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Seismotectonic significance of the 2008–2010 Walloon Brabant seismic swarm in the Brabant Massif, Belgium

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ABSTRACT

Between 12 July 2008 and 18 January 2010 a seismic swarm occurred close to the town of Court-Saint-Etienne, 20 km SE of Brussels (Belgium). The Belgian network and a temporary seismic network covering the epicentral area established a seismic catalogue in which magnitude varies between M_L -0.7 and M_L 3.2. Based on waveform cross-correlation of co-located earthquakes, the spatial distribution of the hypocentre locations was improved considerably and shows a dense cluster displaying a 200 m-wide, 1.5-km long, NW-SE oriented fault structure at a depth range between 5 and 7 km, located in the Cambrian basement rocks of the Lower Palaeozoic Anglo-Brabant Massif. Waveform comparison of the largest events of the 2008–2010 swarm with an M_L 4.0 event that occurred during swarm activity between 1953 and 1957 in the same region shows similar P- and S-wave arrivals at the Belgian Uccle seismic station. The geometry depicted by the hypocentral distribution is consistent with a nearly vertical, left-lateral strike-slip fault taking place in a current local WNW–ESE oriented local maximum horizontal stress field. To determine a relevant tectonic structure, a systematic matched filtering approach of aeromagnetic data, which can approximately locate isolated anomalies associated with hypocentral depths, has been applied. Matched filtering shows that the 2008–2010 seismic swarm occurred along a limited-sized fault which is situated in slaty, low-magnetic rocks of the Mousty Formation. The fault is bordered at both ends with obliquely oriented magnetic gradients. Whereas the NW end of the fault is structurally controlled, its SE end is controlled by a magnetic gradient representing an early-orogenic detachment fault separating the low-magnetic slaty Mousty Formation from the high-magnetic Tubize Formation. The seismic swarm is therefore interpreted as a sinistral reactivation of an inherited NW–SE oriented isolated fault in a weakened crust within the Cambrian core of the Brabant Massif.

1. Introduction

Earthquake swarms or seismic swarms are defined as episodic sequences of a large number of seismic events that are clustered in space and time (Mogi, 1963). In contrast to a classical foreshock–mainshock–aftershock sequence, in which aftershock sequences may consist of numerous lower-magnitude earthquakes, seismic swarms are not marked by one single dominant event. The time history of a swarm is characterised rather by a gradual increase, or sometimes by a burst, of microseismic activity alternating with periods of low seismic rate or seismic quiescence. Sometimes dominant earthquakes may reach larger magnitudes during the course of a seismic swarm. The many small events in a seismic swarm can, however, often not be linked to an identifiable mainshock. This can result from a heterogeneous stress field in a weakened crust that lacks a single well-defined fault structure. If such a fault structure were present, it might be capable releasing higher strain resulting in a higher magnitude earthquake (Fischer et al., 2014; Mogi, 1963). To precisely relocate the numerous events within a seismic sequence, it is necessary to analyse waveform data that are recorded by a dense local seismic network close to the epicentre of an earthquake swarm, allowing the detection of a large number of small events.

In an intraplate continental tectonic setting, seismic swarms are commonly associated with stress perturbations caused by magmatic intrusions, volcanic activity and with gradual fluid transport in the seismogenic part of the crust (Hainzl, 2004; Hiemer et al., 2012; Schenk et al., 2012; Špičák, 2000). In volcanic areas, continental rift and subduction zones, large fluid- and gas movements such as CO_2 release along prominent faults or fault intersections can generate earthquake swarms (e.g. Ibs-von Seht et al., 2008; Lindenfeld et al., 2012). Within the Eurasian tectonic plate, far away from any plate boundary, intense geothermal seismic swarms occur for example in the French/Italian Alps (e.g. Barani et al., 2014; Daniel et al., 2011; Leclère et al., 2012) or in the West Bohemia/Vogtland area in the Eger rift zone (e.g. Fischer et al., 2014; Parotidis et al., 2003; Schenk et al.,

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2012). In these cases, the presence of suprahydrostatic overpressured fluids rising along a fault zone is often considered to trigger fault activity due to pore pressure changes, even when the fault is unfavourably oriented for reactivation (Leclère et al., 2012; Sibson, 1985) and especially if the crust is in a critical state (Parotidis et al., 2003).

Between 2008 and 2010, a seismic swarm occurred in the central part of Belgium in the basement rocks of the Lower Palaeozoic Anglo-Brabant Massif, here further referred to as the Brabant Massif. Although the studied 2008–2010 Walloon Brabant seismic sequence resembles other seismic swarms in terms of its temporal evolution, the lack of a main shock and the restricted spatial distribution, the Brabant Massif is not associated with any of the volcanic, geothermal or tectonic settings described above, nor is any induced seismicity going on. Although the seismicity within the seismotectonic zone of the Brabant Massif is considered as to be rather moderate, still a few of the largest (historical) earthquakes in Western Europe have occurred within this slate belt demonstrating its importance (Camelbeeck et al., 2007). Linking these historical large earthquakes to potential individual fault structures has been difficult in the past, because of a lack of aftershocks and due to the limited seismic network. The rapid installation of a local network in the epicentral area of the 2008–2010 seismic swarm, however, allows us to study for the first time an extremely well documented seismic event in the old geological structure of the Brabant Massif.

The aim of this study is to investigate the specific geological structure that the swarm took place on. First, the hypocentre location is improved by the cross-correlation of waveforms of co-located events. A 3D analysis of the relocated hypocentre distribution delineates the dimension of the structure and frames the seismic sequence within the tectonic structure of the Brabant Massif. Second, stress inversion of the focal mechanisms of the largest-magnitude events is performed, allowing derivation of a best-fitting stress tensor and discussion of its correspondence to the regional stress field in northwestern Europe. Third, waveforms of the largest earthquakes are compared to analogue waveforms of a 1953 seismic event that belongs to a seismic swarm that occurred between 1953 and 1957 at Court-Saint-Etienne in the general same epicentral area as the 2008–2010 swarm. Finally, using the orientation of the seismic swarm, we attempt to link the fault structure to a relevant tectonic structure via aeromagnetic data. These data are matched-filtered to isolate anomalies that are likely to be due to sources at depths of interest. Such structures are of interest as they can play an important role accommodating deformation in a current stress field.

2. Geological and seismological settings

2.1. Regional setting

The study area is located in Belgium in NW Europe, more than 1000 km away from the boundaries of the Eurasian plate. The 2008–2010 Walloon Brabant seismic sequence occurred within the Lower Palaeozoic Brabant Massif, a slate belt situated in the subsurface of the central and northern part of Belgium. Outcrops of the Brabant Massif are sparse and are only present in some incised river valleys along the southern rim of the Brabant Massif. The Brabant Massif dips towards the north and is mostly covered by Cretaceous chalk, Cenozoic sand and clays, and Quaternary loess sediments. Well data indicates that the thickness of the cover rapidly increases to 1000 m at the Belgium–Dutch border (Legrand, 1968). Not much is known about the current deformation in the Brabant Massif, but incised river outcrops suggest uplift of the southern part and gradual deepening of its northern part. To the West, the Brabant Massif laterally extends towards the United Kingdom forming part of the larger tectonic unit of the Anglo-Brabant deformation belt (Verniers et al., 2002) (Fig. 1). To the east, it has been traced using borehole data and geophysical data beneath the thick Devonian and Carboniferous sequences of the Campine Basin (Figs. 1 and 2), as far as to the seismically active Lower Rhine Embayment (Mansy et al., 1999). At its southern border, it is unconformably overlain by undeformed Middle Devonian deposits of the Brabant parautochton (white area between Brabant Massif and Ardenne allochton in Fig. 1). Further to the south, the Brabant parautochton is tectonically overthrust by the Variscan Ardenne allochton along the Midi–Aachen thrust, i.e. the Variscan front of the Rhenohercynean Zone (Fig. 1), during the late stage of the Variscan Orogeny in the Late Carboniferous. Because of its crystalline rigidity, the Brabant Massif acted as a backstop during the Variscan deformation resulting in oroclinal bending of the Palaeozoic deposits of the Ardenne allochton (Van Noten et al., 2012).

2.2. Structural grain of the Brabant Massif

The structure of the Brabant Massif resulted from the ~30 Ma long-lasting, Acadian, Brabantian Deformation event that took place between the late Llandovery (c. 430 Ma) and Emsian (c. 400 Ma) (Debacker et al., 2005; Sintubin et al., 2009). As indicated on lithostratigraphic subcrop maps (Fig. 2a), the Brabant Massif has an apparent symmetrical geometry with a central Cambrian metasedimentary core flanked at both sides by Ordovician and Silurian metasedimentary rocks (De Vos et al., 1993b; Legrand, 1968; Piesens et al., 2005). Structural field work, gravimetric and aeromagnetic anomaly maps (Fig. 2c) show that the Brabant Massif has a dominant NW–SE trending structural grain that curves into a ENE–WSW orientation towards the east. Owing to the high magnetic susceptibility of the rock formations (slate, siltstone, metasandstone, metagreywacke) of the Lower Cambrian Tubize Group, the Cambrian core is clearly visible on the aeromagnetic anomaly map (Chacksfield et al., 1993; Sintubin, 1999). In contrast, the Bouguer gravity anomaly (Fig. 2b) shows the opposite pattern, with positive anomalies related to the Ordovician–Silurian rim and gravity lows associated with the Cambrian core or other deep-seated bodies (Everaerts and De Vos, 2012; Piesens et al., 2005). The arcuate geometry of fold-and-cleavage patterns inferred from potential-field data throughout the slate belt illustrates the change in orientation of the structural grain from west to east (Sintubin, 1999; Sintubin et al., 2009). Tectonic inversion of the Cambrian to Silurian Brabant Basin resulted in the formation of a steep compressional wedge in which the Cambrian core is strongly deformed and is covered by a less deformed Ordovician–Silurian at its peripheral domains. The decrease in deformation from the core to the peripheral domain is characterised by a decrease in metamorphic grade towards the Silurian flanks of the Brabant Massif and by a change of a steep fold belt in the central part to a rather wide and open fold belt in the peripheral part. The SW boundary of the Brabant Massif corresponds with a steep east–west oriented gravity gradient on the Bouguer anomaly map and the highest relief and most prominent magnetic highs on the aeromagnetic anomaly map. *La Bordière fault* defines the southern limit of the Brabant Massif and coincides with a sharp gravimetric anomaly gradient (Legrand, 1968) (Fig. 2b) juxtaposing the dense rocks of the Brabant Massif from less dense Upper Palaeozoic rocks of southern Belgium (Chacksfield et al., 1993).

In the SW part of the Brabant Massif, several NW–SE trending Bouguer anomaly gravity lows (indicated as gl in Fig. 2b) are interpreted as low-density crystalline basement at a minimum depth of 2.5 km (Everaerts et al., 1996; Lee et al., 1993). The NNE–SSW shortening and arcuate shape of the Brabant Massif is believed to be caused by the compression of the Cambrian core of the slate belt against these low-density bodies. This compression led to the lateral tectonic escape of the Cambrian core along dextral transpressional shear zones. These shear zones are interpreted as Palaeozoic NW–SE strike-slip faults as they coincide with several of the NW–SE trending aeromagnetic gradient lineaments shown in Fig. 2c and are also seen as Bouguer gravity gradients (dashed lines in Fig. 2b) (Everaerts and De Vos, 2012; Sintubin, 1999; Sintubin and Everaerts, 2002). Detailed stratigraphic and structural work, however, revealed that these shear zones