P wave velocity model of the crustal structure of the Nubia-Eurasia plate boundary in the Gloria Fault segment

Modelo de velocidades das ondas P para a estrutura da crosta no segmento da Falha da Glória na fronteira de placas Nubia-Eurasia

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Abstract: Interpretation of seismic refraction data across the Gloria Fault segment reveals the existence of a 7.0 to 7.4 kms\(^{-1}\) P wave velocity layer between the lower crust and the upper mantle. The most likely rock materials that can correspond to those velocities in the lithosphere, in such a setting, are serpentinites. The dataset from 18 OBS (Ocean Bottom Stations) are described and analyzed. A seismic velocity model presents the clear evidence of the Gloria Fault and various degrees of deformation of different stratigraphic layers. The geological implications of the obtained velocity/geological models are discussed.

Keywords: Plate boundary, Gloria fault, Velocity model, Crustal structure, Serpentinization.

Resumo: A interpretação de um conjunto de dados de sísmica de refração na Falha da Glória revelou a existência de uma camada, entre a crosta inferior e o manto superior, com uma velocidade das ondas P entre 7,0 e 7,4 kms\(^{-1}\). Os materiais rochosos que podem corresponder a estas velocidades na litosfera, neste tipo de ambiente geodinâmico, são serpentinitos. A descrição e análise de um conjunto de dados obtido através de 18 OBS permitiram a elaboração de um modelo de velocidades que mostra a clara evidência da Falha da Glória e dos vários graus de deformação das diferentes camadas estratigráficas.

As implicações geológicas dos modelos obtidos são discutidas.

Palavras-chave: Fronteira de placas, Falha da Glória, Modelo de velocidades, Estrutura da crosta, Serpentinização.

1. Introduction

The beginning of the Pangea breakup led to the opening of the Central Atlantic, initiating the Azores-Gibraltar fracture zone and the “proto” plate boundary between continental Africa and Iberia/Newfoundland (Dewey et al., 1989; Fernandez et al., 2004; Seton et al., 2012). The plate boundary changed location from north of the King’s Trough at M0 to Eocene times, to King’s Trough, Bay of Biscay and Pyrenees from Eocene to Anomaly 10 to finally jumped to south to the Azores-Gibraltar fracture zone (Srivastava et al., 1990; Roest & Srivastava, 1991). At the beginning (~190 Ma) the spreading rate was very slow (half-spreading rates of ~8 mm/yr) with an increase in spreading at 170 Ma to 17 mm/yr until anomaly M0 (120 Ma) (Dewey et al., 1989; Fernandez et al., 2004; Seton et al., 2012) leading to seafloor spreading and mantle exhumation (Srivastava et al., 1990; Pinheiro et al., 1992).

At this time Africa's motion relative to Europe was sinistral strike-slip (Dewey et al., 1989) and Iberia moved together with Nubia (Sibuet et al., 2004; Seton et al., 2012). Since anomaly 13 Iberia has been part of the Eurasian plate and the Azores-Gibraltar fracture zone presents a dextral strike-slip motion at a rate of 4 mm/yr (DeMets et al., 2010).

The Gloria fault was viewed as a dextral strike-slip transform fault that accommodates the rotation of the Africa with respect to Eurasia oceanic plates between the Azores plateau and the Madeira-Tore Rise (Argus et al., 1989). Bird & Kagan (2004) described the Azores-Gibraltar Fracture Zone that comprises the Gloria Fault segment as sub-divided in various segments with strike-slip, extensional and convergent strain regimes. The zones where the 1941, Ms=8.4 and 1939, Ms=7.1 earthquakes occurred are restricted to the strike-slip deformation regime areas, although at a considerable distance with respect to the trace of the fault. The details of the superficial deformation associated with the Africa-Eurasia
plate boundary along the Gloria Fault are still poorly known, as well as crustal and mantle structure of the plate interaction of this plate boundary.

2. Data

The OBS data were acquired during the METEOR cruise M79-Leg2 in 2009 (Hübischer, 2012). A 130 km refraction profile and a reflection multichannel seismic (MCS) profile (w51) across the Gloria fault was performed (Fig. 1).

To record the seismic energy 18 OBS and a digital streamer with 144 channels made up of six multiple arrays performing an active length of 600m were deployed. The 18 OBS were divided in ten vertical component ocean-bottom seismometers, five ocean-bottom hydrophones and three DEPAS (Deutscher Geräte-Pool für amphibische Seismologie from University of Hamburg) ocean-bottom stations with seismometer and hydrophone. As energy source two 32lt BOLT air guns with 120 bar shooting air pressure were used to generate seismic signals every 60s at 4kn speed velocity over ground.

To work on the data we use a set of programs from the zelt package software named SGY2Z, ZPLOT and RAYINVR (developed by Colin Zelt) and a set of complementary programs adjusted by Luis Matias and Alexandra Afiflado from DEGGE-FCUL (Department of Geographic Engineering, Geophysical and Energy from the Faculty of Sciences of the University of Lisbon). The general methodology and modeling strategies are discussed in Zelt & Ellis (1988), Zelt & Smith (1992), Zelt & Forsyth (1994) and Zelt (1999) and a compilation of the method is described in Afiflado (2006).

The w51 MCS profile was processed by Christian Hübischer and his group from the Geophysics Department of the University of Hamburg.

3. Velocity model

To create the velocity model the interpretation of the w51 MCS profile was the base of depart for the sediments layer. The thickness and velocities for the crust and mantle layers was taken from White et al. (1992). The model was adjusted by comparing the observed and calculated P wave travel times and by comparing the observed data with the synthetic seismic sections generate by the velocity model.

A five layers velocity model was built from the identification of six different phases: Pg1- Phase refracted in layer 3; PgP- Phase reflected in the base of layer 3; Pg2-Phase refracted in layer 4; PnP- Phase reflected in the base of layer 4; Pn1- Phase refracted in layer 5; Pn2- Phase refracted in layer 6. Layer 1 presents a mean thickness between 0.5 km and 1 km, a top velocity of 1.78 km/s and base velocity of 3.37 km/s and looks to be composed by sediments and basaltic flows; layer 2 presents a mean thickness between 1.6 km and 2.6 km, a top velocity between 4.45 km/s-4.70 km/s and base velocity between 5.57 km/s-5.80 km/s and looks to correspond to the upper crust; layer 3 presents a mean thickness between 5 km and 6 km, a top velocity between 6.20 km/s-6.35 km/s and base velocity between 6.70 km/s-6.90 km/s and looks to correspond to the lower crust; layer 4 presents a mean thickness of 5 km, a top velocity of 7.00 km/s and base velocity of 7.40 km/s; layer 5 was established with 4km of thickness, the top velocity was 7.80 km/s and the base velocity 8.00 km/s and corresponds to upper mantle.

In the obtained model for the thickness and velocities it is noticeable that i) the resolution diminishes in depth, ii) the upper crustal thickness decreases in the vicinity of the Gloria Fault, iii) lower crust is thicker to the south of the Gloria Fault (Africa) and iv) L4 is defined based on anomalous velocity values, i.e. values 7.0 and 7.4 km/s, that are too high for lower crust and too low for normal mantle. To look for this evidence six velocity profiles, spaced by 20 km, were extracted from the velocity model. A comparison with a compilation of velocity profiles for normal oceanic crust and fracture zones made by White et al. (1992) (Fig. 2) shows that the general trend of typical oceanic crust regarding to the upper crust is followed by all the profiles. Apart of the differences in the thickness on upper and lower crust the gradient for the P wave velocity is consistent with a normal basalt and gabbro composition.

Before reaching the typical 8 km/s P wave velocity for the upper mantle an anomalous velocity layer (L4) was recognized. Christensen & Salisbury (1975) identify that the following lithologies could give rise to basal Layer 3 velocities of 7.2 km/s: olivine gabbro, partially serpenitized peridotite, pyroxenite and foliated amphibolite. Nonetheless, serpentinization plays an important role in the tectonics of slow spreading ridges (Francis, 1981) and according to Cogné & Humler (2004) the spreading rate in the Atlantic between 80 to 50 Ma was 10 mm/yr. This was the time for the creation of the crust in the study area which is in good agreement to think that the layer 4 is consists of partially serpenitized peridotite.
4. Conclusions

a) The model presents a mean thickness of 8.5 km for the crust and a 4 km for the serpentinite mantle.

b) The comparison between the profiles from the model and the profiles from White et al. (1992) puts in evidence that we do not have a match for normal oceanic crust or typical fracture zones in the study area. The model layer for the upper crust appears to have identical velocity gradient and thickness of normal oceanic crust but differs in the area of the fault.

c) The lower crust thickness is different in the northern and southern parts of the Gloria Fault, which is in agreement with the 20 Ma age difference between the Eurasia plate (youngest) and Africa plate (older).
Fig. 2. Comparison made by the six velocity profiles extract from the model and a compilation made by White et al. (1992) for the typical Atlantic oceanic crust with a range age from 59 to 127 Ma (red polygon) and fracture zones older than 50 Ma (blue polygon).

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References


