

## Reply to comment on 'Tectono-sedimentary evolution of lower to middle Miocene half-graben basins related to an extensional detachment fault (western Crete, Greece)'

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In their comment, van Hinsbergen *et al.* (2008) question the validity of the tectono-sedimentary model proposed by Seidel *et al.* (2007) for the Topolia type breccia deposits in western Crete. Apart from various sedimentological aspects that could probably be sorted out in the field, they argue that the proposed time brackets of *c.* 20–15 Ma for the formation of the Topolia breccia do not reflect the onset of extensional deformation in the overriding plate of the Hellenic subduction zone after collision. Instead, they propose that the onset of extensional deformation is reflected by the formation of a late Miocene (*c.* 12–11 Ma) sedimentary basin unconformably overlying the Topolia type breccia deposits. Notwithstanding the fact that the age of the breccia is otherwise not well constrained apart from being deposited prior to the exposure of the high-pressure metamorphic units in western Crete, we do not concur with this view. Whatever van Hinsbergen *et al.* (2008) mean with terms 'syn-orogenic' and 'post-orogenic' in the present context, they claim that the 'onset of crustal scale, post-orogenic N–S extension' is dated by the oldest Cretan supradetachment basin. They also state that this basin formed 'well after the deposition of the oldest Neogene sedimentary unit in western Crete – the Topolia breccia'. First, it should be noted that Crete represents a narrow horst structure developed in the forearc of the active Hellenic subduction zone since the Pliocene. When dealing with 'crustal scale' processes, the highly incomplete Neogene sedimentary record on the island is

just one limited source of information. Moreover, the buoyant escape model proposed by Thomson *et al.* (1998, 1999) should not be misunderstood to invoke 'extrusion from a subduction channel', as quoted by van Hinsbergen *et al.* (2008), which may rather be a potential cause of the marked uplift of Crete since the Pliocene (Meier *et al.*, 2007). In the buoyant escape model, the partly subducted microcontinent is proposed to return towards the Earth surface by filling the space created by continuous roll back as a more or less coherent slice, becoming exhumed beneath an extensional detachment fault (e.g. Fassoulas *et al.*, 1994; Kiliyas *et al.*, 1994; Jolivet *et al.*, 1996). This process probably causes extensional deformation in the hangingwall to the detachment, which is represented by the unmetamorphic upper units on Crete; extensional deformation continues because of roll back into the present. Any record of normal faulting in the upper plate to the extensional detachment does not allow an unequivocal distinction to be made between the buoyant escape stage (Thomson *et al.*, 1999) and a subsequent stage of more coaxial crustal extension (Thomson *et al.*, 1999; Rahl *et al.*, 2005), stages possibly referred to as 'syn-orogenic' and 'post-orogenic' by van Hinsbergen *et al.* (2008). The structural record of the high-pressure metamorphic Phyllite–Quartzite Unit shows an episode of intense extensional deformation in the brittle field (e.g. Seidel *et al.*, 2005), which took place between *c.* 20 and 15 Ma, according to fission track thermochronometry (Thomson *et al.*, 1998, 1999) and at a depth of less than about 10 km (Küster and Stöckhert, 1997). This extensional deformation therefore happened after the major part of rapid exhumation

from more than 30 km depth, when the high-pressure metamorphic Phyllite–Quartzite Unit already resided in the upper crust. We suspect that this marked extensional deformation of the Phyllite–Quartzite Unit should have had dramatic effects also in the higher levels of the upper crust and consequently at the surface. We therefore envisage that the formation of fault scarps, which are required to produce the fault-bound proximal breccia deposits with a minimum thickness of several hundred meters, preserved in western Crete, could be contemporaneous with the buoyant escape stage as well as with this later stage of extensional deformation. The breccia deposits, as all other units of the upper plate to the extensional detachment, were juxtaposed with the high-pressure metamorphic Phyllite–Quartzite Unit along normal faults during ongoing extension. In our simple structural model (Seidel *et al.*, 2007), we propose that the normal fault presently bounding the Topolia breccia towards the south could have evolved by progressive deformation (or by repeated reactivation) from the fault along which the breccia was deposited on the hangingwall block, the rocks of the Phyllite–Quartzite unit forming the footwall block being juxtaposed with the breccia at a late stage of deformation. The proposed half-graben geometry corresponds to that observed in other extensional tectonic settings (e.g. Davis and Coney, 1979; Wernicke, 1981; Davis *et al.*, 1986) with a more complete structural record and is motivated by the clast size and roundness trends observed in the Topolia area. We feel that the interpretation in terms of a major half-graben bounding fault with progressive exhumation of the footwall block is consistent with

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the record, but other structural interpretations are not excluded. Clearly, we fully agree with van Hinsbergen and Meulenkamp (2006) and van Hinsbergen *et al.* (2008) that the high-pressure metamorphic units were not exposed at the surface at the time of breccia formation, but we do not see an argument to infer a catchment area to the north of the present island. One must consider that the preserved information is highly incomplete for that part of the history and that there were probably quite a number of faults involved. At present, the preserved breccia deposits are bound by faults that were active after breccia formation leaving considerable ambiguity with respect to the original geometry. As such, the isolated breccia occurrence near the harbour of Kastelli mentioned by van Hinsbergen *et al.* (2008) need not be related to the Topolia breccia proper. Moreover, in view of the proximal character of the breccia deposits, the shift of the catchment area to north of the present island, proposed by van Hinsbergen *et al.* (2008), would not solve the problem, apart from being at odds with the observations on clast size and roundness trends in the Topolia breccia. The interpretation of the Topolia breccia as 'a single alluvial to fluvial basin, with a sediment source north of the present-day island', invoked by van Hinsbergen *et al.* (2008) as an alternative to the local-source half-graben scheme proposed by Seidel *et al.* (2007), is difficult to accept for proximal breccia deposits hundreds of meters in thickness. But even with a catchment area off the present-day island, the breccia deposits would require intense tectonic activity, much more pronounced than the comparatively minor deformation inherent in the formation of the later Miocene basins mentioned by van Hinsbergen *et al.* (2008), unconformably overlying the Topolia type breccia deposits. As such, we cannot follow the conclusion drawn by van Hinsbergen *et al.* (2008), who propose that the 12–11 Ma late Miocene basins can be taken to mark the onset of crustal extension. Clearly, marked disintegrated

of the upper crust in the hanging-wall of the extensional detachment must have started earlier and based on the correlation with the late structural record and thermochronometry of the underlying (and now exposed) high-pressure metamorphic Phyllite–Quartzite Unit (Thomson *et al.*, 1999), probably already in the time span between *c.* 20 and 15 Ma, without a possibility to distinguish between a 'syn-orogenic' and a 'post-orogenic' crustal scale deformation in the sense of van Hinsbergen *et al.* (2008).

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