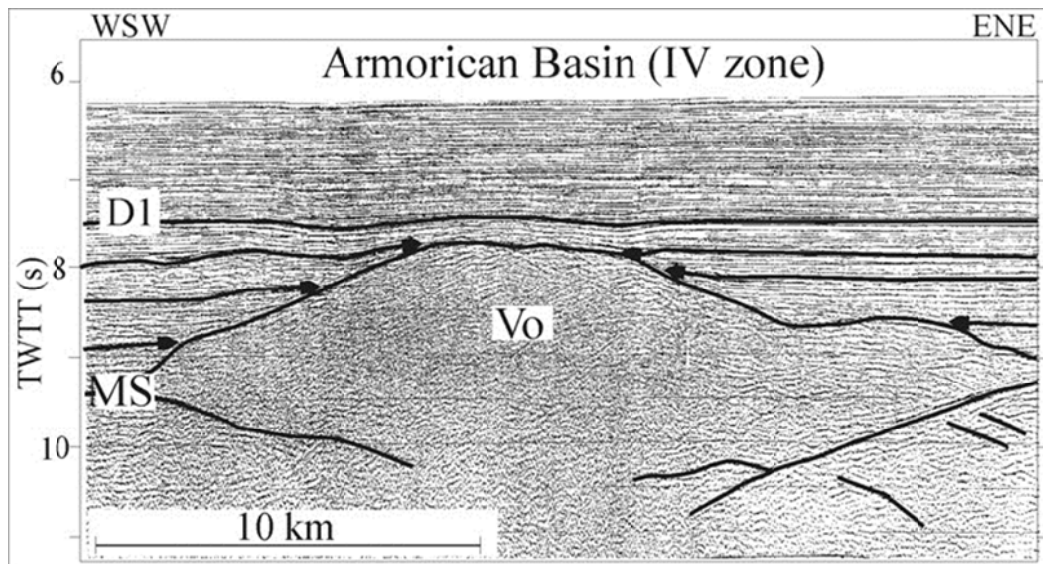
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Fig. 12. (a) Enlargement of SNEA profile (for location see Fig. 6) showing the typical seismic character of the western Armorican ocean–continent transition zone (Fig. 7). Unit 3B covers the MS reflector, which truncates the dipping reflectors beneath. (b) Enlargement of SNEA profile (for location see Fig. 6) showing the different seismic character of Unit 3C observed only in the eastern Armorican ocean–continent transition zone (Fig. 6). (c) Interpreted sections of Loire Maritime 2 profile (LM2). 3B, allochthonous synrift unit; D1, Pyrenean unconformity; MS, basement of the ocean–continent transition zone; DR, dipping reflectors.



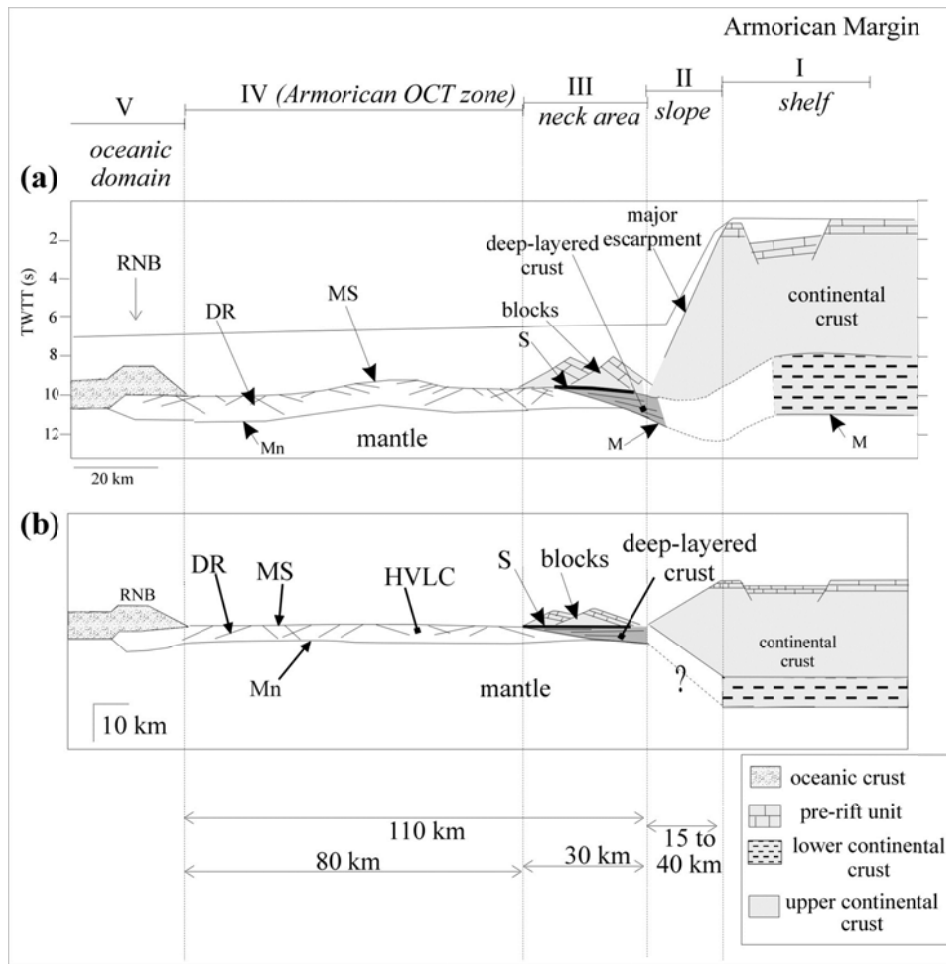
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Fig. 13. Morphology and depth (km) of the substratum of the transitional domain. The time (TWTT (s)) of the digitized MS reflector has been converted to depth (km) with the velocities of Norgasis ocean-bottom seismometers. The magnetic (double black line) and seismic (bold blue line) oceanic limits and the limits of the continental blocks (red line) are shown.

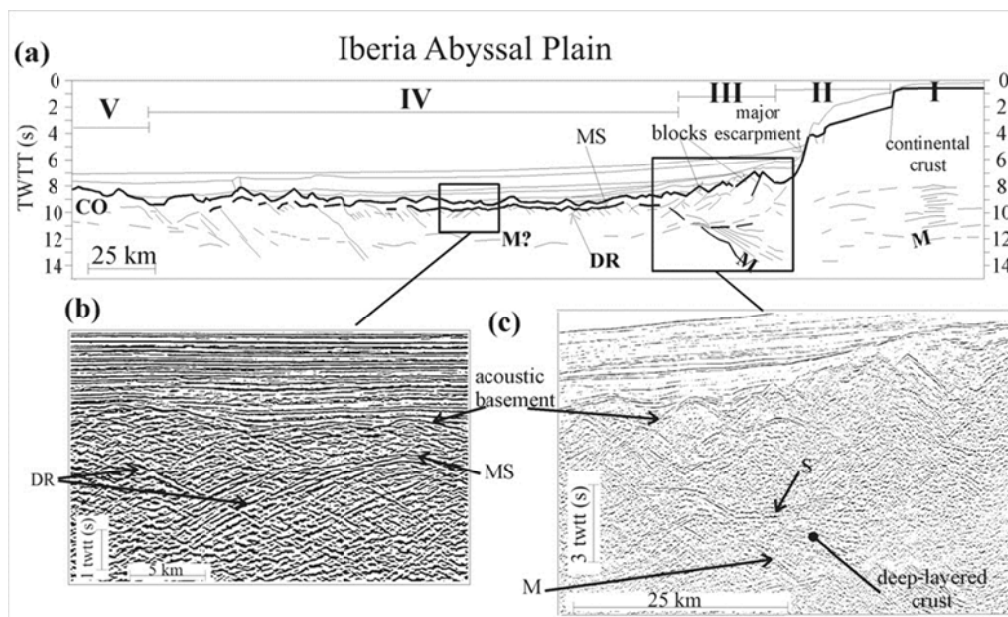


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Fig. 14. An SNEAp AB640 seismic reflection profile across the body with a conical shape (for location see Fig. 6). D1, Pyrenean unconformity; MS, basement of the ocean-continent transition zone; Vo, volcanic body.



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Fig. 15. Schematic illustrations of archetypal crustal section across the Armorican segment of the North Biscay margin. (a) Synthetic time section. (b) Synthetic depth section without vertical exaggeration. MS, Basement of the ocean–continent transition zone; M, Moho; DR, dipping reflectors; S, S reflector; Mn, top of the normal mantle of the Armorican Basin; HVLC, high-velocity lower-crustal layer; NBR, North Biscay Ridge.



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Fig. 16. (a) Reinterpreted section of IAM 9 profile across the West Iberia margin. (b) Enlargement of the acoustic basement of the ocean–continent transition zone (Pickup et al. 1996). (c) Enlargement of the deep-layered crust under the tilted blocks, at the base of the continental slope. MS, Basement of the ocean–continent transition zone; M, Moho; DR, dipping reflectors; S, S reflector; CO, oceanic crust.