Geological record of the transition from induced to self-sustained subduction in the Oman Mountains

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A R T I C L E I N F O

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A B S T R A C T

Along convergent plate boundaries, the negative buoyancy of the lithosphere pulls the slab into subduction. Bending and offscraping of the downgoing plate are processes occurring at subduction zones and acting against plate motions. These localised dissipative processes cause extensional deformation in the bulge-foredeep region and thrusting and folding in the thrust wedge respectively. Within this framework, widespread early subduction-related extensional structures affecting pre-orogenic rocks of the downgoing plate of fossil subduction systems, are commonly interpreted as induced by extension occurring in the forebulge-foredeep zone. Slab pull is to date, rarely considered as a potential causative process when interpreting basin-scale pre-shortening extensional structures. The problem of distinguishing slab-pull and foreland flexuring induced extensional structures relates to the fact that for most belts, slab pull and forebulge-foredeep flexuring are expected to produce extension roughly in the same direction (i.e. parallel to the foredeep-belt system) and, when syn-kinematic strata are not available, discriminating between these two processes is arduous. In this work we present a field investigation of basin-scale extensional faults from the downgoing plate of the Oman Mts. fossil subduction system. Syn-kinematic strata indicate that normal fault development largely predated extension in the bulge-foredeep region. Herein, we argue that such faulting occurred during the transition from induced to self-sustained subduction, when the negative buoyancy of the slab started to exceed the resisting forces and the downgoing plate began to be pulled towards the trench.

1. Introduction

Sinking of the lithosphere into the asthenosphere due to its negative buoyancy is widely recognised as the major driving force of plate tectonics (e.g. Forsyth and Uyedaf, 1975; Conrad and Lithgow-Bertelloni, 2002). Elastic resistance to bending (e.g. Ranero et al., 2003) and viscous resistance of the asthenosphere (e.g. Schellart, 2004) inhibit subduction, particularly during its infancy (McKenzie, 1977; Gurnis et al., 2004). Forced convergence between opposing plates is typically required to induce subduction, before it eventually becomes a self-sustained process (Stern, 2004). The transition from induced to self-sustained subduction typically occurs after a certain volume of oceanic lithosphere has been forced to enter the mantle. At this tipping point, the negative buoyancy of the slab exceeds elastic resistance to lithospheric bending and the viscous resistance of the asthenosphere, such that the downgoing plate is pulled toward the trench (e.g. Faccenna et al., 1999; Hall et al., 2003). This transition produces a major reorganization of forces applied to the edge of the downgoing plate entering into the asthenosphere, with the resulting force applied to the edge of the plate pushing the slab backward and pulling it forward (relative to the trench) during induced and self-sustained subduction respectively (Fig. 1). Once the slab pull starts to exceed the resisting forces, an extensional state of stress can be established in the downgoing plate (e.g. Spence, 1987). As the slab pull force increases, the extensional state of stress affects progressively more distal areas of the unsubducted portion of the downgoing plate, with respect to the location of the trench (Fig. 1). Recognising indicators of stress variation in the downgoing plate at the time of subduction initiation is therefore of paramount importance in plate tectonic studies, as it allows one to determine: (i) whether a subduction system was initiated by induced or spontaneous subduction and (ii) whether transition between the two subduction modes has occurred. Establishing such observations is however, rather challenging, as the effects of the aforementioned transition cannot be seen in the oceanic lithosphere of the downgoing

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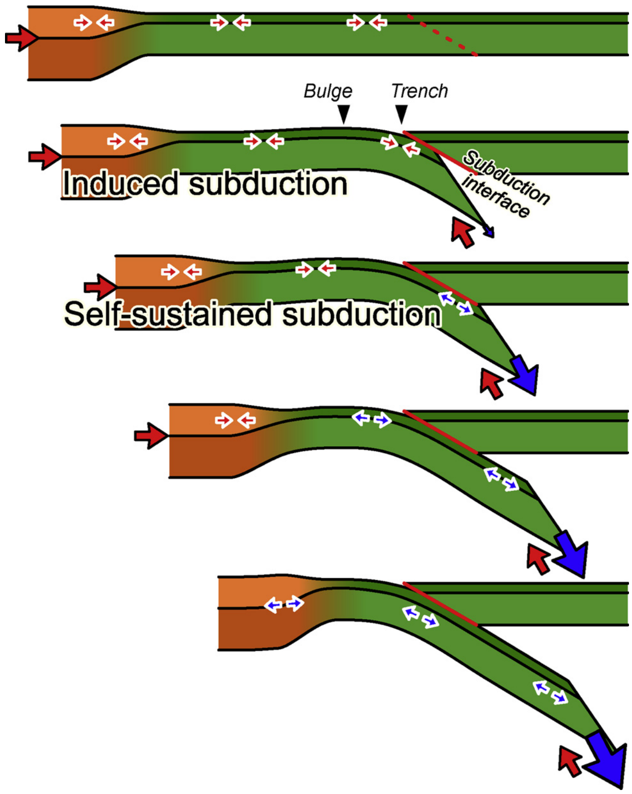

Fig. 1. Scheme showing the transition from induced to self-sustained subduction. Double arrows indicate compressional (red) and extensional (blue) states of stress. Single arrows show forces playing in favor (blue) and against (red) subduction.

plate, which is either subducted or metamorphosed prior to its final exhumation at the surface. Conversely, the continental lithosphere has much greater preservation potential than the oceanic lithosphere in subduction settings. However, the effects of slab pull initiation on the continental lithosphere are largely occluded, due to the absence of a significant syn-tectonic sedimentary record, or masked by younger extensional events, such as stretching induced by arching of the lithosphere in the bulge-foredeep region (Bradley et al., 1991; Lash and Engelder, 2007; Lacombe et al., 2009; Tavani et al., 2015a; Granado et al., 2016). Recognising the slab pull effect thus requires the reconstruction of the chronology of extension relative to development of the forebulge and thrusting in the foredeep. This in turn requires the study of a subduction system in which the age of subduction initiation is well constrained. At present, indeed, only a few structural works have documented the association between extensional structures and slab pull (e.g. Capitanio et al., 2009).

The slab pull effect could be invoked to explain some poorly understood Cretaceous extensional structures developed at the northern margin of Gondwana. There, mid to Late Cretaceous basin-scale extensional structures are documented in the Sirte basin of Lybia (Roohi, 1996; Frizon De Lamotte et al., 2011), in Tunisia (Patriat et al., 2003), in the Adria region (Shiner et al., 2004; Tavani et al., 2015b; Vitale et al., 2018), and in the Arabian margin presently exposed in the Zagros (Wrobel-Daveau et al., 2010). These structures developed coevally with intra-oceanic neo Tethyan subduction occurring immediately to the NE (i.e. the Izmir-Ankara and Oman subduction systems: Jolivet et al., 2016 and references therein), and are consistent with a slab-pull effect. However, accurate timing of processes operating during subduction infancy, and in particular of the transition from induced to self-sustained subduction, is available only for the Oman collisional system. In this work we investigate Cretaceous extensional structures from the Oman Mts. fossil subduction system, to determine whether they developed coevally with slab pull initiation. Kilometre-long extensional faults exposed along the former continental margin of the downgoing

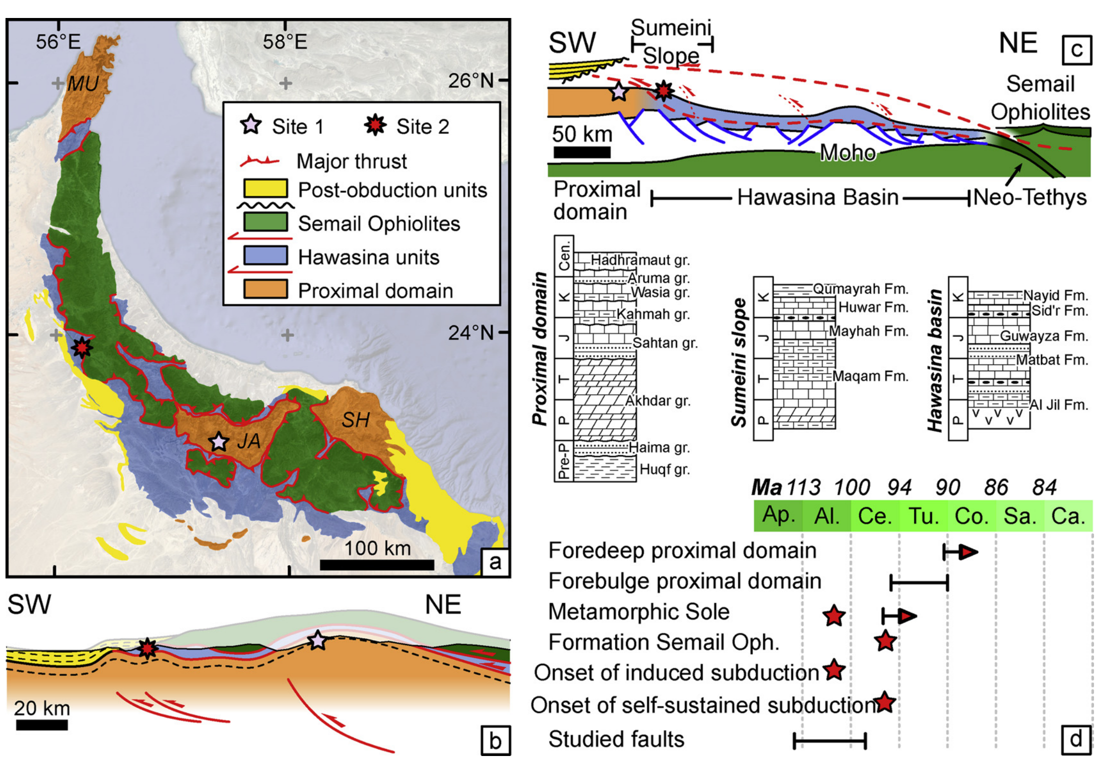

Fig. 2. Geological framework. (A) Simplified geological map of the Oman Belt. (B) Schematic geological section across the central portion of the belt. (C) Diagram showing the early-orogenic framework along with simplified stratigraphic successions of the different Mesozoic domains. (D) Time-table showing the main tectonic events and their timing. Ages for the metamorphic sole are from Guilmette et al., 2018 (star symbol) and from Rioux et al., 2016 (arrow).

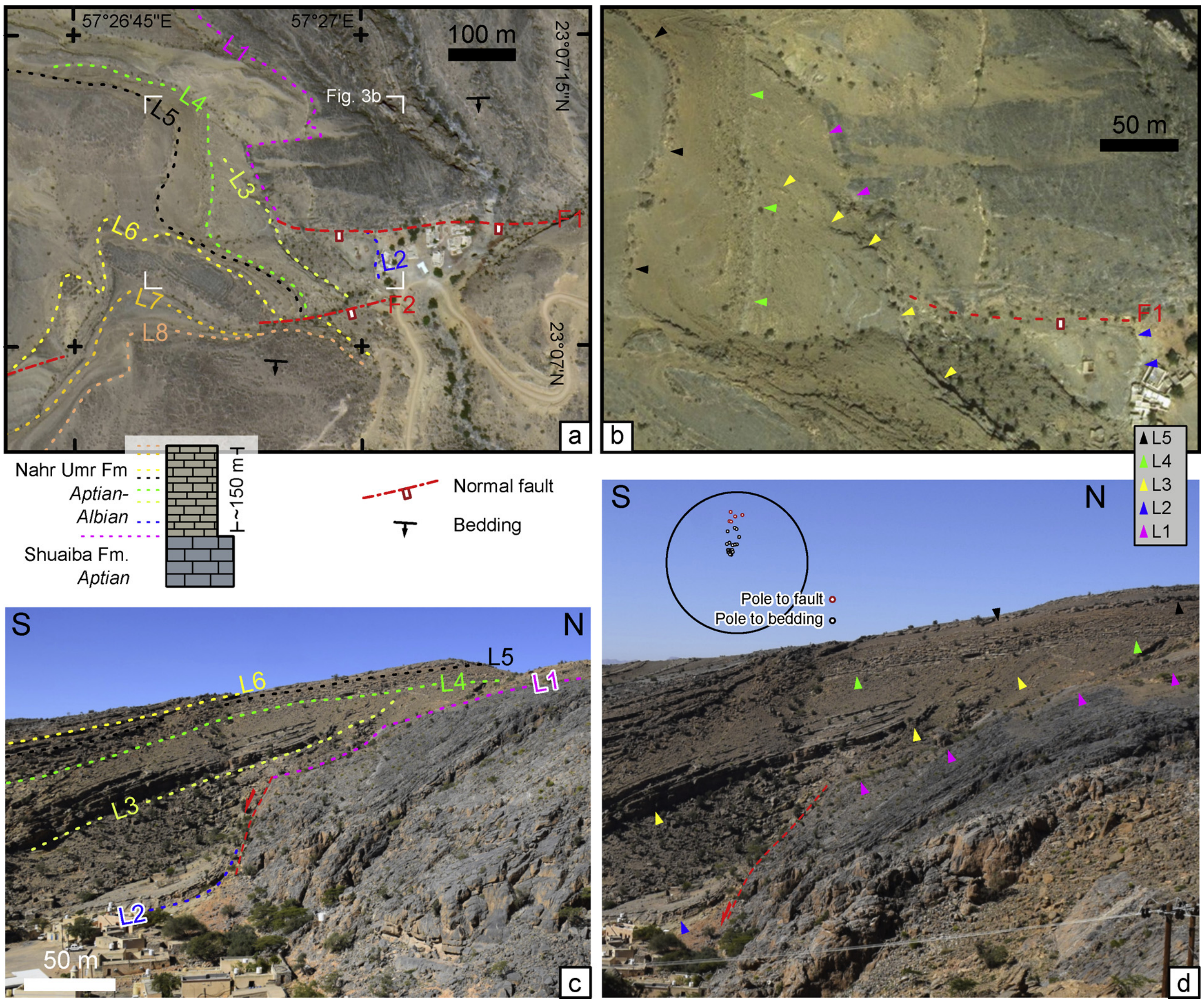

Fig. 3. Extensional faults at Site 1. (A) Orthophoto with faults and key layers indicated, with detail (B) showing the continuity of layers L3, L4, and L3, indicating that these seal the fault F1. (C&D) Views from the east of the northern fault with lower hemisphere, equal-area projection of poles to faults and bedding surfaces.

plate are characterised by means of structural observations and remote sensing, to determine their timing with respect to the major pre- and syn-orogenic events. This allowed us to discard mechanisms other than slab pull as the causative process in their development.

2. Geological setting

The NW-SE trending Oman belt (Al Hajar Mountains) is located on the eastern edge of Arabia (Fig. 2a). It consists of a stack of SW-directed nappes formed during Late Cretaceous SW-directed ophiolite obduction, and was later re-deformed by Cenozoic reactivation (Fig. 2b) (Glennie et al., 1973; Glennie and Reinhardt, 1974; Searle and Malpas, 1980; Searle and Ali, 2009). The Semail ophiolites occupy the uppermost portion of the tectonic pile which overthrusted the distal portion of the Mesozoic Arabian continental margin (Fig. 2b,c), with a thin (< 1 km thick) metamorphic sole separating these two domains. Differentiation of the Arabian continental margin into proximal and distal domains (characterised by shallow-water and deep-water depositional environments respectively), occurred during Permo-Triassic rifting (Robertson and Searle, 1990; Bechennec et al., 1990; Chauvet et al., 2009). This is evidenced by the different Permo-Triassic aged facies of the diverse domains, and by the occurrence of Permian volcanites underlying the Mesozoic stratigraphic succession of the distal domain (Fig. 2c). The Hawasina basin represents an example of such a distal domain. The Hawasina Basin is constituted by ridges and sub-basins and was originally separated from the proximal domain of the margin by the Sumeini Slope (which is grouped together with the Hawasina units in Fig. 2a). The basement of the Hawasina Basin is not known in detail. However, the occurrence of enriched MORB-type pillow basalts in the Permian of the Hawasina Basin (Pillevuit et al., 1997; Lapierre et al., 2004) and the subduction of hundreds of kilometres of the Hawasina Basin's substratum (Cooper, 1988; Béchennec et al., 1990) suggest that many sectors of this basement could have been oceanic in nature (Robertson and Searle, 1990). During Late Cretaceous obduction, the Permo-Mesozoic tectono-sedimentary covers of the different portions of the Hawasina Basin and of the Sumeini slope were detached from their basement and thrusted SW-ward, on top of the proximal domain of the Mesozoic Arabian continental margin (Glennie et al., 1973; Béchennec et al., 1988; Cooper, 1988; Breton et al., 2004; Searle and Ali, 2009). This proximal domain is exposed in the Musandam Peninsula, in the Jabal Akhdar and the Saih Hatat culminations (MU, JA, and SH in Fig. 2a). Lower Maastrichtian to Oligocene conglomerates, shales, and shallow-marine limestones unconformably overlie the tectonic pile (Fig. 2b). These sediments were deposited in a tectonically quiescent period, and were later deformed, along with the Cretaceous belt, during the latest Eocene to Miocene shortening event (Corradetti et al., 2019).

Convergence in Oman initiated with an intra-oceanic subduction (Robertson and Searle, 1990; Searle et al., 2014; Van Hinsbergen et al., 2015; Jolivet et al., 2016) (Fig. 2d). After subduction initiation, at 96-

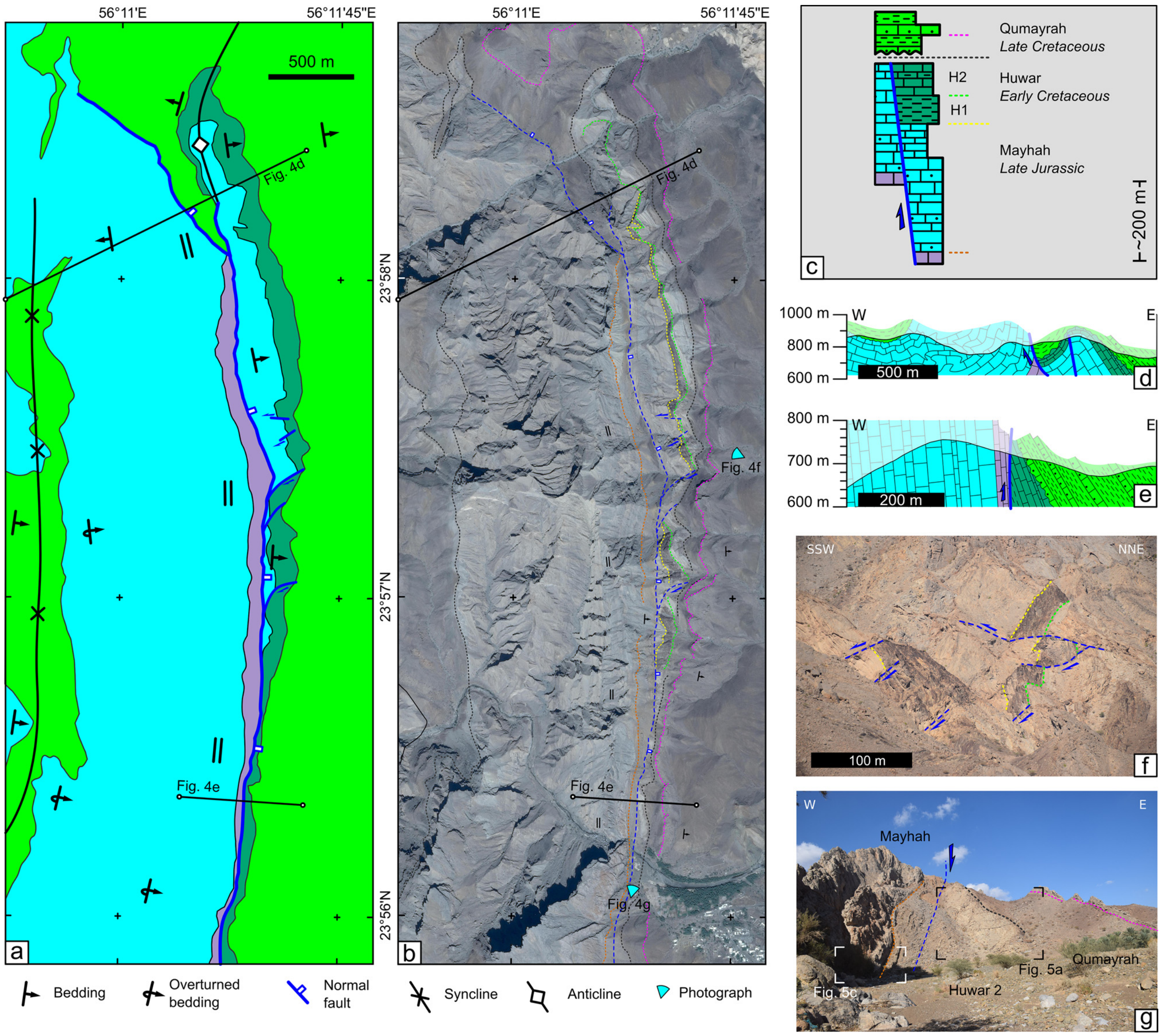

Fig. 4. Geological setting of the Qumayrah village area, where the Site 2 is located. (A) Geological map and (B) orthophoto with faults and bedding indicated (modified from Cooper et al., 2012). (C) Stratigraphic succession of the area. (D&E) Geological cross-sections (trace is in Figure 4a&b). (F) Detail of the radiolarian member of the Huwar Fm. (dark lithology) evidencing tilted normal faults. (G) View from the South of the Qumayrah Fault.

95 Ma, the Semail ophiolites were generated in a supra-subduction environment (e.g. Lippard et al., 1986; Rioux et al., 2012, 2016). Amphibolite facies metamorphism affected the metamorphic sole at the base of the peridotite at 95-92 Ma (54-46 km depth, Searle and Cox, 2002) (Fig. 2d): i.e. coevally with the Semail ophiolite production. Different ages have been documented for mineral assemblages of the metamorphic sole, with U-Pb zircon ages (Rioux et al., 2016 and references therein) indicating ages ranging between 96.5 and 95.5 Ma (with single zircon dates precision of $2\sigma \geq \pm 0.3$Ma mainly from garnet amphibolites and trondhjemite and diorite pods), and Lu-Hf on garnets (from garnet- and clinopyroxene-bearing amphibolites) indicating ages of 103.7 $\pm$ 0.7 Ma (Guilmette et al., 2018). This has been attributed to an initial stage of induced subduction occurred before 104 Ma, followed by a self- sustained subduction stage coeval with back-arc extension and Semail ophiolites production (Guilmette et al., 2018). According to this scenario, the onset of self- sustained subduction occurred between 104 and 96 Ma (Guilmette et al., 2018). The latter stages of ophiolite emplacement over the proximal domain of the Arabian Margin are recorded by eclogite-facies metamorphism in the Saih Hatat culmination at 79 Ma (e.g. Searle et al., 1994; Warren et al., 2003). During obduction, forelandward (i.e. SW-ward) migration of the wedge caused the migration of the foredeep-forebulge system. In the proximal domain of the continental margin, presently exposed in the Jabal Akhdar and Saih Hatat culminations, and in the Musandam Peninsula, this produced uplift of the bulge, recorded in the Jabal Akhdar culmination and in the frontal part of the belt (Jabal Salakh range) by a Turonian erosional episode known as Wasia-Aruma Break (Boote et al., 1990; Robertson and Searle, 1990; Cooper et al., 2014). Uplift was followed by the onset of foredeep sedimentation during the uppermost Turonian-Coniacian (Fig. 2d) (Muti Fm.; Glennie et al., 1973; Robertson, 1987a, b; Searle, 2007). In contrast to the above, the onset of bulging in the Hawasina basin and in the Sumeini slope is poorly constrained. However, the proximity between the Sumeini slope and the proximal domain suggests a Turonian-Coniacian age for the peripheral bulge stage in the Sumeini slope too.

3. Field observations

The first exposure is located on the SW flank of the Jabal Akhdar culmination (Fig. 2a), where the stratigraphic sequence of the proximal

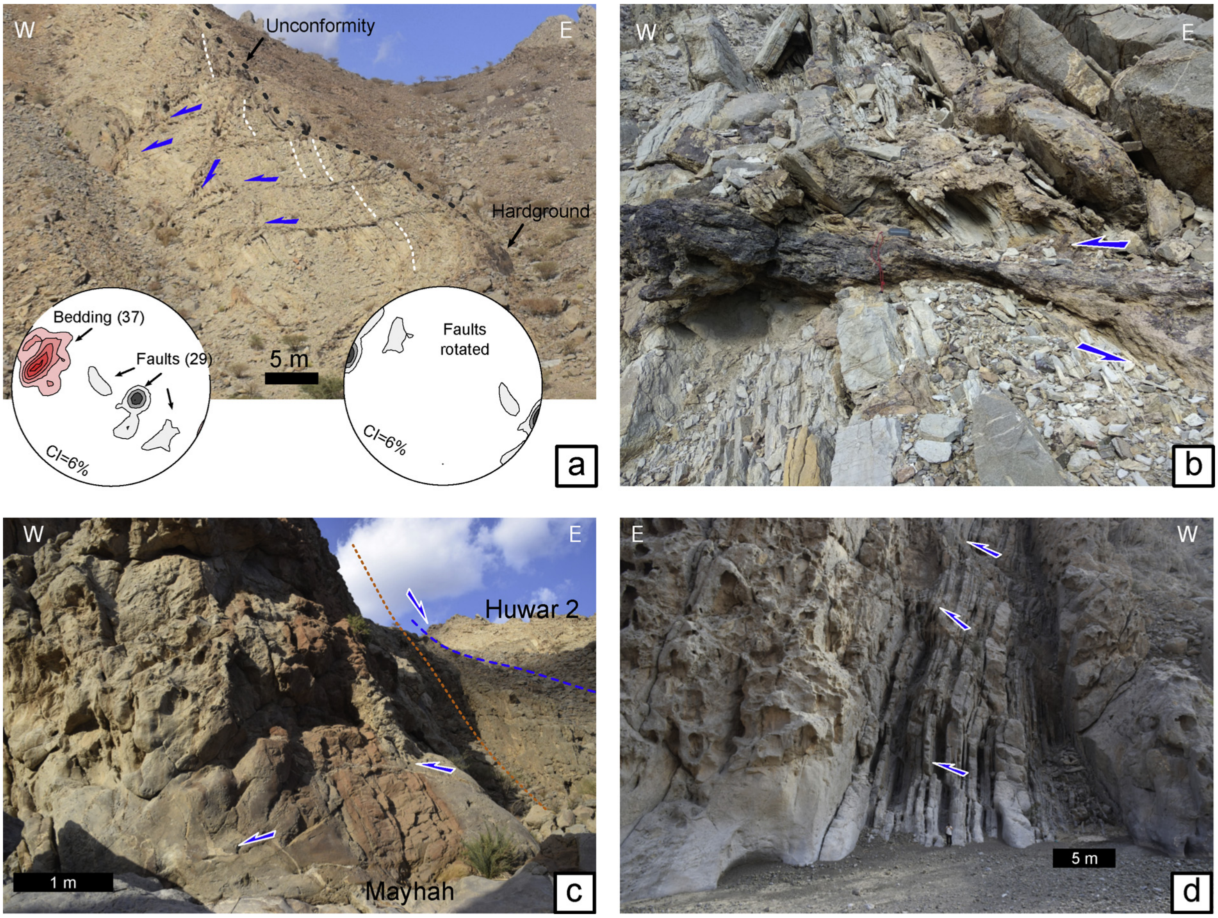

Fig. 5. Details of the extensional fault system at Site 2. (A) Detail showing tilted and faulted Huwar H2 member unconformably overlain by shales of the Qumayrah Fm., with lower hemisphere, equal-area projection of contoured poles to planes of faults (grey-scale) and bedding surfaces (red-scale) in their present-day orientation, and (left) faults after bedding dip removal. (B) Detail of a tilted normal fault seen in Fig. 5a. The 20-cm thick fault zone is altered by iron oxides concretions. (C) Tilted normal faults in Mayhah Fm., in the footwall of the Qumayrah Fault, immediately below the main extensional fault. (D) Tilted normal fault in Mayhah Fm., in the footwall of the Qumayrah Fault (40 m west of the Qumayrah Fault). Notably, meso-scale thrust-related structures (such as S-C or tilted reverse meso-faults) do not occur in (C) or (D), pointing to the absence of significant thrusts within the immediate vicinity of the Qumayrah Fault.

domain of the margin is exposed. There, kilometre long WNW-ESE to E-W striking normal faults affect the Paleozoic to Cretaceous sedimentary succession. These faults are associated with a long lasting extensional history that includes a post-obduction Campanian-Maastrichtian extensional event (Grobe et al., 2018). Some of these faults, however, appear to be older, as they are sealed by the Muti Fm. (Beurrier et al., 1986; Béchennec et al., 1992), whose base is ascribed to the Turonian-Coniacian interval based on biostratigraphy (Robertson, 1987a, b). The Wasia Group includes two formations, namely the Natih shelf carbonates and the underlying Nahr Umr intra-shelf basin with abundant terrigenous input from the Arabian shield. The Wasia Group, which unconformably underlies the Muti Fm., displays growth geometries associated with these extensional faults within the Nahr Umr Fm. This is the case for the fault system illustrated in Fig. 3, where two S-dipping faults affect gently S-dipping strata. Eight key layers are individuated within the Aptian-Albian age carbonates exposed around the fault system (Fig. 3a&b), with Layer 1 corresponding to the top of the Aptian Shuaiba Fm., and layers 2 to 8 within the latest Aptian to Albian Nahr Umr Fm. Layer 8 unconformably overlies Layer 7, with both forming the post-kinematic succession sealing the southern fault (F2). The age of both the Shuaiba and Nahr Umr formations is constrained by biostratigraphy (Witt and Gökdag, 1994; Immenhauser et al., 1999). The post-kinematic succession of the northern fault (F1) instead, includes Layer 3 to Layer 8, with Layer 1 being pre-kinematic and Layer 4 unconformably overlying Layer 3, as illustrated in Figs. 3c&d.

The second exposure is located in the Jabal Qumayrah area (Figs. 2a, 4a), where a tilted km-long extensional fault, hereinafter named the Qumayrah Fault, affects strata of the Mayhah (Early – Late Jurassic, Oolitic calcarenites, fine-graned calcirudites, often dolomitized), Huwar (Late Tithonian – Cenomanian; silicified limestones and cherts), and Qumayrah (Cenomanian – Coniacian?; cherts, massive and coarse-grained calcirudite, shales and calcarenite) formations of the Sumeini slope (Fig. 4a). The age of these formations is constrained by biostratigraphy and chemostratigraphy (Watts and Garrison, 1986; Watts and Blome, 1990; Wohlwend et al., 2017). The Qumayrah Fault is a major N-S striking structure whose trace can be followed for more than 5 km along its strike, before being sealed by post-kinematic Cenomanian strata (Fig. 4a&b). The Qumayrah Fault divides a western footwall block, in which the Late Cretaceous Qumayrah Fm. unconformably rests on top of the Late Jurassic Mayhah Fm., from an eastern hanging wall block, where the Qumayrah Fm. overlies the Early Cretaceous Huwar Fm. (Fig. 4c). The fault has been partly deformed during shortening, and strata of the footwall are upright due to buttressing (Fig. 4a, d, e). The hanging wall block has been slightly deformed too, as indicated by the occurrence of an anticline at the northern edge of the fault (Fig. 4d), pointing to a limited positive reactivation. The geometry of the hanging wall of the Qumayrah Fault in map view is well illustrated in Fig. 4b, where key horizons are indicated. In detail, the Huwar Fm. is further divided into the late Tithonian to upper Valanginian-Hauterivian Huwar 1 member – a dark radiolarian chert interval representing a key marker horizon – and the Valanginian-Hauterivian to Albian Huwar 2 member, which consists of

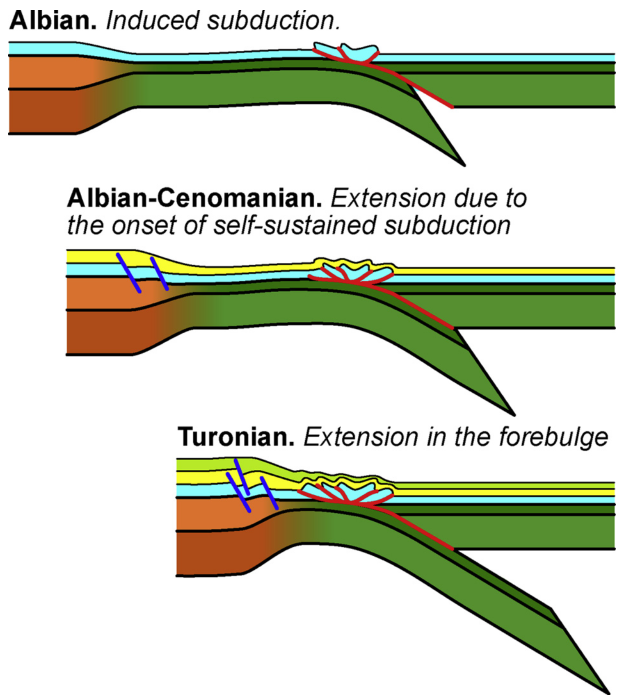

Fig. 6. Schematic reconstruction illustrating the relationships between deformation and sedimentation recorded in the continental margin of the downgoing plate.

micritic limestones and calcarenites (Watts and Blome, 1990; Wohlwend et al., 2017). The Huwar is capped by the Qumayrah Fm., which includes a distinctive calcarenitic level (pink level in Figs. 4b&c). Several splays emanate from the main fault and affect the Huwar Fm. (Fig. 4b), which are clearly sealed by the Qumayrah Fm: this truncation is particularly evident within the' pink 'calcarenitic level of the Qumayrah Fm (Fig. 4b). The Tithonian to upper Valanginian-Hauterivian Huwar 1 radiolarian chert interval holds a pervasive network of tilted extensional faults (Fig. 4). These are observed in the overlying Huwar 2 member, but are not in the Late Cretaceous Qumayrah Fm. The cross sectional geometry of the hanging wall of the Qumayrah extensional fault can be observed west of the Qumayrah village (Fig. 4g), where the fault is near vertical. The 50 to 100 m thick Huwar 1 member is not exposed in this location and an unconformity marked by a hardground separates the Huwar 2 and the base of the Qumayrah Fm. (Fig. 5a). This latter is ascribed to the base of the Cenomanian (Watts and Blome, 1990), placing the age of the fault activity at the Albian to lowermost Cenomanian interval. The strata below the unconformity are affected by faults (Fig. 5a&b) marked by iron oxide concretions (Fig. 5b), which do not affect the beds above the unconformity. These faults are arranged in three NNE-SSE striking sets, two of them having cut-off angles of 60° in opposite directions, thus forming a tilted extensional system (Fig. 5a). The third set includes bedding-perpendicular segments, probably resulting from the linkage of precursor joints during fault growth (Healy et al., 2006). This orientation is only slightly oblique to the view direction of the photo in Fig. 5a, indicating that the trace of the fault and its splays, is highly oblique to the slip direction (i.e. top-to-ESE) (see Fig. 4a). Moreover, the pre-kinematic strata of the Mayhah Fm. immediately to the west of the extensional fault (i.e. in the footwall of the Qumayrah Fault) are affected by low-displacement, tilted extensional faults (Fig. 5c&d), thus providing additional evidence of pre-folding extensional tectonics.

4. Discussion and conclusions

Despite the strong overprint associated with the emplacement of the Semail ophiolites and subsequent syn- to late-orogenic processes (Gregory et al., 1998; Fournier et al., 2006; Tarapoanca et al., 2010; Cooper et al., 2014; Storti et al., 2015; Aldega et al., 2017; Grobe et al., 2018; Hansman et al., 2018), extensional structures predating thrusting and folding are still recognisable in the Oman Mts. The two kilometre-long fault systems described in this work formed in the Albian-lowermost Cenomanian interval, during which time no significant rifting occurred in the area. Furthermore, major submarine landslides or accumulation structures of Albian-lowermost Cenomanian age are unknown. Consequently the observed extension should be attributed neither to rifting (or rifting renewal) nor to gravitational sliding at the continental margin. The striking absence of pre-kinematic radiolarian chert of the Huwar 1 member in the footwall of the Qumayrah Fault further precludes the gravitational sliding as a mode of fault genesis. This absence indicates a syn-kinematic erosion of the footwall, which does not fit with gravitational sliding. Additional deformation mechanisms operating during the typical evolution of passive margins, such as differential compaction of the substratum cannot be invoked either. Though such processes can produce fracturing at the meso-scale, they do not lead to the development of kilometre-long extensional faults (Tavani et al., 2018). Extension due to bulging of the lithosphere (e.g. Ranero et al., 2003; Tavani et al., 2015a; Balestra et al., 2019) is to be excluded too, as the bulge reached the current study areas not earlier than the late Turonian. Indeed, Turonian uplift and erosion in the Jabal Akhdar proximal domain, followed by the onset of foredeep sedimentation, has been classically interpreted as an evidence of late Turonian to Coniacian bulging (e.g. Robertson, 1987; Searle, 2007; Cooper et al., 2014). This is in agreement with the fact that the Semail ophiolites overthrusted the distal domains of the Hawasina basin, originally placed nearly 200 km to the NE of the Proximal Domain/Sumeini slope (Cornish and Searle, 2017), not earlier than the late Turonian, as indicated by the age of the topmost pre-obduction formations of these domains (Bechennec et al, 1990). In addition, at the time of extensional faulting the Semail ophiolites were not yet developed (Fig. 2d), meaning that there was still a considerable amount of oceanic crust associated with the downgoing plate placed in between the trench and the Hawasina basin. Typical trench-bulge distances for subduction zones involving oceanic slabs are in the order of several tens of kilometres (e.g. McNutt and Menard, 1982). It is therefore unrealistic to assume a forebulge in the proximal domain when incipient subduction of oceanic lithosphere was occurring hundreds of kilometres away. Having ruled out other reasonable causes (including local instabilities at the passive margin), the timing of the described extensional structures fits the hypothesis that they developed in response to the onset of slab pull. The Albian-lowermost Cenomanian age of the extensional faults overlaps with the age of subduction initiation, but predates the formation of the Semail supra-subduction ophiolites (Fig. 2d). Therefore these normal faults apparently developed before the transition from induced to self-sustained subduction in the Oman Mts. (Fig. 2d). However, the fact that slab-pull-induced extension in the downgoing plate predates extension in the upper plate is not surprising, as the onset of self-sustained subduction produces a change in the balance of forces, and its effect is expected to be almost immediately transmitted to the downgoing plate interior. Conversely, the onset of self-sustained subduction triggers slab sinking and hinge retreat, which in turn produces extension in the upper plate (e.g. Guilmette et al., 2018), resulting in a delay between the onset of self-sustained subduction and the first evidence of upper plate extension.

Extensional faults from the proximal domain of the Arabian margin indicate that once slab pull starts, it is not instantaneously transmitted from the slab edge to the entire plate, producing a general acceleration of plate convergent motion. Rather, the onset of slab pull determines the establishment of an extensional stress field that can produce extensional failure (Fig. 6), before extension associated with flexuring of the lithosphere in the forebulge. Slab pull produces extension that, for linear trenches, operates prior to, but also in a direction identical to extension produced by flexure of the lithosphere at the forebulge.

Our results provide a new, effective and viable mechanism for the development of extensional structures predating thrusting and folding, which are documented worldwide in foreland fold and thrust belts (Tavani et al., 2015a; and references therein). Based on our observations, extension associated with slab pull initiation should be taken into account, in addition to well-known forebulge-foredeep–related extensional structures, when interpreting pre-orogenic normal fault-fracture networks in fold and thrust belts.

Declaration of Competing Interest

No conflict of interest.

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