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Sequence of thrusting and syntectonic sedimentation in the eastern Sub-Atlas thrust belt (Dadès and Mgoun valleys, Morocco)

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Abstract A series of balanced cross-sections across the Sub-Atlas thrust belt and the northern Ouarzazate basin are used to illustrate the structural geometry and the timing of deformation at the southern front of the High Atlas Mountains of Morocco. The selected area is among the best sedimentary records of mountain building of the entire orogenic system. The study of the relationships between thrusts and synorogenic continental formations enables the unraveling of kinematic sequences and the proposal of a relative chronology of deformation. Active thrusting in the area occurred in a rather continuous fashion from the Oligocene to the Pliocene, punctuated by a major erosional phase imprecisely placed in late Oligocene to early Miocene times. Detrital sedimentary facies indicate that uplift in the hinterland of the High Atlas, to the north of the Sub-Atlas belt, was taking place already by mid Eocene times, although it might have commenced locally even earlier. Within the Sub-Atlas zone, the exposed faults did not propagate in a simple piggy-back fashion but show evidence of a complex, synchronous sequence with events of fault reactivation and out-of-sequence thrusting.

Keywords Thrust · Tectonics–sedimentation · Cenozoic · High Atlas · Morocco

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Introduction

The Sub-Atlas thrust belt and the Ouarzazate basin are located in the southern border of the High Atlas Mountains of Morocco, and contain perhaps the best sedimentary record of mountain building of the Atlas system. In the Sub-Atlas belt (Fig. 1) pre- and syndeformational sediments are well exposed. Nevertheless, the age of deformation is controversial; different ages and different tectonic pulses have been proposed to account for the compressional deformation in the area, ranging from the late Cretaceous to the Quaternary. Laville et al. (1977), based on local unconformities and relationships between tectonics and sedimentation, proposed a long-lasting deformation from the late Cretaceous to the Miocene. Görler et al. (1988), based on conglomeratic fluxes into the basin, distinguished two tectonic pulses, the first during the Oligocene–early Miocene and the second during the late Pliocene–Pleistocene. Fraissinet et al. (1988), based on the unconformities between the main sedimentary formations, proposed four deformation pulses during the Oligocene–Pliocene interval. El Harfi et al. (1996, 2001), based on the distribution of detrital deposits, suggested two main pulses separated by a period of tectonic quiescence, the first during the late Eocene–Oligocene and the second during the Mio-Pliocene or Pliocene. Recently, Frizon de Lamotte et al. (2000), based on the same criteria, proposed two tectonic pulses, the first during the late Eocene age and the second during the Pleistocene–early Quaternary.

Here, due to this lack of agreement between previous authors, we revise the geometry and the timing of the compressional deformation in the eastern Sub-Atlas zone. The area investigated in detail lies

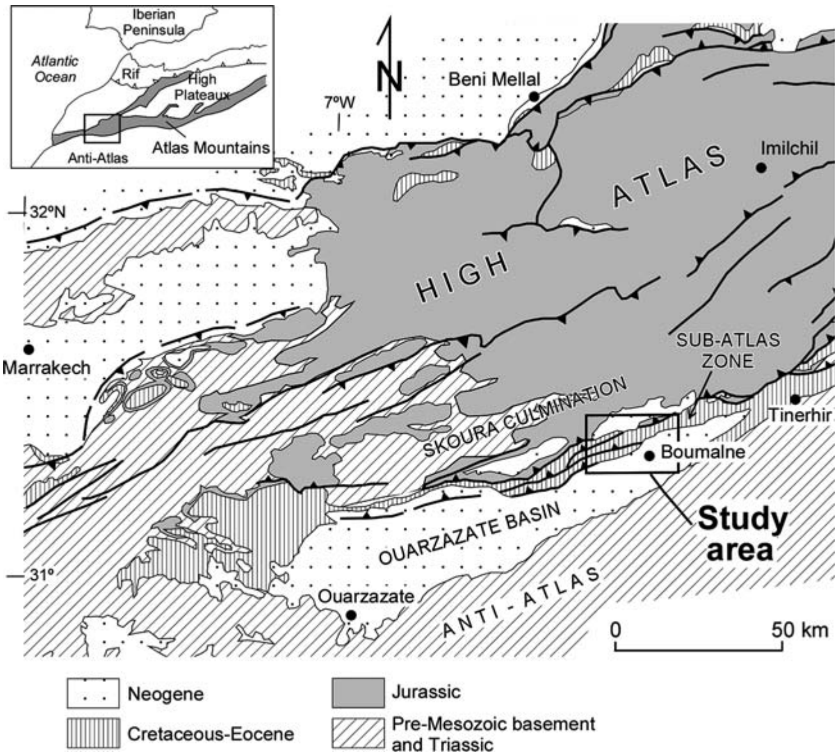


Fig. 1 Structural map of the western part of the central High Atlas of Morocco indicating the location of the study area at the southern margin of the chain (*boxed*)

between the well exposed Dadès and Mgoun valleys where the relations between tectonics and sedimentation can be established, allowing the unraveling of thrust sequences and of the deformation chronology of the Alpine compression at the southern margin of the High Atlas Mountains of Morocco. This study aims to add some chronological constraints on the recent debate on the mechanisms and timing of uplift of the Atlas orogenic system, where crustal shortening and mantle processes have been operating in the past million years (Zeyen et al. 2005; Teixell et al. 2005; Missenard et al. 2006).

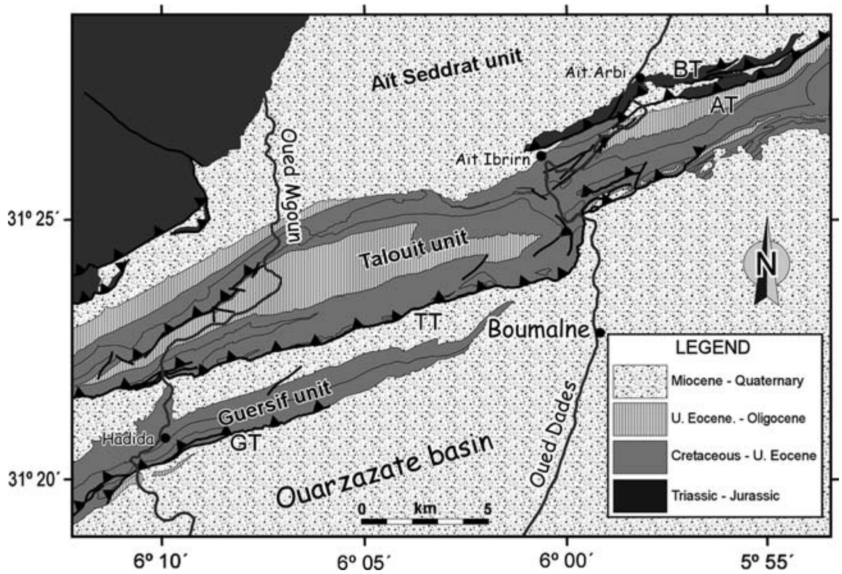
Geological setting: the High Atlas and the Ouarzazate basin

The High Atlas is an intracontinental chain derived from the inversion of a transtensive rift of Triassic to Jurassic age (Mattauer et al. 1977; Jacobshagen 1988). The structural style of the compressive deformation is mainly thick-skinned in the orogen (Frizon de Lamotte et al. 2000; Teixell et al. 2003), whereas in the southern border there is a narrow belt of ENE-trending detached folds and thrusts known as the Sub-Atlas zone (Choubert and Faure-Muret 1962; Laville et al. 1977; Errarhaoui 1997; Beauchamp et al. 1999), where folded Cretaceous to Paleogene rocks override Neogene sediments with a general southward vergence.

The tectonic shortening across the High Atlas has recently been estimated between 15 and 25% (13 to 26 km). It increases from west to east along the strike while the topographic elevation generally decreases. This inverse correlation suggests that crustal thickening does not fully explain the observed topography and suggests a mantle-sourced, thermal contribution to uplift (Teixell et al. 2003), which is corroborated by geophysical data (Seber et al. 1996; Ayarza et al. 2005; Teixell et al. 2005; Zeyen et al. 2005). At present, the nature of tectonic and thermal processes that led to mountain building in the Atlas is well understood, but their chronology is yet to be unravelled.

To the south of the High Atlas, the Ouarzazate basin is a small, elongated basin (145 km long, maximum 35 km across and 800 m deep) of Cenozoic age that constitutes the southern foredeep of the mountain chain (Fig. 1). To the north, the Ouarzazate basin is flanked by the marginal thrust belt of the Sub-Atlas zone, dominated in outcrop by rocks from late Cretaceous to Neogene age (Gauthier 1957) (Figs. 1, 2). To the south, the sedimentary infill of the Ouarzazate basin onlaps the Precambrian of the Anti-Atlas massif, a wide, 100-km-scale arch that plunges beneath the basin. The basin was active and recorded sedimentary aggradation at least from the late Eocene to the

Fig. 2 Structural sketch map of the study area (see Fig. 1 for location), showing the main tectonic units. *GT* Jebel Guersif thrust; *TT* Jebel Talouit thrust; *AT* Algouzi thrust; *BT* Bou Ikhfian thrust



Pliocene (Gauthier 1957; Fraissinet et al. 1988; Görler et al. 1988; El Harfi et al. 2001; Tesón 2005).

Nature of the sedimentary succession

Based on the relationship with the local compressional structures, the sedimentary successions cropping out in the study area have been subdivided into two groups: a predeformational and a syndeformational group. This subdivision is aimed at simplifying description and is only valid for the study area; in fact, the upper predeformational units could be considered as syntectonic at the regional scale of the High Atlas, because they record uplift of distant upstream areas north of the Sub-Atlas zone.

Predeformational succession

The predeformational succession covers the interval from the Triassic to the late Eocene or early Oligocene. The Triassic and Jurassic rocks can be considered as syn-rift deposits and are constituted by red siltstones and sandstones and by massive limestones and dolomites, respectively. Some terrigenous intercalations consisting of red shales, sandstones and minor conglomerates occur at the top of the Liassic succession. The Cretaceous is represented by three characteristic formations: (1) red sandstones and conglomerates of poorly constrained age (''Infracenomanian'' of earlier authors; e.g., Gauthier 1957) that are conformable on Jurassic terrigenous beds or unconformable on the Precambrian and Paleozoic at the borders of the original Mesozoic rift. The conglomerate pebbles are rounded and constituted predominantly of Precambrian and Palaeozoic rocks; (2) a white limestone of Cenomanian–Turonian age (Gauthier 1957; Ettachfini and Andreu 2004), which constitutes a good marker level, and (3) red shales with sparse sandstone and gypsum beds attributed to the Upper Cretaceous (Gauthier 1957; Figs. 3, 4).

The predeformational Paleogene is represented, at the base, by alternating levels of variously colored marls and bioclastic limestones that form the main topographic ridges of the Sub-Atlas zone. The age of this formation is late Maastrichtian to mid Lutetian (Marzoqi and Pascal 2000; Tabuce et al. 2005). The regional distribution of the outcrops of this marine formation in the High and Middle Atlas points to a large epicontinental basin, covering part of the western High and Middle Atlas and the Moroccan Meseta (Herbig and Trappe 1994). Local unconformities and microconglomeratic layers included in this formation (Laville et al. 1977; Froitzheim et al. 1988; Herbig and Trappe 1994) suggest the existence of some ridges of local extent growing within the basin.

On top of the Eocene limestones are red shales with gypsum of the Hadida formation, which pass laterally to sandstones and microconglomerates of the Aït Arbi formation (Gauthier 1957; El Harfi et al. 2001) (Fig. 3). These formations have not been biostratigraphically dated, but with a maximum thickness of 700 m, their conformable disposition on dated Lutetian sediments suggests a mid Eocene to Oligocene age (see also Görler et al. 1988; El Harfi et al. 2001). Even though outcrops of these formations do not show evidence of contemporary deformation, their terrigenous nature and the presence of recycled pebbles from the lower

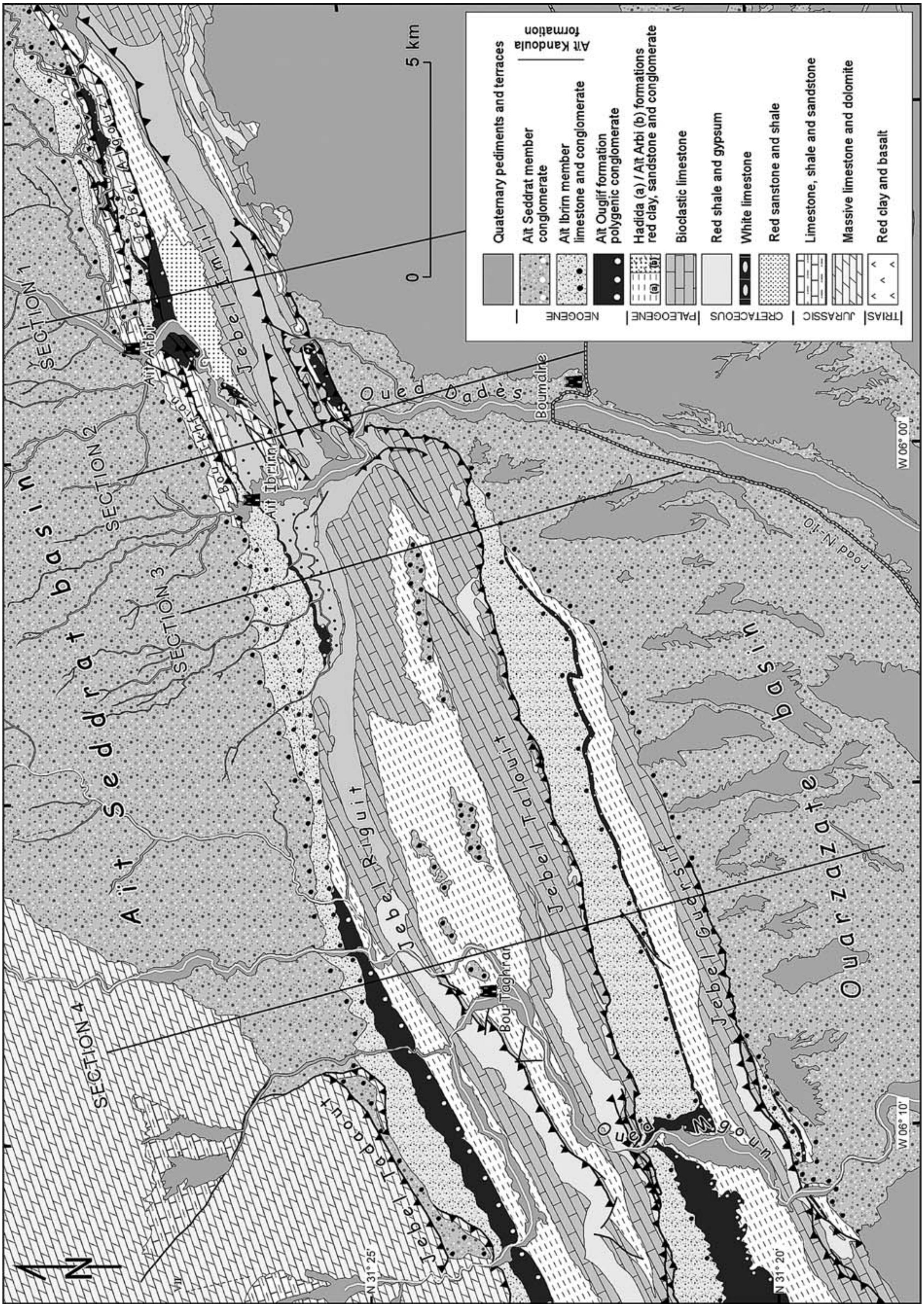


Fig. 3 Geologic map of the Sub-Atlas Zone and northern Ouarzazate basin between the Mgoun and Dadès valleys. This map corresponds to a detailed version of Fig. 2. The *cross-section lines* of Fig. 4 are indicated

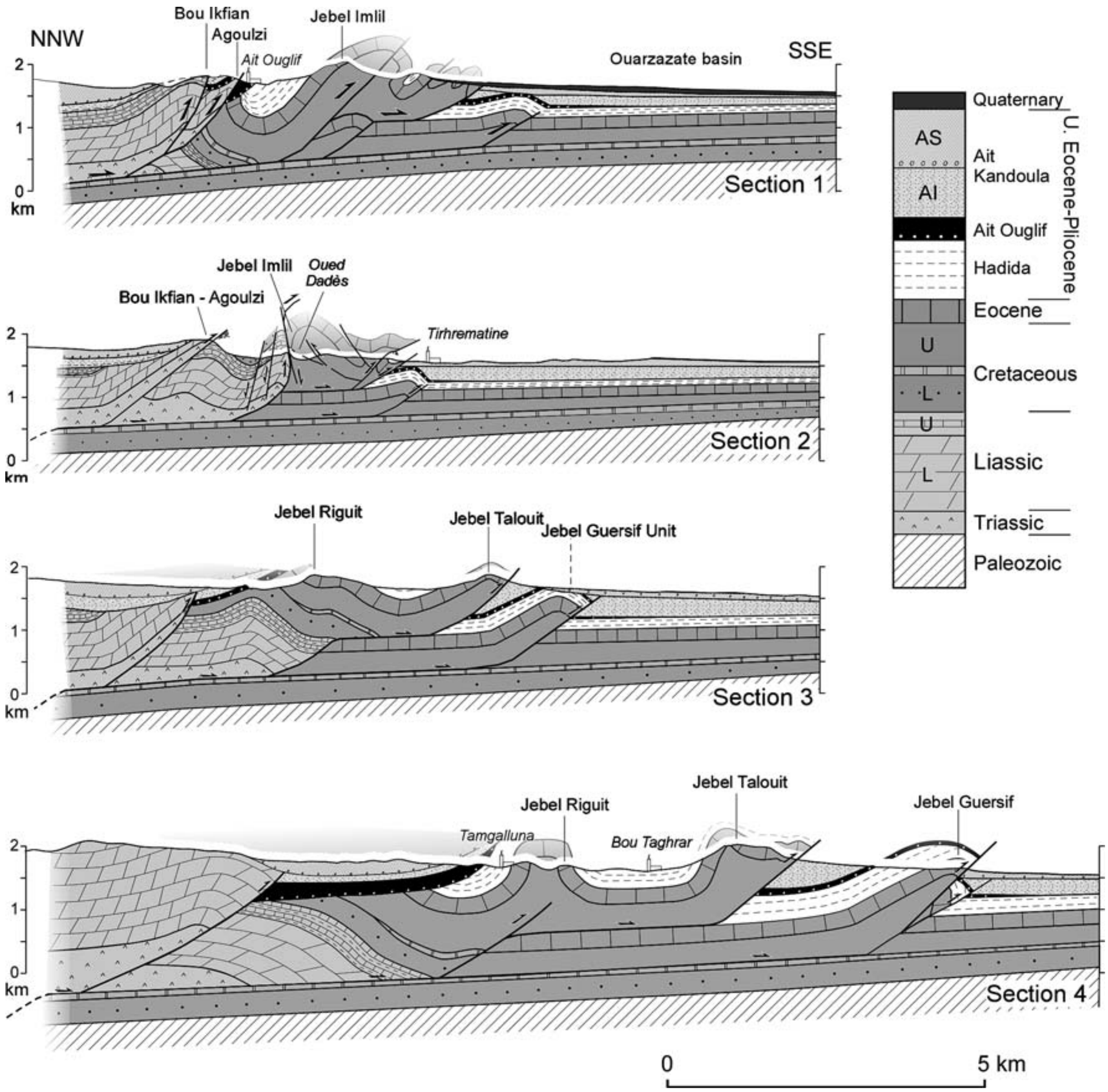


Fig. 4 Serial geologic cross-sections illustrating the structure of the Sub-Atlas thrust belt in the study area (see location in Fig. 3). *AI* Aït Ibrirn member, *AS* Aït Seddrat member

Cretaceous conglomerates (as seen in the Aït Arbi formation) indicate uplift in the internal parts of the High Atlas.

Syndeformational succession

The local syndeformational succession ranges from the lower-middle (?) Miocene to the Pliocene, although in some localities, Quaternary pediments and terraces are also slightly deformed. It begins with the Aït Ouglif formation (Fraissinet et al. 1988; El Harfi et al. 2001), constituted by polygenic conglomerates with rounded to subangular pebbles of Precambrian to Paleogene rocks deposited in an alluvial fan environment. Its mean thickness is 30–40 m. This formation lies with an angular unconformity or paraconformity on previous stratigraphic units. Where preserved in ramp anticlines of the thrust sheets, the Aït Ouglif conglomerate is always unconformable.

The age of this formation is still poorly constrained; it has been attributed to the upper Eocene (Gauthier 1957), to the lower Miocene (Görler et al. 1988) and to the Oligocene (Fraissinet et al. 1988; El Harfi et al. 2001). We come back partly to Görler's attribution, and due to their conformable transition to overlying formations of better known age, we retain a lower to middle Miocene age for this formation. Prior to the deposition of this formation there was an episode of extensive erosion recorded in the entire Sub-Atlas zone, roughly constrained between the probable

Oligocene age of the top of the Hadida/Aït Arbi formations, and the lower to middle Miocene age of the Aït Ouglif formation.

The bulk of the syndeformational sequence is constituted by the Aït Kandoula formation, a succession of shales, lacustrine limestones and alluvial fan conglomerates. The age of this formation is Middle Miocene–Pliocene based on micromammal fossils and magnetostratigraphy in the Aït Kandoula basin, north of the Sub-Atlas zone and west of the study area (Benammi et al. 1995, 1996; Benammi and Jaeger 2001). We subdivide the Aït Kandoula Formation in two members according to lithology and to relationships between sediments and deformation (Figs. 3, 4). The lower member is mainly composed of shales and lacustrine limestones with some conglomeratic intercalations (Aït Ibrirn member, 400 m thick), and in the upper member are massive conglomerates mainly composed of Jurassic clasts (Aït Seddrat member, 300–400 m thick). Upper bounds for the age of the Aït Kandoula formation in the study area and surroundings may be provided by 2.9 ma ankaratrite lavas at Foum el Kous (Schmidt 1992), 22 km to the east of the mapped area, which post-date the stage of aggradation and even the initial erosion of the Ouarzazate basin.

Structure of the Sub-Atlas fold and thrust belt

Previous studies focused on the structure of the Sub-Atlas thrust belt by Fraissinet et al. (1988) and Errarhaoui (1997) present valuable contributions with respect to the imbricate thrust structure and the overall synsedimentary character of the deformation. However, descriptions provided by these authors usually do not allow placing the individual thrust structures in a correlated stratigraphic frame.

A series of balanced cross-sections for the study segment of the Sub-Atlas zone is presented (Figs. 3, 4). Seismic control is lacking for the area, but fairly complete exposure and well-known stratigraphic thicknesses of the involved formations allowed the construction of these geological cross-sections with high degree of confidence. The regional elevation of the sedimentary formations and the top of the basement in the undisturbed foreland (i.e., the Ouarzazate basin) was constrained from seismic profiles to the west of the study area (Errarhaoui 1997; Tesón 2005), and from outcrops of the basin floor to the east in the Tinerhir area (Fig. 1). As shown in the cross-sections, the structure is characterized by folds and thrusts detached in two different levels. In the internal (northern) part, detachment is located in Triassic rocks whereas in the external part the detachment level is located within upper Cretaceous rocks (Fig. 4). We have distinguished three structural units (Fig. 2), which from north to south are: (1) the Aït Seddrat thrust sheet, a large nappe constituted mainly by Jurassic rocks, (2) the Talouit thrust sheet, composed of Mesozoic to Tertiary rocks with a detachment level that climbs southwards from the Triassic to the Upper Cretaceous (Fig. 4), and (3) the Guersif thrust sheet, entirely detached in the upper Cretaceous and carrying rocks from upper Cretaceous to Miocene age.

The bordering faults of the Aït Seddrat thrust sheet are the Bou Ikhfian and Algouzi thrusts (Figs. 3, 4). This thrust sheet defines a wide syncline that supports the Aït Kandoula basin 5 km to the west of the study area, and the Aït Seddrat basin directly to the north (Fig. 3). Jurassic rocks override the Aït Ouglif and Aït Ibrirn formations, while the frontal thrusts are fossilized by the Aït Seddrat member. In the westernmost part of the area studied, west of the Dadès valley, a late reactivation of these thrusts at Jebel Tadaout (Fig. 3) cuts the Aït Seddrat conglomerate.

The frontal thrust of the Talouit unit brings Cretaceous to Eocene rocks on the Aït Kandoula formation (Aït Ibrirn member) (Figs. 3, 4). The thrust sheet contains a complete stratigraphic succession, which starts with Jurassic carbonates exposed in the Dadès valley. The subsurface continuation of the Talouit thrust defines the southern margin of the Jurassic rocks (Errarhaoui 1997; Fig. 4), and hence it can be interpreted as an ancient normal fault defining the Jurassic basin margin, later sealed by the Cretaceous and Paleogene post-rift sequence. In this regard, this Jurassic unit is equivalent to the Toundout nappe of the western Sub-Atlas zone described by Laville et al. (1977). In the central Dadès area, the internal structure of the Talouit unit is complex due to a system of NE-trending high-angle normal faults that overprint the thrust structure (Figs. 3, 4; section 2). The normal faults are those that produce the only exposure of the Jurassic basement of the thrust sheet, mentioned previously. In detail, an internal south-vergent thrust was folded by a north-vergent backthrust system in the Jebel Imlil area (Fig. 4; section 2); the structure was then reactivated or slightly displaced by the late normal faults.

The frontal thrust of the Guersif thrust sheet brings the Cretaceous and Eocene rocks on the Aït Ibrirn member of the Aït Kandoula formation (Figs. 3, 4). In the Dadès valley, this thrust sheet is buried by the conglomerates of the Aït Seddrat member (Fig. 3), whereas in the Mgoun valley, the Aït Seddrat member

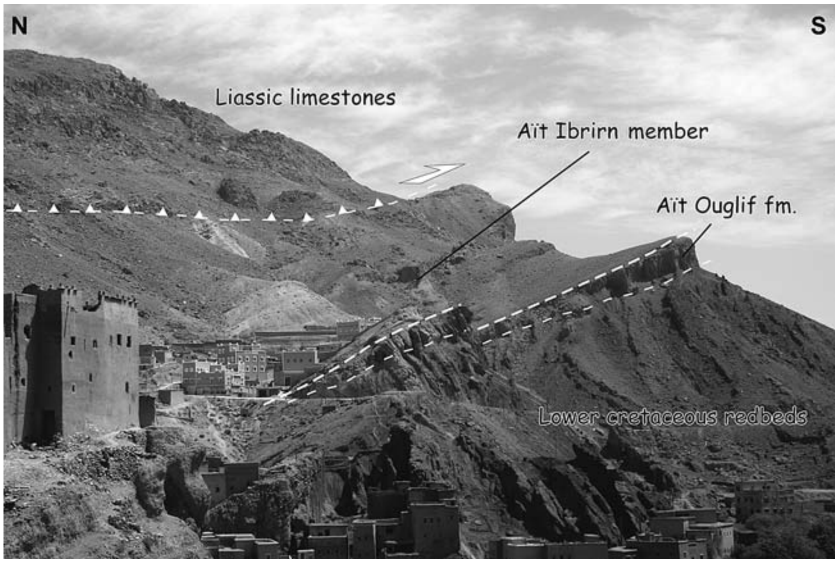


Fig. 5 Field photograph of out-of-sequence thrusting at Aït Ibrirn (Dadès valley). The Aït Ouglif conglomerate rests unconformable on folded lower Cretaceous rocks of the Talouit thrust sheet, and, together with the overlying lower part of the Aït Kandoula formation, is in turn overthrust by the Triassic and Jurassic rocks of the Aït Seddrat nappe (Bou Ikhfian thrust)

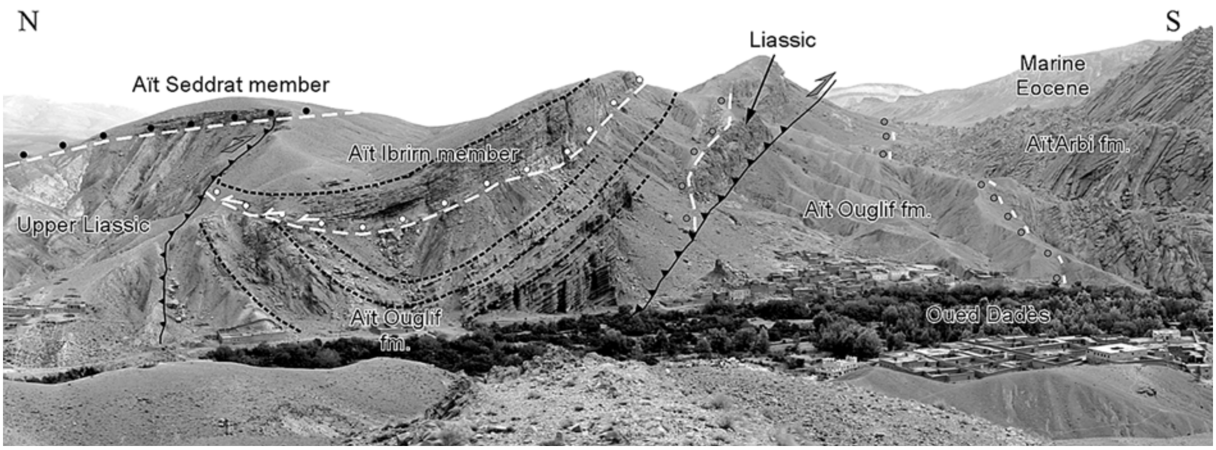


Fig. 6 View of deformed syntectonic formations in the Dadès valley. *Faults* indicated in the picture are the Algouzi and Bou Ikhfian thrusts, which override a large north-dipping panel at the rear of the Talouit thrust sheet. The *village* in the center of the image is Aït Ouglif

is overridden by the Guersif sheet unit due to an out-of-sequence reactivation of the basal thrust. The structural elevation of the sedimentary formations within the Guersif thrust sheet suggests that this unit lacks pre-Cretaceous rocks, and is entirely detached in the upper Cretaceous red shale (Fig. 4).

In the Talouit and Guersif units, fault-bend folds, detachment folds and fault propagation folds are all common (Fig. 4). Detachment folds are defined by Eocene competent limestone and usually cored by the upper Cretaceous shale due to its low competence.

The tectonic shortening, calculated from the restoration of the cross sections in the study area, is about 7–8 km (Tesón 2005), which must represent a large fraction of the total shortening of the High Atlas. Although orogenic shortening estimations for this transect of the High Atlas are lacking, Beauchamp et al. (1999), Benammi et al. (2001) and Teixell et al. (2003) demonstrated that the major shortening occurred in the southern margin of the chain whereas the internal parts of the chain were only slightly deformed.

Timing and sequence of deformation

The excellent exposure in the Sub-Atlas thrust belt (Figs. 5, 6) enables the analysis of tectonics–sedimentation relationships and the unraveling of the relative timing of deformation. The absolute defor-

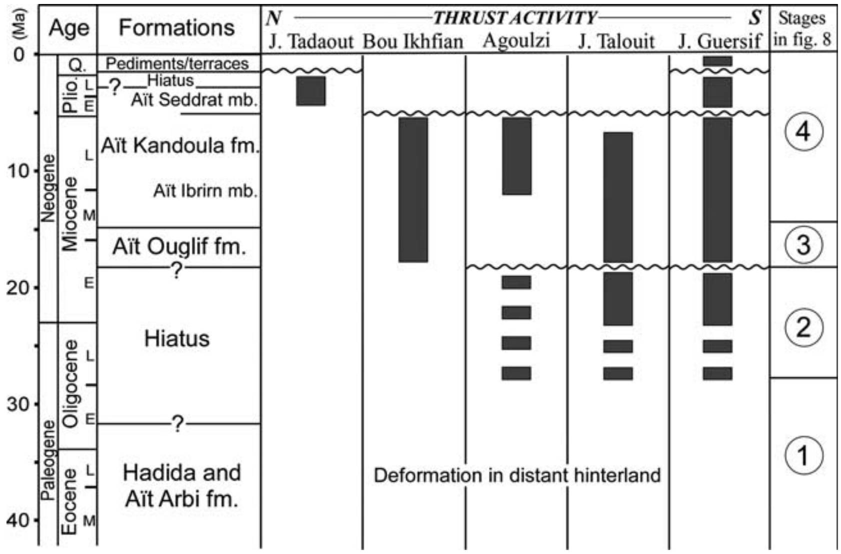


Fig. 7 Summary of thrusting chronology of the Sub-Atlas thrust belt between the Mgoun and Dadès valleys. *Vertical bars* represent the activity time of each thrust. *Undulated lines* indicate unconformable coverage by syntectonic sediments. See text for discussion on the uncertainties of the age attributions

mation chronology is still not completely established because precise ages of the continental synorogenic formations and intervening hiatuses are still not sufficiently constrained (the chronostratigraphic attributions by the previous authors diverge, e.g., Gauthier 1957; Fraissinet et al. 1988; Görler et al. 1988; El Harfi et al. 2001).

The first unequivocal evidence for local deformation is provided by the Aït Ouglif formation. This conglomerate unit is largely unconformable on the first compressional structures. In the Dadès valley, with exposure of the deepest structural levels, the conglomerate is seen over steep Jurassic to Cretaceous strata of the ramp anticline at the level of the Mesozoic of the Talouit unit (Figs. 4, 5).

Further south, the Aït Ouglif conglomerate is locally unconformable on the frontal anticlines associated with the Jebel Talouit and Guersif thrusts (Figs. 3, 4, section 4), indicating early activity of these structures too. Nevertheless, these thrusts were strongly reactivated in later times, since they cut the Aït Kandoula formation (Figs. 3, 4). This reactivation attests for continued displacement transfer from the rear, i.e., through the ramp anticline at the Mesozoic level of the Talouit unit (Fig. 4). Since the Aït Ouglif conglomerate lying on the eroded anticline is almost flat, fold amplification must have stopped by that time, so movement of this fault-bend fold in post-Aït Ouglif times must have taken place in a crestal-broadening fashion (in the sense of Shaw et al. 1994).

The northernmost Bou Ikhfian and Algouzi thrusts also cut the Aït Ouglif formation and the Aït Ibrirn member, indicating that they splay from the Talouit sheet in an out-of-sequence fashion (Fig. 5). The Aït Seddrat member fossilized these thrusts (Fig. 6), although in the western end of the study area it is affected by another out-of-sequence thrust at Jebel Tadaout (Fig. 3). In the frontal part of the Sub-Atlas belt, the Aït Seddrat member overlaps the Jebel Guersif thrust in the Dadès valley area, and it is cut by the Jebel Talouit thrust in the north (Fig. 3).

Hence, the propagation sequence of the Sub-Atlas imbricate fan is not simply a piggy-back one, but includes reactivation, out-of-sequence and synchronous thrusting. The activity times of each individual thrust fault on the basis of its relationships to the syntectonic sedimentary formations is shown in Fig. 7. This diagram reflects a protracted deformation history, beginning some time before the Aït Ouglif sedimentation (Oligocene?), and continuing at least to the deposition of the Aït Seddrat formation (undetermined Pliocene). Shortening rates were low, and averaged about 0.3 mm/a (7–8 km during 20–25 Ma). During the Quaternary, tectonic shortening continues at the front of the Guersif unit, as demonstrated by the presence of gentle deformed pediments and terraces in the western continuation of this thrust in the Toundout valley (Couvreur 1973; Sébrier et al. 2006).

Some authors postulated that the onset of the Atlas orogeny took place in the late Cretaceous or the early Tertiary (Laville et al. 1977; Froitzheim et al. 1988; Amrhar 1995). In the study area, no clear evidence of the start of the deformation during these early times was found; however, some sedimentologic arguments point to a beginning of the Atlas uplift prior to the development of the thrust sequence recorded in the Sub-Atlas zone. Siliceous pebbles found in the middle Eocene to Oligocene Aït Arbi formation were previ-

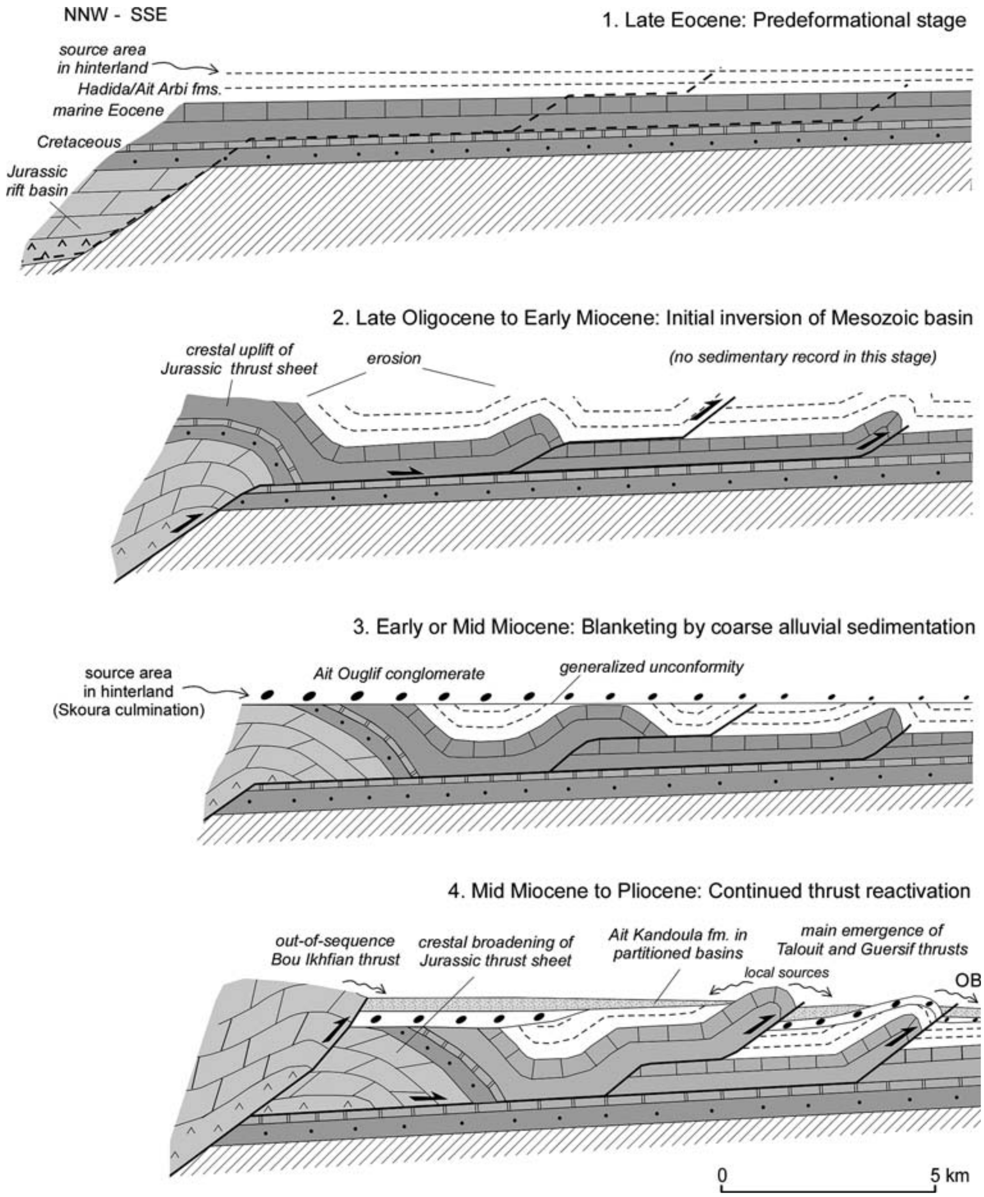


Fig. 8 Sequential evolution of the Sub-Atlas thrust belt and adjoining basins in selected stages, based on the restoration of section 4. Stage 3 should be regarded as a snapshot describing the base of the Aït Ouglif event, and not as a protracted period of tectonic quiescence. *OB* Ouarzazate basin

ously interpreted as derived from the Anti-Atlas (El Harfi et al. 2001); however, their composition, mature texture and restricted geographic extent suggest that they rather derive from the erosion of the lower Cretaceous conglomerates of the High Atlas. We interpret that they record mountain building in the High Atlas hinterland north of the Sub-Atlas thrust belt, probably in the Skoura antiformal culmination and other northern structures (Fig. 1). Hence, the conglomerates of this formation, together with the laterally equivalent Hadida formation, can be regarded as the earliest foreland basin deposits to the south of the Atlas Mountains. It is possible that some growth folding had already started at that time in local structures within the Sub-Atlas belt, such as the syncline south of Aït Ouglif (Dadès valley; Fig. 4; section 1), which traps a thick accumulation of Aït Arbi conglomerates, but direct geometric evidence in the field is lacking. With regard to even earlier deformation, the unconformities described at the base of the marine Paleogene by Laville et al. (1977), and the siliceous clasts locally present in the Eocene limestones also point to some local uplift in the Atlas domain, although the paleogeographic reconstruction indicates a marine gulf linked to the Atlantic margin at that time and not a peripheral foredeep of an orogen (Herbig and Trappe 1994).

Figure 8 illustrates the thrust sequence of the study segment of the Sub-Atlas belt based on the restoration of cross-section 4. Stage 1 represents the sedimentation of the Hadida and Aït Arbi formations. No local deformation has been assumed for this stage, although deformation and uplift should occur in the Atlas hinterland. Stage 2 represents the initial inversion of the Jurassic basin margin, the Mesozoic rocks climbing the ramp in a crestal uplift stage of fault-bend folding.

The emergence of the faults to the surface at this stage produced the initial deformation at the Jebel Talouit and Guersif thrusts (Fig. 8-2). The Aït Ouglif sedimentation punctuates this deformation, covering unconformably these structures (Fig. 8-3). The polygenic nature of the Aït Ouglif pebbles reflects erosion from Jurassic carbonates (locally sourced) and lower Cretaceous (and, possibly, Aït Arbi) conglomerates, that we attribute to continued uplift of the Skoura culmination (which was actually the root zone of the Sub-Atlas thrusts). Finally, Fig. 8-4 describes deformation during the sedimentation of the Aït Kandoula formation, including crestal broadening of the Mesozoic anticline in the rear of the Talouit thrust sheet, displacement transfer and reactivation of the Talouit and Guersif thrusts, and out-of-sequence thrusting of the Aït Seddrat unit (e.g., Bou Ikhfian thrust). The sedimentary basin during this stage was compartmentalized by the emergence of the thrust sheets, and small terrigenous to lacustrine basins, fed by the erosion of the anticline crests, were isolated.

Conclusions

The structural geometry of the Sub-Atlas thrust belt between the Dadès and Mgoun valleys (southern margin of the Moroccan High Atlas) defines an imbricate fan of three main thrust sheets (Ait Seddrat, Talouit and Guersif). Detachment levels climb from the Triassic to the upper Cretaceous, whereas the leading edges of the thrust sheets override and partition the synorogenic basins of Neogene age.

The study segment of the Sub-Atlas belt provides constraints to the debate on the timing of compressional deformation in the southern High Atlas Mountains. Timing of the mantle-sourced uplift and its relationship with tectonic shortening processes in the Atlas orogenic system is still to be unravelled, but this work sets main constraints about the compressional deformation sequence and timing. With the available dating of the sedimentary formations, a detailed analysis of the relationships between tectonics and sedimentation (cross-cutting relationships, unconformities) suggests that the main shortening activity in the Sub-Atlas thrust belt spanned rather continuously from the Oligocene to the Pliocene, and continues at a lower rate until recent times. A widespread detrital influx into the basin at mid Eocene times suggests that deformation in the High Atlas hinterland was probably regionally significant in these early times, although emerging thrusts had not reached the Sub-Atlas zone yet.

Main thrust faults did not propagate in a simple piggy-back fashion, but display a complex, synchronous sequence with events of fault reactivation and out-of-sequence thrusting. Geological cross sections suggest shortening values of about 7 to 8 km for the Sub-Atlas belt, accommodated at a low rate of about 0.3 mm/a in average.

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