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Isabelle Thinon, Luis Matias, Jean-Pierre Réhault, Alfred Hirn, Luis Fidalgo-González, et al.. Deep structure of the Armorican Basin (Bay of Biscay): a review of Norgasis seismic reflection and refraction data. Journal of the Geological Society of London, 2003, 160, pp.99-116. 10.1144/0016-764901-103 . hal-00647845

## HAL Id: hal-00647845 https://brgm.hal.science/hal-00647845

Submitted on 5 Jan 2012  $\,$ 

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

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- 11 Received 30 July 2002; revised typescript accepted 27 August 2002
- 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 considered this margin typical of a non-volcanic passive margin: it is characterized by a striking 16 17 succession of tilted blocks beneath which occurs the S reflector and the continent-ocean boundary is abrupt. This paper examines the Armorican segment and is based on a study of all early seismic 18 profiles together with new multichannel reflection and refraction seismic data (Norgasis cruise). An 19 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<sup>-1</sup>) 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 than a long ridge elongated off the whole North Biscay margin. 26

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

#### 28 Introduction

From the West Iberia margin to the Goban Spur margin, the zone of transition between thinned continental crust and typical oceanic crust was first proposed as a sharp transition, less than 10 km wide (e.g. Montadert *et al.* 1979a; Avedik *et al.* 1982; Derégnaucourt & Boillot 1982; de Graciansky *et al.* 1985; Ginzburg *et al.* 1985; Boillot *et al.* 1987a; Whitmarsh *et al.* 1990; Pinheiro *et al.* 1992; Sibuet *et al.* 1992). More recently, off the West Iberia non-volcanic margin, seismic profiles and deep boreholes have indicated that this contact may involve a wider (30–120 km) zone referred to as the ocean–continent transition (e.g. Pickup *et al.* 1996; Whitmarsh & Sawyer 1996) or the zone of
exhumed continental mantle (Whitmarsh *et al.* 2001). The ocean–continent transition zone coincides
with a very quiet magnetic zone (Whitmarsh *et al.* 1990) whose magnetic anomaly amplitudes are
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 (between 4.5 and 6.0 km s<sup>-1</sup>), and are underlain by a lower basement layer characterized by high 40 41 velocities of 7.2-7.6 km s<sup>-1</sup> 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<sup>-1</sup> (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<sup>-1</sup>), and higher than that 46 measured within the oceanic layer 3 or within lower continental crust (6.5-7 km s<sup>-1</sup>). This is a 47 common feature of this and other ocean-continent transition zones and is referred to as the highvelocity lower-crustal layer (Louden & Chian 1999). Currently, ocean-continent transition zones are 48 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 61 al. 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 during prolonged periods of extension with little or no melt generated from the upper mantle. 63 Although the 7.2-7.6 km s<sup>-1</sup> velocities are characteristic of the ocean-continent transition zones of 64 65 some non-volcanic rifted margins, they are also observed beneath volcanic passive margins, where they would represent underplating gabbros or a lower continental crust intruded by mafic intrusions 66 (Skogseid et al. 1992; White 1992a). One of the differences between volcanic and non-volcanic 67 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 sediments consist of Jurassic (150-140 Ma, Kimmeridgian to Portlandian) platform carbonates. As 96 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

The Bay of Biscay was surveyed extensively in the 1970s and 1980s (Debyser *et al.* 1971;
Montadert *et al.* 1971, 1974, 1979; Derégnaucourt & Boillot 1982; Barbier *et al.* 1986; Le Pichon &
Barbier 1987). Oceanographic surveys began again in 1994 and 1997 (e.g. the Norgasis cruise;
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 byBacon et al. (1969). 130 Resting on the basement, Unit 3B, a seismically transparent unit characterized by velocities of 4.4 131 km s<sup>-1</sup> (Bacon et al. 1969) to 4.6 km s<sup>-1</sup> (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 137 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 140 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 Bay of Biscay (Debyser et al. 1971: de Charpal et al. 1978: Le Pichon & Sibuet 1981), Using new 148 149 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.

The Norgasis seismic reflection profiles were acquired with a 96-channel streamer (25 m interval) using a single bubble array with 8–10 air guns and a generator capacity from 804 to 1230 in<sup>3</sup> (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 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<sup>-1</sup> in its original velocity (from 7.2 to 7.4 km s<sup>-1</sup>). 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<sup>-1</sup>) and the high-velocity lower-crustal layer (7.4 km s<sup>-1</sup>) 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.

More recently, we made a multibeam survey of this zone during the Zee-Gascogne cruise in 1997 on board the R.V. *l'Atalante* (Fig. 1). During this cruise, seismic reflection profiles were acquired at 10 knots with a six-channel streamer, two Generator injector guns (105 in<sup>2</sup> each), and with 10 s TWTT recording. A conventional seismic processing scheme (stack, migration) using the ProMax software was applied to the Norgasis and Zee-Gascogne reflection data (Thinon 1999). This complete and homogeneous dataset (18 000 km) provides new information on the deep structure across this non-volcanic passive margin.

#### 222 Interpretation

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

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

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

Faulted and tilted continental blocks are restricted to a deeper 30 km wide domain, at the base of 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<sup>-1</sup> 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<sup>-1</sup>; 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<sup>-1</sup>, 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)

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

*Substratum of transitional domain.* Throughout the transitional domain, the top of the substratum is an acoustic reflector, the MS reflector that gives rise to reflections with high amplitudes and low frequencies that were in the past considered to represent the S reflector. The MS reflector systematically truncates the dipping reflectors (DR) in the underlying material (Figs 7, 8 and 12). It is situated at depths of 8–9.5 s TWTT (Figs 7 and 12). Its regional topography is characterized by large wavelength bulges (Fig. 13), generally independent of the overlying structures. This indicates 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  $(5.0-5.2 \text{ km s}^{-1} \text{ v}. 7.4 \text{ km s}^{-1})$ , at the top of a 3-4 km thick layer with velocities of 7.4-7.5 km s<sup>-1</sup> (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<sup>-1</sup> v. 8.0 km s<sup>-1</sup>) 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<sup>-1</sup>, 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).

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

*Boundaries of the transitional domain.* The geometric relations between the continental and the transitional domains are well imaged on the Norgasis lines (Fig. 11). The oceanward continental block of the 'neck area' corresponds approximately to the junction of the S, M and MS reflectors (Figs 7, 8 and 11). Figure 11c shows that the MS reflector lies on the prolongation of the junction between the M reflector (continental Moho) and S reflector (base of tilted blocks and top of the 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<sup>-1</sup> 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<sup>-1</sup> layer 2 and 353 6.2–7.0 km s<sup>-1</sup> 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 (6-8 km). This last observation agrees with other seismic measurements along the non-volcanic 355 356 North Atlantic continental margins (Ginzburg et al. 1985; Whitmarsh et al. 1990; Pinheiro et al. 1992;

- White 1992a, 1992b), which show an abnormally thin oceanic crust immediately adjacent to the continent–ocean transition. White (1992a, 1992b) suggested that the cause might be very slow seafloor spreading.
- The magnetic anomalies of this domain have high amplitudes and a weakly marked linearity. They are similar to those of the North Atlantic Ocean between the magnetic Chron 34 and the magnetic 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 signature of a basement ridge, the North Biscay Ridge, observed only on the OC17 profile (Figs 3b 368 369 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).
- Comparison of the seismic and magnetic data shows that the boundary between the oceanic domain and the transitional domain identified on the seismic profiles coincides globally with a strong magnetic gradient (Figs 2 and 6). However, off the eastern Armorican margin, a small shift exists between the limits of the oceanic domain deduced from seismic data and magnetic data (Fig. 6). This shift coincides with the presence of the ST bodies and some basement ridges, oriented WNW– ESE and situated in the western prolongation of the Dôme Gascogne. These elements may have a strong magnetic signature that influences the oceanic limit drawn from the magnetic data.

#### 380 **Discussion**

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#### 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 signature, occurrence of shallow high-velocity materials (high-velocity lower-crustal layer), and the 388 389 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 features in comparison with other ocean–continent transition zones. It is marked by the presence of Moho reflections (Mn reflector) beneath the ocean–continent transition zone, by the occurrence of volcanic bodies in proximity to the oceanic domain and by particularities of the high-velocity lowercrustal layer, which is covered directly by sediments and overlies Moho reflections. Locally this highvelocity lower-crustal layer penetrates the oceanic domain, where it underplates the thin Cretaceous oceanic crust.

398 Nature and origin of the high-velocity lower-crustal layer. The high-velocity lower-crustal layer has 399 velocities (7.4–7.5 km s<sup>-1</sup>) lower than those measured within the normal mantle (8 km s<sup>-1</sup>), and 400 greater than those measured within oceanic layer 3 or the lower continental crust (6.5-7 km s<sup>-1</sup>). 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 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

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image of Unit 3C (Fig. 12b) exhibits numerous similarities with images from the Iberia Abyssal Plain
basement (see Fig. 16b, below; Pickup *et al.* 1996). If this last interpretation is correct, it implies that
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

The 'neck area' (zone III, Fig. 6) is the last domain oceanward, at the foot of the continental slope, where structural elements attributed to the continental crust have been observed. The rare tilted blocks are restricted to this area. They comprise a low crustal thickness and a thick pre-rift unit. They could also represent a few fragments of the more superficial part of the upper continental crust. There are very few tilted blocks along the Armorican margin (five at most) compared with the Western Approaches margin (e.g. Montadert *et al.* 1979*b*; 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 442 al. (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.

Previous interpretations of the S reflector suggested that it corresponds to: (1) the brittle–ductile
transition within continental crust (de Charpal *et al.* 1978; Le Pichon & Barbier 1987), or as deduced
from the Galicia margin; (2) a detachment fault that penetrates the entire lithosphere (Boillot *et al.*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**

Integration of the new seismic reflection and refraction data, from the continental shelf to the oceanic domain, highly constrains the shallow and deep structure of the Armorican margin. Interpretation of the entire seismic dataset has resulted in a structural map of the Armorican margin that defines five main domains: the unextended continental domain (the shelf), the thinned continental domain (the slope), the oceanic domain, a wide transitional zone and a domain called the 'neck area'. We know now the 3D crustal geometry of the Armorican margin. A structural sketch 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<sup>-1</sup>) 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.

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

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

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

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

In comparison with other divergent margins, we consider that the Armorican margin is not atypical. The Labrador, West Greenland, Orphan Basin, Southern Grand Banks, Nova Scotia and Flemish Cap margins present some very similar features: they all have narrow continental slopes, which are the loci of the main crustal thinning (see Keen & Dehler 1997; Louden & Chian 1999, p. 745; Fig. 1). The crustal geometry of the Armorican margin (Fig. 15) can also be compared with the West Iberia margin, which also exhibits a narrow continental slope linked to a sharp escarpment, and a high 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.* 1979*b*). 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?

#### 554 Acknowledgements

555 The first author thanks J.-L. Olivet and D. Aslanian for their interest and help during the evolution of 556 this study, and Ifremer and Institut Universitaire Européen de la Mer (IUEM) for their support. We 557 are grateful to D. Needham, J. Girardeau and G. Cornen for critical reviews of the manuscript and 558 constructive discussions, and reviewers are gratefully acknowledged for their pertinent comments 559 helpful for the final manuscript. C. Truffert (BRGM) provided much appreciated help in the 560 processing of the magnetic data. R. Le Suavé and G. Auffret gave us free access to the data 561 collected during the Zee-Gascogne and Sédifan cruises, respectively.

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### **FIGURES**



**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



**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

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.



**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).



*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.



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 offsetdependent 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, reflected 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.



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



**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.



*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.



**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



**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.



**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.



**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.



**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.



**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.



**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.