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Folding and inversion tectonics in the Anti-Atlas of Morocco

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[1] The late Variscan Anti-Atlas of Morocco shows some conspicuous deviations from the standard anatomy of foreland fold-and-thrust belts. Large basement inliers crop out at a very short distance of less than 50 km behind the southeastern front of the fold belt, reminiscent of Windriver-style basement uplifts. In contrast to the Rocky Mountain foreland, however, the Anti-Atlas basement uplifts punctuate tightly folded Paleozoic cover series similar in tectonic style to the Appalachian Valley and Ridge province. Cover shortening is exclusively accommodated by buckle folding, and the Anti-Atlas fold belt lacks any evidence for duplexing or thrust faults other than the occasional steep reverse fault found near basement inliers. Basement domes have classically been considered as the result of vertical tectonics in a horst and graben fashion, or, alternatively, as large "plis de fond" [*Argand*, 1924], basement folds. Unfolding of a large portion of an Ordovician quartzite marker bed reveals a minimum shortening of 17% (30 km). Balancing this section at the crustal scale indicates a lower crustal detachment level at 18 to 25 km depth. Basement shortening is inferred to be accommodated through massive inversion of former extensional faults, inherited from a Late Proterozoic-Lower Cambrian rifting phase.

INDEX TERMS:
8005 Structural Geology: Folds and folding;
8015 Structural Geology: Local crustal structure;
8102 Tectonophysics: Continental contractional orogenic belts;
8159 Tectonophysics: Rheology—crust and lithosphere;
9305 Information Related to Geographic Region: Africa;

KEYWORDS:
Anti-Atlas, tectonics, balanced cross section, folding, foreland fold and thrust belt.

1. Introduction

[2] The Anti-Atlas foreland fold belt of southwestern Morocco is part of the Variscan Appalachian-Ouachita-Mauretanides chain. It shows some important deviations from typical foreland fold and thrust belts [*Rodgers*, 1990]. Large structural domes, the so-called "boutonnieres" [*Choubert*, 1963] of Proterozoic basement rocks, crop out at a very short distance behind the southeastern front of the orogen. Detailed mapping of the Anti-Atlas did not reveal any evidence for major thrusts within this fold belt. Layer-parallel décollements or detachments are required between different structural levels, but no such décollement has ever been mapped to step up in a ramp-flat geometry. The complete lack of thrusts lead earlier authors to consider the entire basement uplifts as crustal-scale folds [*Choubert*, 1963; *Choubert and Faure Muret*, 1971], inspired by the plis de fonds of *Argand* [1924, Figures 5a, 5b, and 5c].

[3] The sedimentary cover of the Anti-Atlas includes the uppermost Proterozoic and an up to 10 km thick Paleozoic series. These mildly deformed and nearly unmetamorphosed strata of the Anti-Atlas have received much attention from stratigraphers and paleontologists [*Destombes et al.*, 1985; *Bertrand-Sarfati et al.*, 1991; *Villeneuve and Cornee*, 1994; *Piqué*, 2001]. Special attention has been given to the Proterozoic-Cambrian boundary [*Boudda et al.*, 1979; *Bertrand-Sarfati*, 1981; *Buggisch*, 1988b; *Latham*, 1988; *Geyer and Landing*, 1995; *Benssaou and Hamoumi*, 2003]. In contrast to this wealth of stratigraphical and paleontological literature, structural publications dealing with the tectonics of the western Anti-Atlas remain scarce [*Leblanc*, 1972; *Michard*, 1976; *Soulaimani*, 1998; *Guiton et al.*, 2003].

[4] On the basis of paleomagnetic observations near the South Atlas and the Tizi'n'Test faults, a strike-slip origin for the entire Anti-Atlas chain has been proposed [*Mattauer et al.*, 1972]. This view is widely shared by French authors [*Leblanc*, 1972; *Donzeau*, 1974; *Jeannette*, 1981], and folding of the sedimentary cover has mostly been explained as draping over vertically uplifted basement blocks and/or as wrenching above inherited, subvertical "zones of weakness" [*Faik et al.*, 2002]. Horizontal shortening in the central Anti-Atlas has been ignored, or estimated to be on the order of 5 to 10% at most [*Leblanc*, 1972; *Donzeau*, 1974]. Published cross sections of the western Anti-Atlas depict the basement inliers as horst and graben structures, indistinct broad basement domes or delimited by vertical strike-slip faults [*Michard*, 1976; *Michard and Sougy*, 1977; *Piqué et al.*, 1991]. Clearly, these cross sections imply no crustal-scale shortening within the basement, and fold trains within the synclinoria remain unexplained in terms of material balance, a question which has never been addressed in the Anti-Atlas. The striking difference between the Devonian "Jbel Rich" folds and the virtually undeformed Jbel Ouarkziz monocline near the southern front of

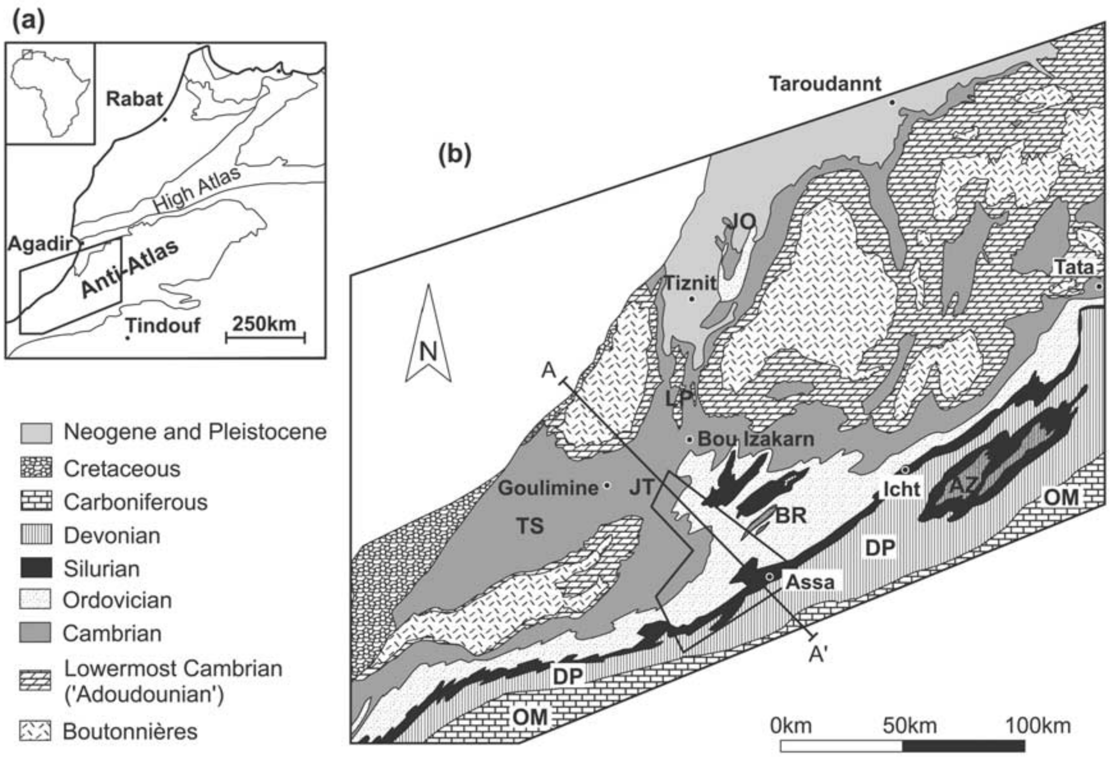

Figure 1. (a) Location of the Anti-Atlas in southwestern Morocco. The polygon represents the study area. (b) Simplified geological map of the westernmost Anti-Atlas showing the irregular shape and distribution of the structural domes. These Proterozoic "boutonnieres" consist of pre-Cambrian crystalline basement and Late Proterozoic volcaniclastic series of variable thickness. The black line A-A′ shows the location of the large-scale cross section of Figure 10. The polygon represents the location of the 3-D fold model shown in Figure 6. Abbreviations are as follows: AZ, Adrar Zouggar mountain; BR, Bani range; DP, Drâa plain; JO, Jebel Ouarzemine; JT, Jebel Tayyert; LP, Lakhssass plateau; OM, Ouarkziz monocline; TS, thin skin part of the "internal" Anti-Atlas.

the Anti-Atlas has been considered as evidence for a major strike-slip fault [*Weijermars*, 1993], of a Neogene age. The apparent "en echelon" arrangement of the Rich folds was further used to deduce a right lateral shear sense and amount of shear, assuming a subvertical plunge of fold axes [*Weijermars*, 1993]. In reality, however, fold axes plunges are very gentle, less than 30°, and the asymmetric fold pattern of the lower Drâa valley implies very weak, if any, wrenching deformations [*Soulaimani et al.*, 1997]; paleostress indicators are systematically orientated nearly perpendicular to the fold axial planes, rather than at an inferred angle of some 50° [*Weijermars*, 1993]. New observations in the southwestern Anti-Atlas have recently led to a general revision of the classic model of vertical and strike-slip tectonics [*Soulaimani et al.*, 1997; *Soulaimani*, 1998]. Convincing evidence exists for southwestward thrusting of the Lower Drâa inlier onto the sedimentary cover, and the dextral strike-slip boundary postulated by *Weijermars* [1993] is not substantiated by any field evidence.

[5] In this study, we present a detailed structural analysis of a large portion of the SW Anti-Atlas. A continuous fold train in a competent Ordovician quartzite bed is mapped in detail and rendered in a three-dimensional (3-D) structural model. Unfolding of this marker bed provides minimum estimates of horizontal shortening as well as a measure of regional variations in folding intensity. The consequences of this tectonic shortening are discussed at the larger, crustal scale of this "thick skinned" foreland fold belt.

2. Geological Framework of the Western Anti-Atlas

2.1. Basement Preconfiguration

[6] The Anti-Atlas fold belt is located at the northwestern border of the West African craton and continues southward into the Mauretanides. Large-scale basement cored domes with irregular shapes form a continuous area of positive structural relief (Figures 1 and 2). The long axis of the Anti-Atlas belt axis is oriented SW-NE and measures some 700 km.

[7] The pre-Cambrian basement has a complex geologic history. It was consolidated during the Panafrican orogeny between 620 and 580 Ma [*Choubert*, 1963; *Choubert and Faure Muret*, 1971; *Leblanc and Lancelot*, 1980; *Hassenforder*, 1987]. In detail it shows considerable

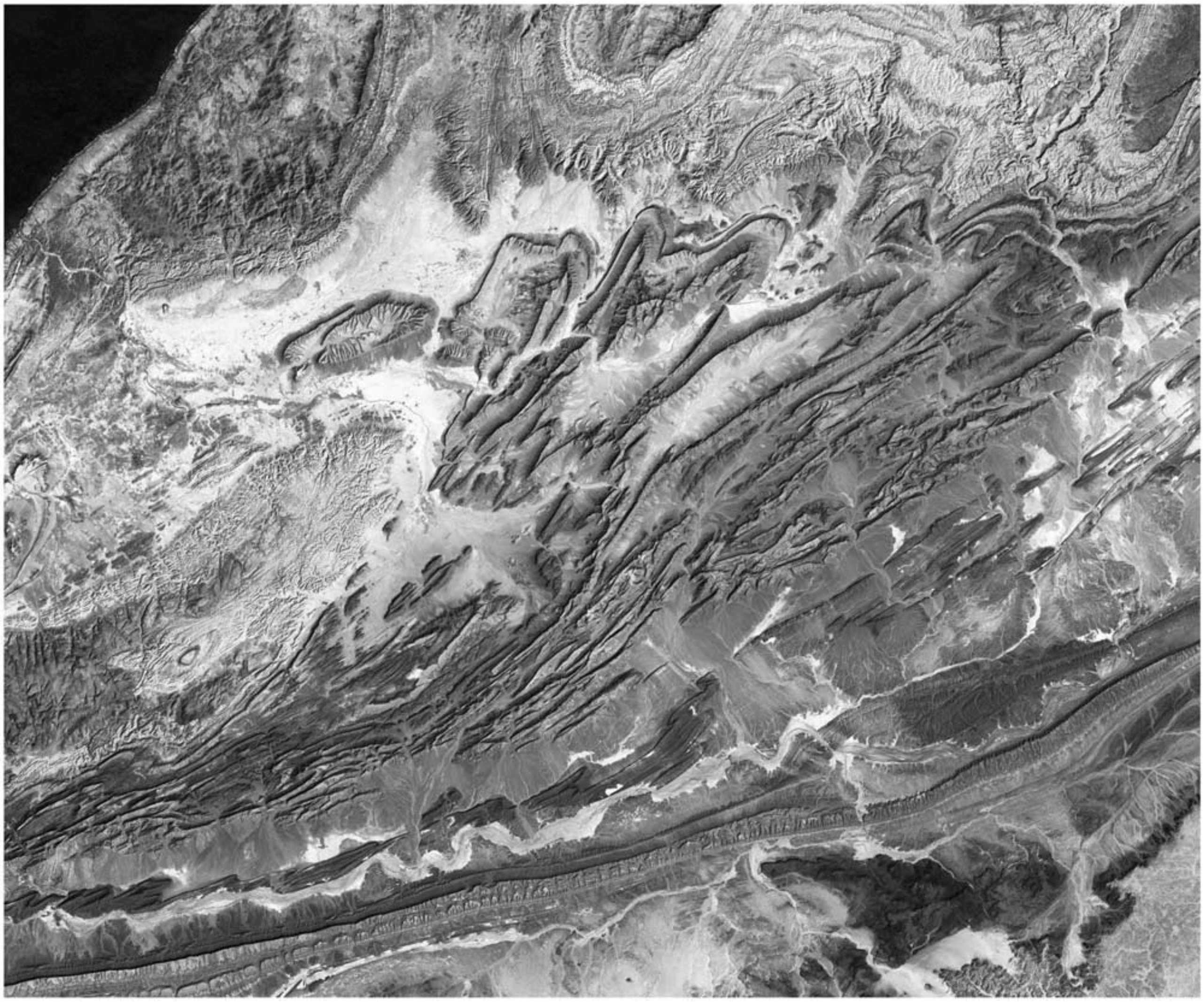

Figure 2. Landsat image of the westernmost Anti-Atlas chain. Compare with Figure 1 for major landmark features such as the basement inliers (dark green) with their ‘‘autochthonous’’ Cambrian cover in light green and tan colors. Quartz-rich lithologies of the Jbel Bani, the Jbel Rich, and Jbel Ouarkziz appear in dark purple. Image courtesy of NASA and Earth Satellite Corporation (available at http://zulu.ssc.nasa.gov/mrsid/mrsid.pl). See color version of this figure at back of this issue.

heterogeneity. It contains the suture zone between the stable West African craton and a Proterozoic mobile belt to the north. Remnants of this suture zone are exposed in several basement domes, for example in the Bou Azzer and in the El Graara inliers. In the central Anti-Atlas this suture zone is known as ‘‘accident majeur de l’Anti-Atlas.’’ Southwest of this suture zone, Eburnean (approximately 1900 Ma) augen gneisses, metadolerites, and metamorphic rocks, the so-called Zenaga and Kerdous Series, crop out in several of the structural domes. The Proterozoic rocks are further characterized by numerous synorogenic granite and granodiorite intrusions.

[8] The Panafrican orogeny is followed by an important phase of continental sedimentation, which results in locally thick clastic and volcaniclastic deposits, the so-called PII-III unit, separated from the crystalline basement by a major unconformity. The geodynamic context of the PII-III is still a matter of debate [*Piqué et al.*, 1999; *Piqué*, 2001]. In the case of the overlying PIII unit of clastics and volcaniclastics, separated by a minor unconformity from the PII, there is increasing evidence in favor of an extensional tectonic regime, in relation with a new rifting cycle starting in the Late Proterozoic [*Piqué et al.*, 1995; *Piqué*, 2001; *Soulaimani et al.*, 2003]. Extensional structures have been described within the Proterozoic basement in the westernmost Anti-Atlas [*Soulaimani et al.*, 1997] and large-scale half-grabens are documented in the central Anti-Atlas [*Azizi-Samir et al.*, 1990]. The post-Panafrican geological history of the Anti-Atlas is now considered as an aborted rift [*Piqué et al.*, 1995; *Soulaimani et al.*, 2003].

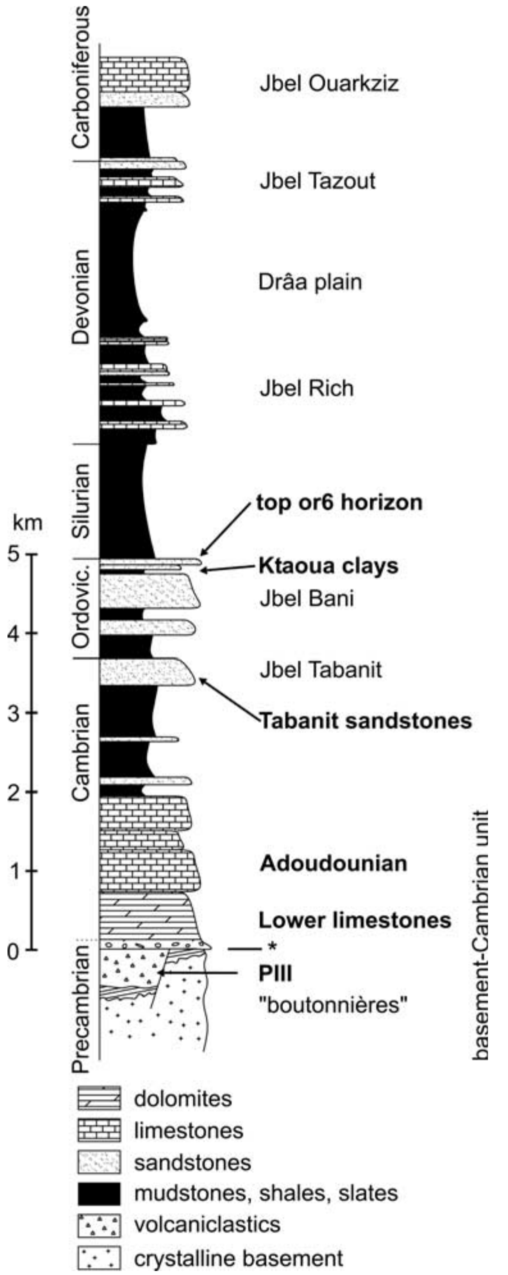

Figure 3. Synthetic stratigraphic column of the western Anti-Atlas. Incompetent units are shown in black. The asterisks marks the slight unconformity which is classically used to define the outlines of the boutonnieres (basement inliers). The commonly used, informal lithostratigraphic/mechanical units are labeled at the right-hand side; chronostratigraphy is shown to the left.

2.2. Sedimentary Cover

[9] Above the PIII conglomerates, a slight unconformity marks the onset of epicontinental shallow marine sedimentation in the Anti-Atlas, a limit often referred to as the "basement cover contact." A marked contrast in color, sediment type and rheology makes the PIII-Adoudounian contact an easy target for mapping both in the field and on satellite images (Figure 2). The same unconformity is also used to define the outline of the basement inliers.

[10] A new sedimentary cycle leading to the formation of the Anti-Atlas basin starts in the uppermost Proterozoic and ends in the upper Carboniferous. In the western Anti-Atlas, a total of up to 12 km of varied clastic and shallow marine sediments were deposited, while eastward, in the central Anti-Atlas, the thickness is decreasing to about 8 km [*Michard*, 1976; *Piqué and Michard*, 1989; *Kogbe*, 1998]. The Anti-Atlas basin fill is characterized by a high amount of fine-grained detrital, clay rich sediments: muddy siltstones and shales (Figure 3). Competent quartzites, sandstones, limestones, and conglomerates form thin but continuous beds at different stratigraphic levels. Coarse-grained beds are increasing in thickness and abundance toward the easternmost Anti-Atlas.

[11] The thin competent marker beds form very continuous spectacular crests easily visible in aerial and satellite images (Figure 2). Major stratigraphic intervals such as the Ordovician and the Devonian series have been named according to the mountain belts they form: Jbel Bani (for Ordovician), Jbel Rich (Devonian), and Jbel Ouarkziz (Carboniferous). We continue to use this terminology in order to describe the main competent units, despite the fact that they are not formally accepted "lithostratigraphic" formation names.

2.3. Variscan Orogeny

[12] The age of deformation within the western Anti-Atlas is generally considered as middle to late Carboniferous on the basis of the youngest age of (slightly) deformed and tilted sediments on the one hand, and superregional considerations and comparisons on the other hand [*Michard*, 1976; *Piqué and Michard*, 1989; *Piqué*, 1994]. The oldest sedimentary formations sealing eroded Anti-Atlas folds are of Upper Cretaceous age and do not provide any useful timing constraints. Dolerite dikes such as the famous dike of Foum-Zguid and sills clearly postdate Anti-Atlas folding [*Sebai et al.*, 1991], but their lower Jurassic age does not put any tight time constraint on the deformation age of Anti-Atlas folding either, since folding might still be anything between "middle" Carboniferous and Late Triassic. Some attempts at a direct age determination of the tectonic cleavage have been made using radiometric methods, applied to fine grained illite-muscovite assemblages of the westernmost Anti-Atlas [*Bonhomme and Hassenforder*, 1985]. Isotopic ages of 370 Ma and 290 Ma were attributed to a pre-Variscan phase of deformation and the peak of regional metamorphism, respectively. Similar results have been obtained in isotopic studies of the Variscan Meseta of northern and central Morocco [*Huon*, 1985]. Since metamorphism in many parts of the Variscan chain of Morocco is barely of lower greenschist facies, and deformation only locally penetrative, any K-Ar and Rb-Sr ages should be regarded very cautiously [*Villa*, 1998]. Correlations between the northern Moroccan Meseta and the Anti-Atlas are not straightforward either. The two

Figure 4. A gently inclined fold limb north of Assa. The entire dip slope of this fold limb is formed by one stratigraphic horizon of Upper Ordivician age: the or6c-quartzites. The foreground of the photograph shows the Drâa plain corresponding to a lowland eroded into Silurian shales; view is to the north.

provinces are separated by the South Atlas shear zone, a major terrane boundary within the Variscan chains of Morocco [*Mattauer et al.*, 1972]. An overall geodynamic interpretation of a complete Canadian-Moroccan transect through the entire Appalachian-Meseta-Anti-Atlas system has been presented as early as 1971 [*Schenk*, 1971].

2.4. Anti-Atlas Foreland Fold Belt

[13] In terms of folding and deformation style, four structural units with different wavelengths and amplitudes can be distinguished in the Anti-Atlas. They correspond to four distinct stratigraphic levels separated by the thick incompetent units of the Middle Cambrian, Silurian, and Upper Devonian respectively. None of the fold levels shows a clear and consistent vergence.

[14] The lowermost unit is represented by the basement domes and their autochthonous cover consisting of Late Proterozoic PIII conglomerates and Lower Cambrian limestones and dolomites of the so-called "Adoudounian." These basement domes have accessible, mapped amplitudes of up to 2 km, and the corresponding wavelength is on the order of several kilometers to tens of kilometers. The true "amplitude" of basement "folds" is in excess of 10 km, however, larger than the total thickness of Paleozoic cover series. Locally, thin layers of incompetent shales and evaporites within the autochthonous Adoudounian series act as minor detachment horizons, resulting in second-order folding at a much smaller scale (10 to 100 m). Toward the central Anti-Atlas the "Lie de vin" formation acts as a distinct local décollement horizon (Figure 2). Fold axis trends within the Adoudounian are strongly influenced by the preconfiguration of the underlying basement domes, and they often deviate considerably from the large-scale SW-NE Anti-Atlas orientation [*Soulaimani et al.*, 1997; *Soulaimani*, 1998]. Higher up in the stratigraphic column, two structural units are characterized by detachment folds of the Ordovician Jbel Bani quartzites (Figure 4) and the Lower Devonian Jbel Rich limestones, respectively. In the western Anti-Atlas they are separated from each other by about 1000 to 1500 m of Silurian shales and mudstones. The folding style of these two units is very similar, albeit with different wavelengths and amplitudes. Spectacular kilometer-scale Bani folds are found especially in the western Anti-Atlas between the towns of Bou Izakarn Assa and Icht (Figure 5). The Jbel Rich folds display smaller wavelengths and amplitudes at the 500 m scale. The uppermost structural unit, a thick series of Carboniferous sandstones and limestones of the Jbel Ouarkziz, is separated from the Jbel Rich folds by a thick series of Upper Devonian shales. Morphologically, the Jbel Ouarkziz defines a very continuous monocline, where the Carboniferous series dip gently south-southeastward into the adjacent Tindouf basin (Figure 2). Some authors (e.g., M. Zizi, personal communication, 2000; see also http://www.onarep.com) consider the lack of folding within the tilted Jbel Ouarkziz as evidence for a Late Devonian age of folding in the Anti-Atlas, an interpretation inspired by comparisons with northern and central Morocco and with the Canadian Appalachians [*Piqué and Michard*, 1989]. Direct field evidence in the form of an unconformity has never been identified in the sedimentary series of the Anti-Atlas, where a tectonic interpretation (décollement) is more generally accepted [*Soulaimani*, 1998]. While the deformation front of the Anti-Atlas system is classically seen at the base (NNW) of the Jbel Ouarkziz, we propose a new interpretation as a triangle structure with a forethrust ending somewhere blindly SSE of the Jbel Ouarkziz, at depth within the Tindouf basin. From this most external tip point of the Anti-Atlas thrust system, a NNW vergent backthrust climbs north-northwestward to emerge into the shale series separating the Jbel Ouarkziz monocline from the Jbel Rich fold train within the Lower Drâa valley. This interpretation is corroborated by structural observations, notably the tilting of the Jbel Ouarkziz monocline and bed internal layer

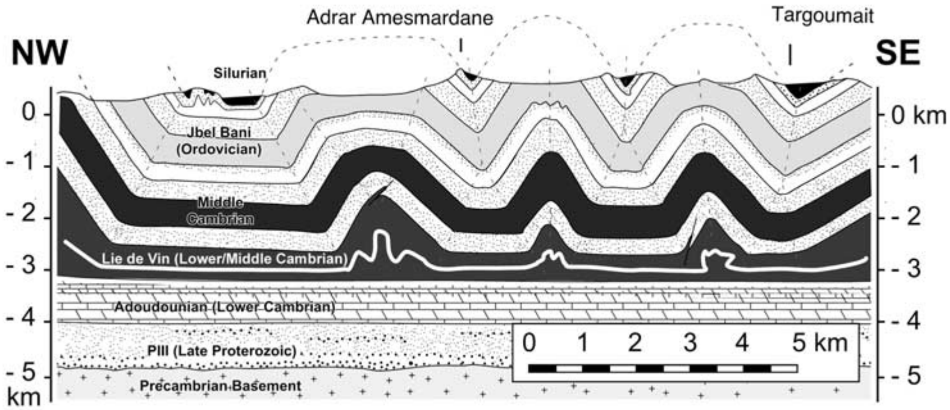

Figure 5. Intermediate-scale section across the folded Jbel Bani, in the northwestern sector of the 3-D fold model shown in Figure 6. A major décollement is inferred within the incompetent Lie de vin formation of Lower/Middle Cambrian age. The structural style below this décollement is unknown; for lack of better knowledge, it is depicted here as a flat-lying, undeformed panel of autochthonous basement-Lower Cambrian unit.

parallel shortening (LPS) deformation features within the youngest outcropping Carboniferous limestones [*Burkhard et al.*, 2001].

3. Anti-Atlas Cross-Section Balancing

3.1. Theoretical Considerations

[15] In areas of tectonic activity, deformation typically occurs at all scales, ranging from structures in crystal lattices to structures forming mountain ranges [*Ramsay and Huber*, 1983; *Mitra*, 1992; *Mitra*, 1994]. Three scales of observation are commonly used to describe structures: microscale (<0.01 m), outcrop scale (0.01–100 m), and map scale (>100 m). Primary sedimentary structures are used as proxies for the qualitative description and quantitative determination of strain at all scales.

[16] Cross-section balancing is dealing with map-scale structures such as folds, faults, and thrusts. Smaller-scale structures accommodate additional deformation, which should be considered in overall estimations of tectonic shortening. Over the years, various investigators have shown that small-scale structures do make a significant contribution to total tectonic deformation. *Mitra* [1994] measured strains in sections parallel to the transport direction across the Sevier fold-and-thrust belt. He could demonstrate that failure to include microscale deformation in the restoration of regional cross sections results in overestimation of the wedge taper for the thrust belt. Another study examined the contribution of both microscale and outcrop-scale data to regional deformation using the foreland thrust belt of the central Appalachians where small-scale structures account for more than 75% of the shortening in the roof sequence of this thrust belt [*Smart et al.*, 1997]. *Hogan and Dunne* [2001] conducted a study in the Upper Devonian Chemung-Brallier Formation boundary in the western Valley and Ridge province, where map-scale structures account for about 50% of the total shortening, whereas outcrop-scale and microscale shortening yield each close to 25% of the overall shortening.

[17] Qualitatively, layer parallel shortening features are widely recognized as predating folding and thrusting and this is very nicely demonstrated in the Jura mountains [*Plessmann*, 1972; *Homberg*, 1996].

[18] For the estimation of the total shortening in the Anti-Atlas we concentrated on map-scale and outcrop-scale structures. Since there are other important unknowns of the shortening at the scale of the orogen we were content with a qualitative assessment of microscale shortening. In order to minimize the effects of LPS, we choose a very competent quartzite horizon, devoid of any LPS features visible at the outcrop to hand specimen scale. We believe that this marker bed deformed essentially by buckling at the outcrop to map scale.

3.2. Study Area and the Ordovician Quartzite Marker Bed

[19] Our estimation of the shortening is mainly based on measurements and observations made in the Bani mountain range, an area of folded Ordovician rocks between the towns of Bou Izakarn, Aouinat Torkoz, and Icht (Figure 1b) with excellent outcrop conditions (Figure 2). The structural style is quite homogeneous, and fold trends are easily recognized.

[20] We concentrated on a very competent horizon within the so-called "2nd Bani" formation of Ashgillian age. This uppermost competent unit of the Jbel Bani is labeled or6c according to *Guerraoui et al.* [1997] (Figure 3); on older maps, however, this thin marker bed is not always distinguished from the immediately underlying or6b or or6s. It consists of massive microconglomeratic quartz arenites, interpreted as being of glacial origin. This particular horizon is ideal for the reconstruction of the map-scale structures: (1) It is ubiquitous in the study area and has a relatively small thickness of 50 to 90 m [*Destombes et al.*, 1985],

which is rather thin compared to the average wavelength of the folds. (2) The massive or6c quartzites are much more resistant to weathering and erosion than the Silurian shales immediately above. The top of the or6c horizon is often very well preserved and forms the present-day surface, often in the form of dip slopes. In consequence, where the or6c-quartzites crop out, the topographic contours can directly be used as structure contours (Figure 4). Owing to very favorable outcrop conditions, large areas could be checked for outcrop-scale structures through the examination of aerial photographs and verification in the field. (3) By coincidence, the present-day valley floors in the western Anti-Atlas lie close to the inflection point of the folded Jbel Bani series. The dip at the inflection point is one of the most important pieces of information for the reconstruction of a fold, especially when considering variations of fold vergence. Outcrop conditions allow for a straightforward construction of fold closures below and above topography.

3.3. Shortening Estimates Based on Folding

[21] We constructed first a structure contour map of the top or6c surface at the 1:50,000 scale, using the most recent geologic and topographic maps available, with an equidistance of 25 m. Additional information includes our field observations, bedding dip measurements, and interpreted aerial photographs.

[22] Because of the abundance and the continuity of the outcrops we were able to reconstruct most of the fold limbs and many fold closures quickly and with high accuracy. In a second step, the structure contour lines for the eroded and subsurface parts of the folds were added using auxiliary cross sections and classical techniques of down-plunge projection along the measured fold axis direction [*Ramsay and Huber*, 1987, p. 356ff]. Since the stratigraphic thicknesses of all the Ordovician formations are very well known, and not internally deformed to any measurable degree, we were able to reconstruct the hidden parts of the folds with nearly the same level of accuracy as the outcropping limbs.

[23] The structure contour map was then transformed to a digital 3-D surface with standard digitizing procedures. For most of our further analyses, we used this 3-D surface in a TIN format, in which a network of small flat, but irregular triangles approximates the real surface. In a TIN, the size of the triangles is variable and depends, in our case, on the spacing of the structure contour lines. A TIN is ideal for representing surfaces with features at variable scales since even smaller structures are adequately represented. The error stemming from replacing a curve by straight line segments (or in 3-D: a curved surface replaced by interconnected triangles) is negligible. In our case, even smaller folds consist of numerous triangles.

[24] Using a common geographic information systems software package (ArcInfo, ArcView) allowed calculation of statistics for many of the relevant fold parameters such as amplitude, wavelength, symmetry, vergence, and variations in the direction of the fold axis. Similarly, the extraction of a section at a random point in the model and the calculation of the curve's length is a more or less automatic procedure. In this way, we calculated the map-scale shortening for a virtually unlimited number of cross sections.

[25] In a second phase, sections were extracted from the model and enlarged for those locations where outcrop-scale second-order folds had been observed in the field. These structures were then added "by hand" in order to estimate their relative contribution to overall shortening.

[26] The assessment of outcrop-scale and microscale shortening is mainly based on field observations. The fieldwork had essentially two goals: (1) an assessment of the outcrop-scale structures and the estimation of their frequency and distribution and (2) the search for macroscopic indicators of microscale deformation, their description, and the estimation of their frequency and relative importance.

[27] Outcrop-scale deformation features include meter-scale minor folds and, more frequently, minor faults with displacements of the order of centimeters to decimeters. We could hardly find any indicators for microscale deformation features within the quartzites. Axial planes of folds in the Jbel Bani quartzites are materialized through the development of systematic sets of joints [*Guiton et al.*, 2003] rather than a cleavage, let alone a schistosity. Even shaly interlayers still mostly display primary bedding and the occurrence of LPS feature in the form of a weakly developed pencil cleavage or other tectonic cleavage is rather exceptional. Generally, deformation intensity increases downward in the stratigraphic pile, and the development of cleavage is more frequently observed near the basement inliers. Higher up in the stratigraphy, within limestones of the Devonian Jbel Rich and the Carboniferous Jbel Ouarkziz, we identified layer parallel stylolite peaks as clear indicators of early, layer parallel shortening in a SSE-NNW direction, at a high angle to the strike of this monocline.

4. Results

4.1. Map-Scale Shortening

[28] The top of the Ordovician or6c horizon has been reconstructed for the entire western part of the Jbel Bani (Figures 5 and 6). Our structure model covers an area of approximately 1600 km^2. It nicely illustrates the low complexity of this folded structure. Because of favorable outcrop conditions even some of the larger outcrop-scale structures are resolved, especially in the southern quadrangle (see Figure 7 for the use of "quadrangles" and "parts"). Clearly, folding is the predominant deformation style at the map scale. A few minor tear faults appear in the northern quadrangle, but they are local phenomena, normally limited to one fold limb.

[29] Throughout the area the folds show a high lateral continuity and cylindricity. Individual fold hinges are near horizontal over several kilometers. Plunging fold hinges are generally found in pairs of en echelon lateral transitions. The overall orientation of fold axes is very constant at approximately N038°.

[30] Fold limbs near the inflection point have dips between 35° and 75° (Figure 8). Most folds, irrespective of wavelength and amplitude are symmetrical and upright.

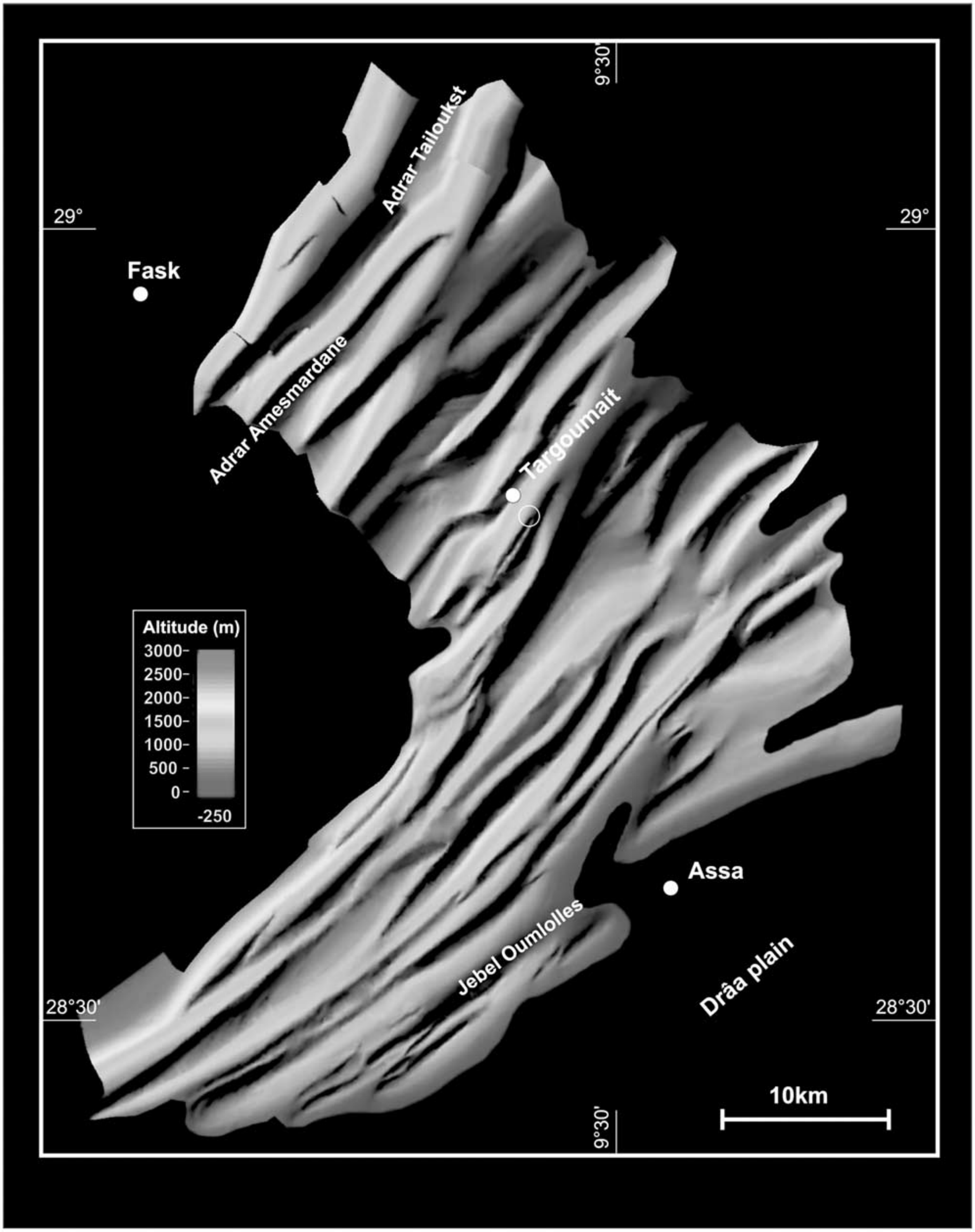

Figure 6. Three-dimensional rendering of a structure contour map constructed for the top of the or6c marker horizon in the western part of the Jbel Bani. The small white circle shows the location of the photograph of Figure 8. Note the decreasing amplitude and wavelength toward the south. See color version of this figure at back of this issue.

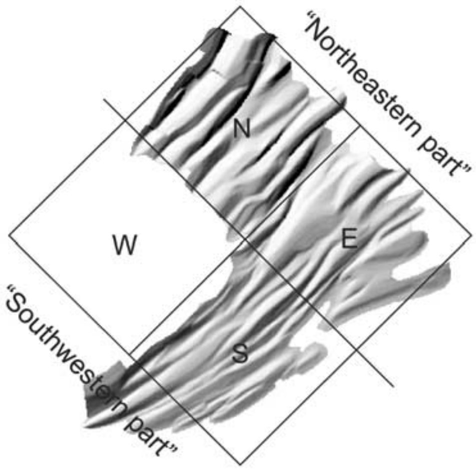

Figure 7. The location of the different areas mentioned in the text: N, northern quadrangle; E, eastern quadrangle; S, southern quadrangle; W, western quadrangle.

The differences of dip of corresponding fold limbs (at the inflexion point) is normally less than 5°. At the map scale there are no overturned limbs and there is no clear fold vergence.

[31] The wavelengths and amplitudes of the folds decrease significantly from north to south. In the northern quadrangle amplitudes of up to 1100 m are attained (northernmost anticline; ''Adrar Tailoukst'' in Figure 6), but decrease toward the eastern quadrangle to typical values of 450 to 550 m. The wavelengths for the northern quadrangle varies between 3.5 km and 4 km. Especially in the northern ''high-amplitude'' quadrangle the limb dips at the inflexion point are up to 60°. Fold limbs are rather straight and the deformation is concentrated to relatively narrow, angular hinge zones. The fold's tops are flat and large. In this quadrangle the folds are very near to the box fold endmember of detachment folds [*Namson*, 1981].

[32] The eastern and southern quadrangles are characterized by a higher variability in wavelength and amplitude. The larger and more continuous folds have amplitudes of 200 to 300 m and wavelengths of 1.5 to 2 km. The amplitudes of the smaller folds vary strongly; their wavelengths are on the order of 500 m or less. Some of these folds are at the transition to outcrop-scale structures. In the southern quadrangle the fold hinges become more rounded. The limb dips vary between 35° and 45° but they rarely exceed 40°. Only the limb dips of folds at the transition to outcrop scale attain high values up to 60°.

[33] A striking feature of the 3-D model of the Ordovician marker bed is the complete absence of thrust faults and/or duplex structures. This absence is real and not just an interpretation. Outcrop quality over large distances is excellent and allows ruling out any hidden, unresolved structures at this level. The low complexity of the structures at the map scale, essentially a continuous surface without thrusts or faults, allows for a straightforward calculation of the shortening. Sections at the borders of the model that contain gaps (because of ''holes'' of missing data) were ignored. The shortening estimates obtained from unfolding the structure contour map in different parts of the model are shown in Figure 9.

[34] The calculated shortenings correspond to the map-scale portion of the total shortening. In the southern quadrangle some structures at the transition to outcrop scale are integrated. The shortening shows considerable variability. It varies between less than 5% and more than 30%. However, if the model is subdivided into a northern ''high-amplitude domain'' (corresponding to the northern quadrangle) and a ''low-amplitude domain'' (eastern and southern quadrangles), the shortening within each of these domains is rather homogeneous. The northern quadrangle displays shortening between 12 and 20% with an average of 16.3% (Figure 9a). The eastern and the southern quadrangle together show shortening between 4% and 14% with an average of 8.7% (Figure 9b). Accordingly, the (long) sections through the whole northeastern part show shortening values between the two mentioned above. They

Figure 8. Tight syncline near the town of Targoumait. The width of the small valley is about 60 m. The or6c horizon forms the limbs of the syncline. This structure is at the transition from outcrop scale to map scale. It is well resolved in the 3-D model.

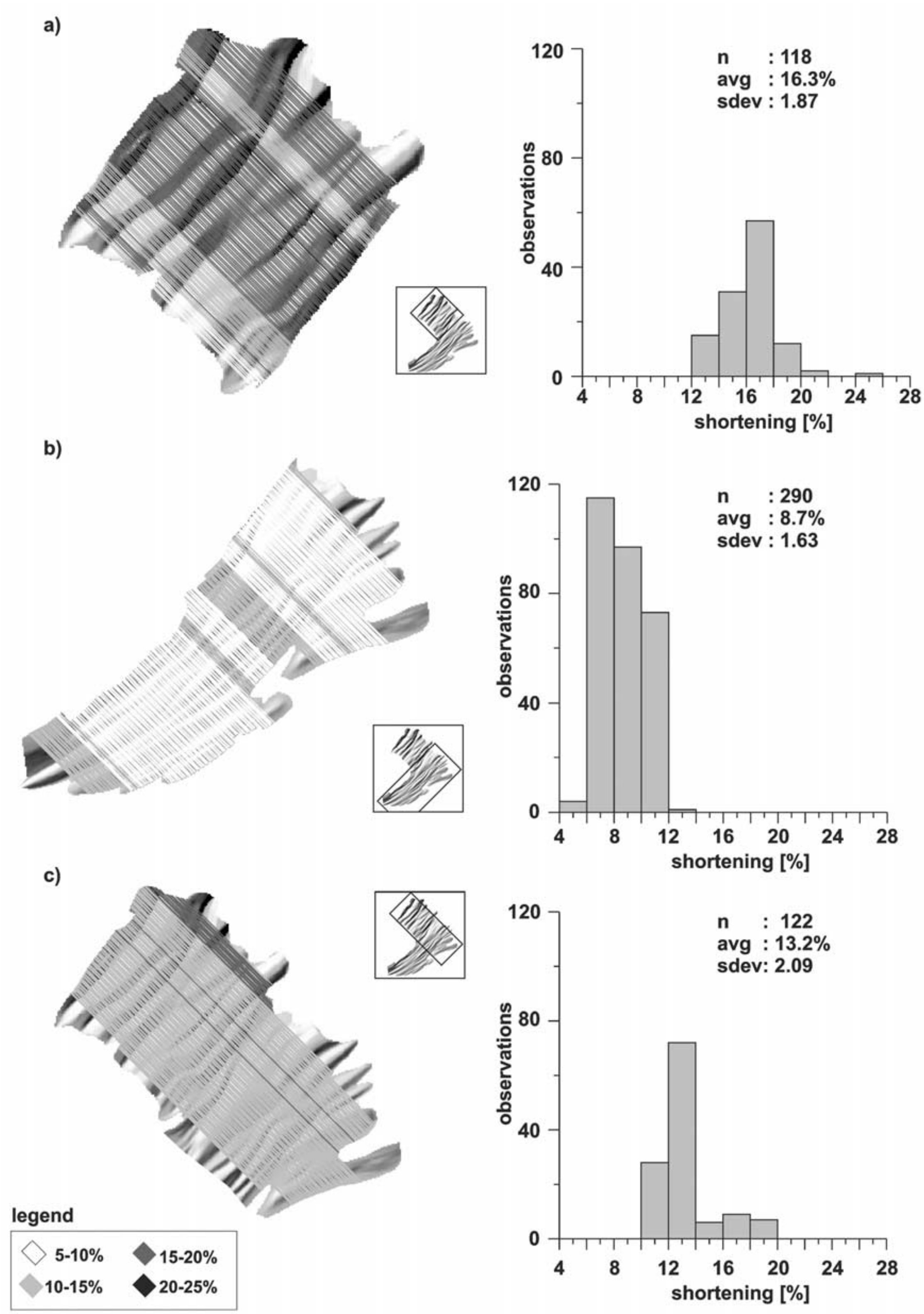

Figure 9. Map-scale shortening calculations for different parts of the top or6c-surface. Abbreviations are; n is number of sections; avg is average shortening; sdev is standard deviation.

vary between 10% and 20% with an average of 13.2% (Figure 9c).

4.2. Outcrop-Scale Shortening

[35] Field observations and the examination of aerial photographs allowed checking different structural positions for outcrop-scale structures. Also, very continuous outcrops, such as the Jbel Oumloulles Anticline west of Assa (Figure 6, southern quadrangle) allowed checking for outcrop-scale structures at several structural positions of the same fold.

[36] The most common outcrop-scale structures are wide flexures of the or6c horizon. Typically, they are characterized by very low amplitude to wavelength ratio. Shortening associated with these structures is minimal, on the order of a few meters in the case of individual, outcrop-scale structures. Their spacing is on the order of hundreds of meters to kilometers. More complex outcrop-scale structures such as tight minor folds and reverse faults are very rare. Such structures could add another few tens of meters of shortening per kilometer of cross section at the map scale. Together, outcrop-scale shortening is estimated to account for less than 5% total shortening at the map scale.

4.3. Microscale Shortening

[37] The or6c horizon exhibits no metamorphic overprint. Illite cristallinity measurement in the Ktaoua clays below and the Silurian shales above indicate that the horizon is located near the transition from diagenetic to anchizone conditions (U. Helg et al., Illite cristallinity patterns on the southeastern flank of the Variscan Anti-Atlas, Morocco, submitted to *Terra Nova*, 2004; hereinafter referred to as Helg et al., submitted manuscript, 2004) estimated at 200°C [*Frey*, 1987]. The or6c quartz arenites lack any cleavage. Rare stylolites are clearly of diagenetic origin, with peaks at high angle to bedding and no tectonic stylolites have been found within the Jbel Bani quartzites despite an intense search for such structures. Synfolding deformation is mostly by brittle fracturing and veining, most prominent in fold hinges, where this deformation is leading to some layer-parallel stretching, rather than shortening [*Guiton et al.*, 2003]. Prefolding LPS fabrics have not been identified. If anything, there might be a hidden component of layer parallel shortening in the rearrangement of the quartz grains on the truly microscopic scale [*Engelder and Marshak*, 1985]. Studies in the Appalachian Tuscarora Sandstone, a framework-supported quartz arenite, have shown that significant amounts of up to 22% horizontal shortening may be accommodated by such subtle microscale deformations [*Onasch*, 1993]. A similar study was conducted in the Tuscaroro Sandstone of the western margin of the Appalachian foreland near Keyser, West Virginia [*Harrison and Onasch*, 2000], where pressure solution together with micro fracturing and crystal plastic strain results in 10% layer parallel shortening, 10% less than the LPS values previously reported by *Onasch* [1993] from a more internal position within the Appalachian chain. Our own preliminary thin section observations made within the or6c quartzite reveal a very similar style of microscale deformations as those observed in the Appalachian Tuscarora Sandstone. In one site near Targoumait (Figure 5), we measured a grain shape preferred orientation of the detrital quartz grains of R_s = 1.15 within the bedding plane. This shortening is oriented NW-SE at a high angle to the local fold axis and it could well be interpreted as due to an early, prefolding shortening of up to 13%, if it was accommodated by pure "volume loss." In any case, LPS shortening in quartz arenites seems to be much more restricted than the highly variable values reported from shales and other clay-rich lithologies [*Kisch*, 1991].

[38] In summary, an absolute minimum horizontal shortening for the studied area, based on the map-scale balancing of folds (13%) with the addition of a conservative estimate of outcrop-scale shortening (2%), is 15%. It is more difficult to give an estimate for the most likely total tectonic shortening. It could be as large 30%, twice the minimum shortening, if microscale deformation was of the same magnitude as reported from comparable Appalachian Tuscaroro Sandstone. A more conservative estimate, based on our preliminary thin section observations, is in the range 15 to 25%.

5. Extent of the Folding

[39] In northwestern parts of the Anti-Atlas belt, Ordovician quartzites have been removed by erosion leaving only small isolated remnants, while southward, they disappear below younger Devonian and Carboniferous strata. The question therefore arises to what extent shortening estimates based on the tightly folded Jbel Bani can be extrapolated to the rest of the Anti-Atlas mountain belt.

[40] In contrast to many previous authors, we assume that similarly folded Ordovician sediments were present above at least the western basement inliers. In our interpretation, the present-day Jbel Bani is but a small erosional remnant of a much larger Ordovician sedimentary basin, preserved only in structural depressions in between basement uplifts. The present-day outline of the folded belt is by no means coincident with the original extent of the folded belt. There is no change in folding style or shortening from the northern quadrangle toward the eroded eastern quadrangle. Similarly, there is no significant decrease in shortening from the eastern toward the southern quadrangle, toward the outline of the Jbel Bani. This is also corroborated by our field observations outside of the study area: most of the mapped fold terminations along the outline of the Jbel Bani are erosional and markedly different, in terms of cylindricity, geometry, fold axis plunge, from real, conical en echelon fold terminations occasionally observed within the fold belt. Remnants of folded Ordovician series are also found outside of the Jbel Bani: The Middle Cambrian Tabanit sandstones, which are structurally coupled to the overlying Ordovician, form an open syncline at the Jbel Tayyert mountain which is the lowermost, not yet eroded part of a large, map-scale fold.

[41] Another indication for a much larger extent of the folding is the Jbel Ouarzemine mountain (JO on Figure 1b), which consists of a map-scale fold in Ordovician rocks. It is

a nearly isoclinal east vergent overturned syncline, located approximately 60 km north of the Jbel Bani. Its axis trends N021°, which is in perfect continuity to the northernmost Bani folds. On its western side it is bounded by a large thrust. Obviously, this structure accommodates more shortening than the open upright folds of the northernmost Jbel Bani, and we conclude that the observed trend of increasing shortening from southeast to northwest continues northwestward outside of our study area [*Belfoul et al.*, 2002].

[42] Folding does not stop at the southeastern border of the Jbel Bani either. A general south-southwestward decrease in deformation intensity is expected, however, and we take this into account in our "conservative" cross section where we let the Bani folds rapidly decrease in amplitude, wavelength, and shortening toward the south. A possible indication for the extent of the folding below the Drâa plain comes from the Adrar Zouggar and Addana mountains [*Desthieux*, 1977] (AZ in Figure 1b). They represent the most external Ordovician outcrops of the Anti-Atlas chain. Both of these Anticlinoria are basement cored domes as documented by drilling in the case of Adrar Zouggar [*Michard and Sougy*, 1977]. Unfortunately, this area is inaccessible for field studies at present. Detailed cross sections of the Addana and Adrar Zouggar mountains [*Desthieux*, 1977] document folding of the same style, wavelength, and "intensity" as our own observations at the southern border of the Jbel Bani.

[43] We conclude that the folded Ordovician strata initially occupied a much larger area than seen today. Folding of the Jbel Bani is not restricted to structural lows where it is preserved from erosion, but covered also large parts above the present day basement inliers. However, as seen in our model and on the basis of additional information from isolated further outcrops, both amplitude and wavelength generally decrease from northwest to southeast. This decrease is most probably related to regional-scale facies changes. The thickness of the Middle Cambrian shales, the décollement horizon, also decreases toward the southeast [*Boudda et al.*, 1979; *Benssaou and Hamoumi*, 2003], thereby progressively losing its ability as an efficient detachment horizon. Toward the southeastern front, the spacing between individual folds becomes more pronounced. This is seen as a consequence of the combined effect of a decrease in layer thickness and an increase in the ratio of competent versus incompetent beds.

[44] Another important unknown in terms of total tectonic shortening at the orogen scale is the southwestern border of the Anti-Atlas, the so-called "Jbel Ouarkziz" (OM in Figure 1b). There is strong geomorphic evidence for this monocline being the surface expression of a major triangle structure. The Ouarkziz mountain chain forms a straight ridge with altitudes of up to 500 m (200–300 m above the valley floors) over more than 300 km length, a landmark feature often photographed from space (images available at http://earthobservatory.nasa.gov/Newsroom/NewImages/images.php3?img_id=7285). This ridge is also used over a large distance as the "natural" boundary between Morocco, Algeria, and the western Sahara, respectively. There is only one road cut (Assa-Zag) at present, where the Ouarkziz chain is accessible for in situ study. In structural terms, it is essentially a large continuous monocline, dipping at $\sim 15°$ to 30° south-southeastward, forelandward into the Tindouf basin. This monocline stands in stark contrast to the tightly folded Devonian Jbel Rich underneath, from which it is separated by the lowland of the Drâa plain, running along a thick series of Upper Devonian-lower Carboniferous shales and siltstones. The Ouarkziz monocline bears all the key features of a typical triangle structure [*Banks and Warburton*, 1986]. Unfortunately, in the absence of seismic lines, we have no means to determine the additional shortening accommodated by underthrusting below this structure other than minimum estimates based on total structural relief.

6. Tectonic Implications and Discussion

6.1. Crustal-Scale Balancing

[45] Having gained some new information about the shortening in the sedimentary cover of the western Anti-Atlas, some further implications for the basement are now considered. For the illustration of our concepts a section has been chosen where the interaction between basement and sedimentary cover can be best demonstrated. This section runs across the central part the Jbel Bani (see Figure 1b), where our 3-D model has been constructed. We will concentrate first on the tectonic implications, which come from the minimal shortening recorded in the sedimentary cover. A conservative section has been constructed. For the folded Ordovician only map-scale and outcrop-scale shortening were considered. Northwest of the outcropping Jbel Bani folds, now eroded from above the basement high, the structural style is assumed to be the same geometry as within the 3-D model, i.e., open, upright folds with steep limbs. In order to account for an observed increase in thickness of the Ordovician toward the northwest [*Destombes et al.*, 1985], folds were drawn with a slightly increasing amplitude and wavelength in this direction.

[46] Southeast of the model, the Bani folds are depicted as dying out rapidly below the Drâa plain. The Jbel Ouarkziz triangle structure is interpreted as stemming from one single thrust stepping up south-southeastward from the Middle Cambrian shales to Upper Devonian shales before ending blindly somewhere below the Tindouf basin in a tip point, from which a postulated backthrust/décollement horizon must lead back up to the Earth's surface, most likely into the lower Drâa valley. We hold this backthrust responsible for the marked disharmony between the folded Jbel Rich below and the nonfolded Jbel Ouarkziz monocline above.

[47] The top of the lower Cambrian has been reconstructed using down-plunge projections and depth to detachment calculations [*Epard and Groshong*, 1993; *Bulnes and Poblet*, 1999]. In the western Anti-Atlas the Adoudounian (Lower Cambrian) consists mainly of massive platform carbonate series [*Destombes et al.*, 1985; *Benssaou and Hamoumi*, 2003]. Incompetent horizons are scarce and sediments up to the Middle Cambrian are strongly coupled

to the basement (compare Figures 2 and 3). A décollement within these sediments, at least in the western Anti-Atlas, is very unlikely and not corroborated by any structural observations. Therefore the whole pile from the basement up to the top of the Lower Cambrian can be considered as one structural unit, the ‘‘basement-Cambrian’’ unit in our rheologic terminology.

[48] As seen on most classical cross sections of the Anti-Atlas and confirmed by our own field observations, the folded Ordovician is clearly decoupled from this basement-Cambrian unit. Folding is not restricted to structural lows between the basement inliers, however, and entire fold trains can be projected sideways above the larger basement domes. This also means that Bani folding at any given location cannot be linked to one specific (assumed) thrust fault in the basement. Shortening in the Ordovician needs to be compared to the overall shortening on a much larger scale, so as to include the entire basement-cover contact. In order to balance the section at the crustal scale, an equal amount of shortening is necessary in the sedimentary cover and the basement. Shortening within the basement is assumed to take place along widely spaced, mostly hidden large-scale reverse faults. Geometric constraints will be discussed below.

6.2. Thin Skin Interpretation

[49] With the Ordovician strongly decoupled from the basement-Cambrian unit and a shortening increasing toward the northwest, the question arises to which degree the observed folding could be the result of a thin skin thrusting-folding phase, predating the formation of the basement uplifts. The westernmost part of the Anti-Atlas, west of the Bas Drâa inlier, is known as a thin skin domain indeed [*Belfoul*, 1991; *Soulaimani*, 1998; *Piqué*, 2001; *Belfoul et al.*, 2002]. It is characterized by map-scale thrusts and reverse faults within Lower Cambrian units. Accordingly, shortening observed in the Ordovician Jbel Bani might be rooted either here or in even more internal positions to the west, currently hidden in the Moroccan offshore. We consider such a distant source for the shortening of the Jbel Bani as very unlikely, however. Such a thin skin interpretation still requires two new postulates, which are not substantiated by any observations. First, a regional-scale basal décollement within Middle Cambrian shales would have to be increasingly important westward, prior to cutting down into the basement-Cambrian unit, in locations where neither direct nor indirect evidence for any major basement ramp is present. Secondly, the basement uplifts of the SW Anti-Atlas would have to be explained by a second deformation phase, in an essentially ‘‘vertical tectonics’’ fashion, without any sizeable amount of horizontal shortening. This latter view of the Anti-Atlas basement uplifts does not provide any explanation for the observed crustal-scale thickening.

[50] The fold axis directions within the Adoudounian of the western Anti-Atlas show considerable heterogeneity [*Soulaimani*, 1998]. In many places, they parallel in a conspicuous manner the outlines of the basement domes. This results in places in fold interference patterns with nearly perpendicular fold axes [*Caritg et al.*, 2003]. Most authors agree that the intimate relationship of fold axes directions and basement outlines is due to ‘‘forced’’ folding of the sedimentary cover, accommodating differential movements of basement blocks. It is difficult to envisage these complicated fold pattern as the result of an early, thin skin folding phase. Thus, at least in the Adoudounian of the western Anti-Atlas, there exists no evidence for folding independent of the main deformation phase, which is clearly thick skinned, involving basement. The tectonic style of these folds at the lowermost stratigraphic levels is very similar to the Rocky Mountains of Wyoming [*Stone*, 1993, 2002].

[51] Higher up in the stratigraphy, however, within the Ordovician Jbel Bani, fold axes directions are very regular and increasingly independent of the underlying basement structures. Some minor variations in fold axis directions can still be explained as deflections around the basement inliers, albeit strongly attenuated above the Lie de vin detachment level. No polyphase deformation history is required to explain these variations in fold axis trends in the western Anti-Atlas. Note that further east, near Tata, there is evidence for a two-phase history in the form of conspicuous fold interference patterns [*Caritg et al.*, 2003], but even there, both folding phases are best explained in a thick skinned fashion, because of reactivation of former basement faults, in close analogy with Laramide uplifts of the Rockies [*Stone*, 1993; *Marshak et al.*, 2000; *Bump*, 2003].

[52] Paleostress measurements in the western Anti-Atlas fold belt reveal that the local axes of major compressive stress are subperpendicular to the fold axes throughout the Jbel Bani. The paleostress field also shows some heterogeneity and it is characterized by a predominant local component. It seems therefore that the paleostress field is also conditioned by the basement blocks rather than being the result of a distant push. In conclusion, an early thin skin phase is regarded as a very improbable mechanism to explain the folding of the Ordovician Jbel Bani.

6.3. Thick Skin Interpretation

[53] A rapid glance at the tectonic map of the SW Anti-Atlas reveals the existence of ‘‘thick skin’’ tectonics, necessary to explain the large domes of Proterozoic-Cambrian [*Rodgers*, 1995]. The central question is not the existence, but the nature and origin of these basement domes, as well as their relationship with the buckle folding observed in the cover series around and above them. Direct evidence for large-scale thrusts within the basement-Cambrian unit is elusive in the western Anti-Atlas. This lack of direct evidence should not be used as an argument against the existence of thrusts or reverse faults at depth, however. There is a series of explanations why such thrusts have not been mapped, nor postulated. The most important reason is the present-day erosion level, revealing only a very small, topmost 2 km of more than 10 km of structural relief of these basement uplifts. This estimate is based on the known depth to basement in the distant foreland of the Tindouf basin, where it is located at approximately -8 km [*Michard*, 1976; *Bertrand-Sarfati et al.*, 1991]. Similarly, the preshortening

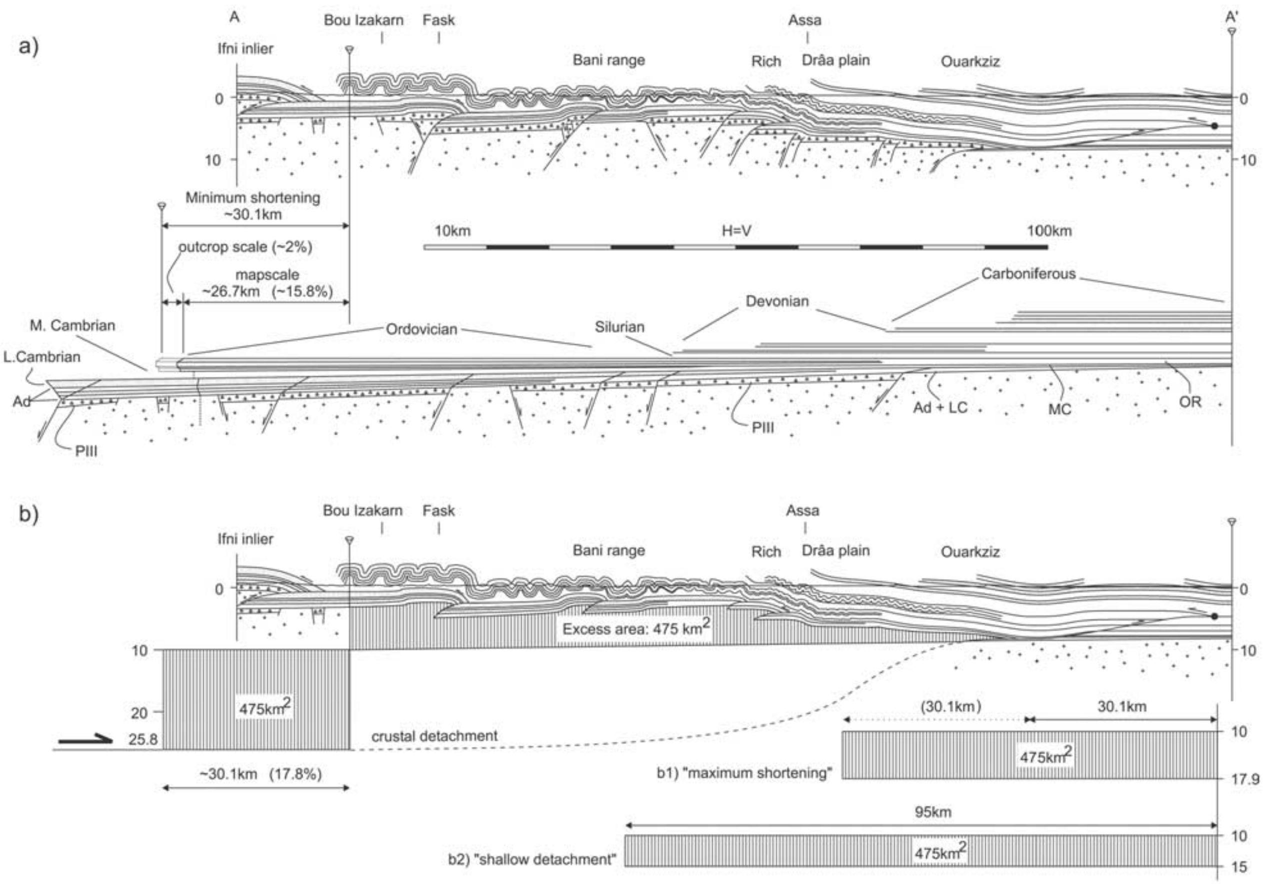

Figure 10. (a) Large-scale cross section through the entire western Anti-Atlas and corresponding unfolded section, with a conservatively estimated minimum horizontal shortening of 30 km. Abbreviations are as follows: Ad, Adoudounian; LC, Lower Cambrian; MC, Middle Cambrian; OR, Ordovician. Stippled area shows the most competent units (mainly quartz arenites, sandstones, and conglomerates). (b) Visualization of the "excess area" concept used to estimate the depth to an inferred middle to lower crustal detachment. The cross-sectional area calculations are based on the assumption of an initially 10 km deep sedimentary basin, necessary to accommodate the thickness of the Paleozoic series of the Anti-Atlas chain. A maximum estimate of shortening of the order of 60 km (30 km folding + 15% LPS) requires a midcrustal detachment at about 18 km depth (labeled b1). A shallow detachment at 15 km depth leads to an unreasonably high shortening of 95 km (labeled b2).

top basement level in the southwestern Anti-Atlas region must have been at some −10 km below sea level in order to accommodate the total thickness of Paleozoic sediments, which comprise shallow marine series up to the middle Carboniferous (Visean).

[54] Another reason why no large-scale thrusts were found in the SW Anti-Atlas is the presence of widespread thick Middle Cambrian shales. As depicted in Figure 10, we assume a tectonic style where thrust ramps, breaking through the Lower Cambrian Adoudounian, turn to horizontal "flats" within incompetent Middle Cambrian shales. Minor thrusts may level out even deeper in the stratigraphy, within thin evaporitic shales of the Adoudounian unit.

[55] An important reverse fault or thrust has been described in the westernmost Anti-Atlas, at the southern border of the Lower Drâa inlier, where Proterozoic basement is thrust toward the SE onto the sedimentary cover [*Soulaimani*, 1998]. Here, intense deformation under lower greenschist facies lead to the development of a foliation with steeply plunging stretching lineations and a marked internal deformation of rhyolite pebbles of the PIII; all these deformation features are compatible with a south-southeastward thrust.

[56] At the scale of the entire Anti-Atlas, crustal-scale balancing considerations are used to constrain permissible basement geometries at depth. One of the key features of the Anti-Atlas is its high structural relief. A conservative estimate of the total thickness of the Paleozoic cover series of the SW Anti-Atlas yields 10 km thickness. Most of this thickness is preserved today in first-order synclinoria, in between the basement domes. However, there is good evidence for a similar amount of erosion from atop the westernmost basement inliers (Bas Drâa, Ifni, Kerdous), which were invariably at lower greenschist facies metamor-

phic conditions during late Variscan deformation. This metamorphism is obvious from the deformation style at the brittle/ductile transition and from our own illite crystallinity measurements [*Buggisch*, 1988a; *Burkhard et al.*, 2001; Helg et al., submitted manuscript, 2004]. In contrast to previous authors, who mostly considered these basement domes as all-time "horsts," with a reduced thickness of total Paleozoic cover, we assume an essentially continuous thickness of sedimentary series across these basement domes. We therefore assume a similar structural relief to exist at the basement cover contact, where it is documented, and at the Devonian/Carboniferous boundary, where it is mostly based on indirect evidence.

[57] The present-day topography and average elevation of the Anti-Atlas could be at least partly due to Neogene reactivation in relation with High-Atlas inversion tectonics [*Beauchamp et al.*, 1999; *Frizon-de-Lamotte et al.*, 2000]. The Anti-Atlas is also the location of Pliocene volcanic activity, e.g., in the Jbel Siroua, and broad thermal uplift at a regional scale has to be taken into consideration. Accordingly, the removal of more than 10 km of Paleozoic sediments from above the basement inliers may not be solely the result of late Variscan tectonics but the combined effect of late Variscan tectonics, Neogene Atlas inversion, and thermal hot spot related uplifts and concomitant erosion. The total amount of non-Variscan uplift/erosion in the Anti-Atlas is ill constrained. Upper Cretaceous sediments of the "Hamada" clearly seal Variscan folds; their present-day elevation is about 400 m above sea level in the eastern Anti-Atlas but close to sea level near the Atlantic coast. In between, this same marker horizon is completely eroded from above the SW Anti-Atlas, leaving some room for speculation. Thermal uplift, related to the volcanic activity of the Jbel Sirhoua, can only be estimated from general considerations. Surface uplift related to plates moving over hot spots may reach 1 to 3 km, affecting large areas of up to 2000 km across [*Coward*, 1994; *Eisbacher*, 1996].

[58] Nevertheless, even with the assumption of up to 3 km of Neogene surface uplift and High Atlas reactivation, the leveling of more than 7 km of structural relief remains to be explained by late Variscan tectonics with concomitant uplift and erosion. This same amount of structural relief is not only attained at the scale of the orogen but also at the scale of individual basement domes. Between the Jbel Bani and the surrounding basement inliers Kerdous, Lower Drâa, and Ifni (Figure 2), the structural relief is in excess of 7 km. Given the modest amount of cover shortening observed in the Anti-Atlas fold belt, this structural relief is difficult to explain by an antiformal stack of thin basement slabs above a relatively shallow crustal detachment [*Boyer and Elliott*, 1982]. As an example, if we assume 5 km thick basement slabs overlain by a 10 km thick sedimentary series, a horizontal shortening of roughly 95 km would be necessary to create the observed cross sectional excess area of 475 km^2 (Figure 10b, "shallow detachment" diagram labeled b2). This exceeds by far our estimations of the horizontal shortening observed within the sedimentary cover. The discrepancy is too large to be explained by ill evaluated, underestimated outcrop-scale or microscale shortening. Even if we assume a total submap-scale shortening of the same magnitude as map-scale shortening, the calculated depth to detachment remains at about 18 km (Figure 10b, "maximum shortening" diagram labeled b1). Considering only the minimal shortening (the total observed shortening), the detachment depth comes to lie at about 26 km depth.

[59] As an alternative to the antiformal stack of thin basement slabs we postulate basement inversion of thick block-like basement parts with the dimension of the present day boutonnieres, uplifted along steep reverse faults. The depth of the detachment horizon again determines the thickness of such basement blocks.

6.4. Inversion Tectonics

[60] The term inversion is used here in the sense of *Coward* [1994] to describe regions which have experienced a reversal in uplift or subsidence. The Anti-Atlas nicely fits this definition; after an extended period of subsidence during the Paleozoic, the Anti-Atlas basin is the locus of tectonic shortening, uplift, and erosion during the later stages of the Variscan orogeny, possibly with some reactivation in Neogene times [*Frizon-de-Lamotte et al.*, 2000].

[61] Most authors agree that Variscan basements tectonics of the Anti-Atlas is dominated by the reactivation of former weakness zones. In earlier publications these weakness zones are invariably considered to be subvertical [*Michard*, 1976]. Sedimentological evidence for an important post-Panafrican extensional phase has accumulated during the last two decades [*Soulaimani et al.*, 2001; *Benssaou and Hamoumi*, 2003]. Decametric fault-bounded graben structures are documented within the Proterozoic basement in the westernmost Anti-Atlas [*Soulaimani*, 1998]. Regional changes in thickness and facies of Cambrian series have been used to map the location and general orientation of graben structures [*Benssaou and Hamoumi*, 2003] (Figure 11).

[62] In the light of this post Panafrican rifting phase a Variscan inversion of former PIII normal faults seems the most likely explanation for the mostly still hidden reverse faults bounding the present day basement inliers. Rather than all-time horsts [*Michard*, 1976; *Piqué*, 1994; *Benssaou and Hamoumi*, 2003], basement inliers in our interpretation represent the former basin floors. The fault reactivation nicely explains the strong basement involvement and the high basement topography in an otherwise only mildly deformed tectonic setting.

[63] Tectonic inversion is, at least partly, corroborated by the thickness distribution of the rift/early postrift sediments [*Boudda et al.*, 1979]. The thickness of the lowermost Adoudounian unit, also called "Lower Limestones" (Figure 11), is highest in the area of the basement domes of the western Anti-Atlas, decreasing both east and southward, forelandward. In other words, the location of maximum subsidence by rifting corresponds roughly with the area of maximum uplift by late Variscan tectonic inversion.

[64] The widespread occurrence of PIII rift sediments found along the borders of almost all boutonnieres is an additional indication for these to be former grabens, rather

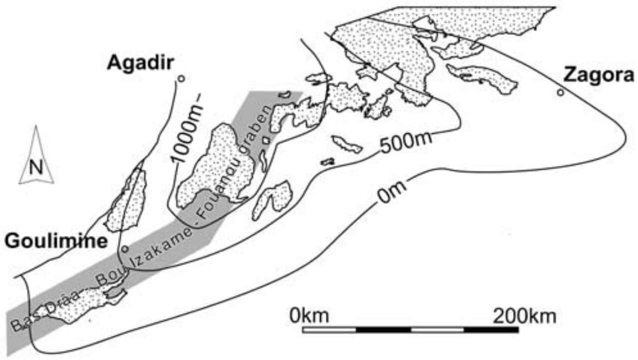

Figure 11. Thickness distribution of the lowermost sedimentary unit of the Paleozoic Anti-Atlas basin according to *Boudda et al.* [1979]. The Bas Drâa-Bou Izakarne-Fouanou graben structure of Middle Cambrian age is shown according to a recent detailed stratigraphic analysis of *Benssaou and Hamoumi* [2003, Figure 2].

than horsts. We have to admit, however, that there is no observational control, nor is there any subsurface data available about the regional-scale variations of PIII series for the large "synclinal" areas in between the present day basement domes. According to our inversion hypothesis, there should be only a very limited amount of PIII series in these areas of the former, Late Proterozoic horsts.

7. Conclusions

[65] The Anti-Atlas of Morocco represents a special type of foreland fold belt with a striking absence of observable thrusts. Cover shortening is accommodated by tight detachment folding of competent quartzite and carbonate layers, but no ramping or duplexing has ever been observed. This tectonic style is dictated by the mechanical stratigraphy of the Paleozoic cover series, dominated by thick intervals of shale and siltstones. Four competent units are distinguished (Figure 3), from bottom to top: (1) basement-Lower Cambrian, (2) Jbel Bani (Ordovician), (3) Jbel Rich (Devonian), and (4) Jbel Ouarkziz (lower Carboniferous). The basement-Lower Cambrian unit (1) is involved in massive inversion of Late Proterozoic normal faults, leading to Laramide style basement domes of several tens of kilometers wavelength, with "amplitudes" or structural relief on the order of 7 to 10 km: the "first-order" basement folds or so-called "plis de fond" [*Argand*, 1924]. A more than 1 km thick series of Middle Cambrian shales allows for a complete decoupling and disharmony between this basement unit (1) and the overlying Paleozoic cover series (2 and 3), which react essentially by polyharmonic multilayer folding [*Ramsay and Huber*, 1987]. Larger wavelengths (1 km) and amplitudes (500 m) are observed within folds of the thicker Ordovician sandstones of the Jebel Bani than within the Devonian Jbel Rich folds (500 m, 200 m). Carboniferous series (4) are only preserved at the very mountain front, where they are involved in a triangle structure. The total width of the SW Anti-Atlas foreland belt is about 150 km and basement domes occur at less than 30 km behind the mountain front. A minimum total amount of 30 km (17%) of cover shortening is obtained from restoration of the Jbel Bani fold train. Balancing considerations require a middle to lower crustal detachment at 18 to 25 km depth for the basement inversion.

[66] **Acknowledgments.** We wish to thank A. Soulaimani for introducing us to the geology of the Anti-Atlas, his hospitality, and for many inspiring discussions. Ongoing support by the Moroccan "Ministère de l'Energie et des Mines, Direction de la Géologie" and the state petroleum agency ONAREP is gratefully acknowledged. We wish to thank personally M. Dahmani, A. M. Morabet, and M. Zizi for their help. We are grateful to Bert Bally, André Michard, Alain Piqué, Jean-Paul Schaer, and Dieter Schroeder for encouragement and for sharing their ideas and data about the Anti-Atlas of Morocco. Financial support by the Swiss National Science Foundation, grants 21-52516.7 and 20-63790.00 is gratefully acknowledged.

References

Argand, E. (1924), La tectonique de l'Asie, *Proc. Int. Geol. Congr.*, *13th*, 171–372.

Azizi-Samir, M. R., J. Ferrandini, and J. L. Tane (1990), Late Pan-African (580–560 M.A.) tectonic activity and volcanism in the Anti-Atlas Mountains of Morocco: Geodynamic interpretation on the scale of north-west Africa, *J. Afr. Earth Sci.*, *10*, 549–563.

Banks, C. J., and J. Warburton (1986), "Passive-roof" duplex geometry in the frontal structures of the Kirthar and Sulaiman mountain belts, Pakistan, *J. Struct. Geol.*, *8*, 229–237.

Beauchamp, W., R. W. Allmendinger, M. Barazangi, A. Demnati, M. Alji, and M. Dahmani (1999), Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological-geophysical transect, *Tectonics*, *18*, 163–184.

Belfoul, M. (1991), La structuration hercynienne de la région de Labiar-Oued Boussafene (Guelmim, Anti-Atlas sud-occidental), thèse 3ème cycle thesis, 162 pp., Univ. de Rabat, Rabat, Morocco.

Belfoul, M. A., F. Faik, and B. Hassenforder (2002), Evidence of a tangential tectonic event prior to the major folding in the Variscan belt of the western Anti-Atlas, Morocco, *J. Afr. Earth Sci.*, *32*, 723–739.

Benssaou, M., and N. Hamoumi (2003), The Lower-Cambrian western Anti-Atlasic graben: Tectonic control of paleogeography and sequential organisation, *C. R. Geosci.*, *335*, 297–305.

Bertrand-Sarfati, J. (1981), Problème de la limite Précambrien-Cambrien: La section de Tiout (Maroc): Les stromatolithes et leur biostratigraphie, *Newsl. Stratigr.*, *10*, 20–26.

Bertrand-Sarfati, J., A. Moussine-Pouchkine, P. Affaton, R. Trompette, and Y. Bellion (1991), Cover sequences of the West African Craton, in *The West African Orogens and Circum Atlantic Correlatives*, edited by R. D. Dallmeyer and J. P. Lecorche, pp. 65–82, Springer-Verlag, New York.

Bonhomme, M., and B. Hassenforder (1985), Le métamorphisme hercynien dans les formations tardi-et post-panafricaines de l'Anti-Atlas occidental (Maroc): Données isotopiques Rb/Sr et K/Ar des fractions fines, *Sci. Geol. Bull.*, *38*, 175–183.

Boudda, A., G. Choubert, and A. Faure-Muret (1979), Essai de stratigraphie de la couverture sédimentaire de l'Anti-Atlas: Adoudounien - Cambrien Inférieur, *Notes Mem. Serv. Geol. Maroc*, *271*, 96.

Boyer, S. E., and D. Elliott (1982), Thrust systems, *AAPG Bull*, *66*, 1196–1230.

Buggisch, W. (1988a), Diagenesis and very low-grade metamorphism of the Lower Cambrian rocks in the Anti-Atlas, Morocco, in *The Atlas System of Morocco: Studies on Its Geodynamic Evolution*, *Lecture Notes in Earth Sci.*, vol. 15, edited by Volker H. Jacobshagen, pp. 123–128, Springer-Verlag, New York.

Buggisch, W. (1988b), The Precambrian-Cambrian boundary in the Anti-Atlas (Morocco), in *The Atlas System of Morocco: Studies on Its Geodynamic Evolution*, *Lecture Notes in Earth Sci.*, vol. 15, edited by Volker H. Jacobshagen, pp. 81–90, Springer Verlag, New York.

Bulnes, M., and J. Poblet (1999), Estimating the detachment depth in cross sections involving detachment folds, *Geol. Mag.*, *136*, 395–412.

Bump, A. P. (2003), Reactivation, trishear modeling, and folded basement in Laramide uplifts: Implications for the origins of intra-continental faults, *GSA Today*, *13*(3), 4–10.

Burkhard, M., S. Caritg, U. Helg, and C. Robert-Charrue (2001), Forced, disharmonic multilayer buckle folding in the late Variscan Antiatlas of Morocco, paper presented at American Association of Petroleum Geologists 2001 Annual Meeting, Tulsa, Okla.

Caritg, S., M. Burkhard, R. Ducommmun, U. Helg, L. Kopp, and C. Sue (2003), Fold interference patterns in the late Paleozoic Anti-Atlas of Morocco, *Terra Nova*, *16*, 27–37.

Choubert, G. (1963), Histoire géologique du précambrien de l'Anti-Atlas, *Notes Mem. Serv. Geol. Maroc*, *162*, 352.

Choubert, G., and A. Faure Muret (1971), Anti-Atlas (Maroc), in *Tectonique de l'Afrique* (*Tectonics of Africa*), *Earth Sci. Ser.*, vol. 6, pp. 163–175, UNESCO, Paris.

Coward, M. (1994), Inversion Tectonics, in *Continental Deformation*, edited by P. L. Hancock, pp. 289–304, Pergamon, New York.

Desthieux, F. (1977), Etude tectonique et métallogénique du Jbel Addana, Ordovicien des plaines du Dra. Maroc présaharien, *Notes Serv. Geol. Maroc*, *268*, 209–236.

Destombes, J., H. Hollard, and S. Willefert (1985), Lower Paleozoic rocks of Morocco, in *Lower Palaeozoic of North-Western and West-Central Africa*, vol. 4, edited by C. H. Holland, pp. 91–336, Trinity Coll., Dep. of Geol., Dublin, Ireland.

Donzeau, M. (1974), L'Arc Anti-Atlas-Ougarta (Sahara nord-occidental, Algérie-Maroc), *C. R. Acad. Sci., Ser. II*, *278*, 417–420.

Eisbacher, G. H. (1996), *Einführung in die Tektonik*, 374 pp., F. Enke, Stuttgart, Germany.

Engelder, T., and S. Marshak (1985), Disjunctive cleavage formed at shallow depths in sedimentary rocks, *J. Struct. Geol.*, *7*, 327–343.

Epard, J. L., and R. H. J. Groshong (1993), Excess area and depth to detachment, *AAPG Bull.*, *77*, 1291–1302.

Faik, F., M. A. Belfoul, M. Bouabdelli, and B. Hassenforder (2002), The structures of the Late Neoproterozoic and Early Palaeozoic cover of the Tata area, western Anti-Atlas, Morocco: Polyphased deformation or basement/cover interactions during the Variscan orogeny?, *J. Afr. Earth Sci.*, *32*, 765–776.

Frey, M. (1987), *Low Temperature Metamorphism*, Chapman and Hall, New York.

Frizon-de-Lamotte, D., B. Saint-Bezar, R. Bracene, and E. Mercier (2000), The two main steps of the Atlas building and geodynamics of the western Mediterranean, *Tectonics*, *19*, 740–761.

Geyer, G., and E. Landing (1995), *MOROCCO '95—The Lower-Middle Cambrian Standard of Western Gondwana*, 269 pp., Inst. für Paläontol., Univ. Würzburg, Würzburg, Germany.

Guerraoui, A., M. Bensaid, and M. Dahmani (1997), Carte géologique du Maroc 1:100'000, Fask, *Notes Serv. Geol. Maroc*.

Guiton, M. L. E., W. Sassi, Y. M. Leroy, and B. D. M. Gauthier (2003), Mechanical constraints on the chronology of fracture activation in folded Devonian sandstone of the western Moroccan Anti-Atlas, *J. Struct. Geol.*, *25*, 1317–1330.

Harrison, M. J., and C. M. Onasch (2000), Quantitative assessment of low-temperature deformation mechanisms in a folded quartz arenite, Valley and Ridge Province, West Virginia, *Tectonophysics*, *317*, 73–91.

Hassenforder, B. (1987), La tectonique panafricaine et varisque de l'Anti Atlas dans le Massif du Kerdous (Maroc), Thèse Sci. thesis, 249 pp., Univ. L. Pasteur Strasbourg, Strasbourg, France.

Hogan, J. P., and W. M. Dunne (2001), Calculation of shortening due to outcrop-scale deformation and its relation to regional deformation patterns, *J. Struct. Geol.*, *23*, 1507–1529.

Homberg, C. (1996), Analyse des déformations cassantes dans le Jura et modélisation numérique des perturbations des contraintes tectoniques autour d'accidents majeurs, Ph.D. thesis, 282 pp., Univ. Pierre et Marie Curie, Paris.

Huon, S. (1985), Clivage ardoisier et réhomogénisation isotopique K-Ar dans des schistes paléozoiques du Maroc: Etude microstructurale et isotopique: Conséquences régionales, Ph.D. thesis, 124 pp., Univ. Louis Pasteur, Strasbourg.

Jeannette, D. (1981), Influence de structures panafricaines sur des déformations hercyniennes (Bordure septentrionale de la boutonnière d'Igherm, Anti-Atlas occidental), *Notes Serv. Geol. Maroc*, *42*, 25–32.

Kisch, H. J. (1991), Development of slaty cleavage and degree of very-low-grade metamorphism: A review, *J. Metamorph. Geol.*, *9*, 735–750.

Kogbe, C. A. (1998), Sedimentary Basins of Morocco, *Afr. Geosci. Rev.*, *5*, 232.

Latham, A. (1988), The Precambrian-Cambrian transition in Morocco: Problems and perspectives, in *Trace Fossils, Small Shelly Fossils, and the Precambrian-Cambrian Boundary*, edited by E. Landon, G. M. Narbonne, and P. M. Myrow, *Bull. N. Y. State Mus.*, *1076*, 14 pp.

Leblanc, M. (1972), Sur le style disharmonique des plis hercyniens, à la base de la couverture, dans l'Anti-Atlas central (Maroc), *C. R. Acad. Sci., Ser. D*, *275*, 803–806.

Leblanc, M., and J. R. Lancelot (1980), Interpretation geodynamique du domaine pan-africain (Precambrien terminal) de l'Anti-Atlas (Maroc) a partir de donnees geologiques et geochronologiques, *Can. J. Earth Sci.*, *17*, 142–155.

Marshak, S., K. Karlstrom, and J. M. Timmons (2000), Inversion of Proterozoic extensional faults: An explanation for the pattern of Laramide and Ancestral Rockies intracratonic deformation, United States, *Geology*, *28*, 735–738.

Mattauer, M., F. Proust, and P. Tapponnier (1972), Major strike-slip fault of late Hercynian age in Morocco, *Nature*, *237*, 160–162.

Michard, A. (1976), Eléments de géologie Marocaine, *Notes Mem. Serv. Geol. Maroc*, *252*, 408.

Michard, A., and J. Sougy (1977), L'orogenèse hercynienne à la lisière nord-ouest de l'Afrique (Structure des chaînes primaires du Maroc au Sénégal), in *La Chaîne varisque d'Europe moyenne et occidentale*, *Colloq. Int. Cent. Natl. Rech. Sci.*, vol. 243, pp. 605–640, Cent. Natl. de la Rech. Sci., Paris.

Mitra, G. (1994), Strain variation in thrust sheets and across the Sevier fold-and-thrust belt (Idaho-Utah-Wyoming): Implications for section restoration and wedge taper evolution, *J. Struct. Geol.*, *16*, 585–602.

Mitra, S. (1992), Balanced structural interpretations in fold and thrust belts, in *Structural Geology of Fold and Thrust Belts*, edited by S. Mitra and G. W. Fisher, pp. 53–77, Johns Hopkins Univ. Press, Baltimore, Md.

Namson, J. (1981), Structure of the western Foothills Belt, Miaoli-Hsinchu area Taiwan: (1) Southern Part, *Pet. Geol. Taiwan*, *18*, 31–51.

Onasch, C. M. (1993), Determination of pressure solution shortening in sandstones, *Tectonophysics*, *227*, 145–159.

Piqué, A. (1994), *Géologie du Maroc: Les Domaines Structuraux et Leur Évolution Structurale*, 283 pp., Ed. PUMAG, Marrakech, Morocco.

Piqué, A. (2001), *Geology of Northwest Africa*, 310 pp., Gebrüder Bornträger, Stuttgart, Germany.

Piqué, A., and A. Michard (1989), Moroccan hercynides: A synopsis: The paleozoic sedimentary and tectonic evolution at the northern margin of West Africa, *Am. J. Sci.*, *289*, 286–330.

Piqué, A., J. J. Cornée, J. Müller, and J. Roussel (1991), The Moroccan Hercynides, in *The West African Orogens and Circum Atlantic Correlatives*, edited by R. D. Dallmeyer and J. P. Lecorche, pp. 229–263, Springer-Verlag, New York.

Piqué, A., M. Bouabdelli, and J.-R. Darboux (1995), Le rift cambrien du Maroc occidental, *C. R. Acad. Sci., Ser. II*, *320*, 1017–1024.

Piqué, A., M. Bouabdelli, A. Soulaimani, N. Youbi, and M. Iliani (1999), Upper Neoproterozoic PIII conglomerates in the Anti-Atlas, southern Morocco: Pan-African molasses, or indicators of upper Proterozoic rifting?, *C. R. Acad. Sci., Ser. II*, *328*, 409–414.

Plessmann, W. (1972), Horizontal-Stylolithen im französisch-schweizerischen Tafel-und faltenjura und ihre Einpassung in den regionalen Rahmen, *Geol. Rundsch.*, *61*, 332–347.

Ramsay, J. G., and I. M. Huber (1983), *The Techniques of Modern Structural Geology*, 307 pp., Academic, San Diego, Calif.

Ramsay, J. G., and I. M. Huber (1987), *The Techniques of Modern Structural Geology*, 697 pp., Academic, San Diego, Calif.

Rodgers, J. (1990), Fold-and-thrust belts in sedimentary rocks, part 1: Typical examples, *Am. J. Sci.*, *290*, 321–359.

Rodgers, J. (1995), Lines of basement uplifts within the external parts of orogenic belts, *Am. J. Sci.*, *295*, 455–487.

Schenk, P. E. (1971), Southeastern Atlantic Canada, northwestern Africa, and continental drift, *Can. J. Earth Sci.*, *8*, 1218–1251.

Sebai, A., G. Feraud, H. Bertrand, and J. Hanes (1991), AR-40/AR-39 dating and geochemistry of tholeiitic magmatism related to the early opening of the central Atlantic rift, *Earth Planet. Sci. Lett.*, *104*, 455–472.

Smart, K. J., W. M. Dunne, and R. D. Krieg (1997), Roof sequence response to emplacement of the Wills Mountain duplex: The roles of forethrusting and scales of deformation, *J. Struct. Geol.*, *19*, 1443–1459.

Soulaimani, A. (1998), Dynamique et interactions Socle/Couverture dans l'Anti-Atlas occidental (Maroc): Rifting fini-protérozoïque et orogenèse hercynienne, Thèse d'état, Univ. Cadi Ayyad, Marrakech, Morocco.

Soulaimani, A., C. Le Corre, and R. Farazdaq (1997), Deformation hercynienne et relation socle/couverture dans le domaine du Bas-Draa (Anti-Atlas occidental, Maroc), *J. Afr. Earth Sci.*, *24*, 271–284.

Soulaimani, A., A. Piqué, and M. Bouabdelli (2001), La série du PII–PIII de l'Anti-Atlas occidental (Sud marocain): Un olistostrome à la base de la couverture post-panafricaine (PIII) du Protérozoic supérieur, *Earth Planet. Sci. Lett.*, *332*, 121–127.

Soulaimani, A., M. Bouabdelli, and A. Piqué (2003), L'extension continentale au Néo-Protérozoïque supérieur-Cambrien inférieur dans l'Anti-Atlas (Maroc), *Bull. Soc. Geol. Fr.*, *174*, 83–92.

Stone, D. (1993), Basement-involved thrust-generated folds as seismically imaged in the subsurface of the central Rocky Mountain foreland, in *Laramide Basement Deformation in the Rocky Mountain Foreland of the Western United States*, edited by R. Chase, E. Erlsev, and C. Schmidt, *Spec. Pap. Geol. Soc. Am.*, *280*, 271–318.

Stone, D. (2002), Morphology of the Casper Mountain uplift and related subsidiary structures, central Wyoming: Implications for Laramide kinematics, dynamics, and crustal inheritance, *AAPG Bull.*, *86*, 1417–1440.

Villa, I. (1998), Isotopic closure, *Terra Nova*, *10*, 42–47.

Villeneuve, M., and J. J. Cornee (1994), Structure, evolution and palaeogeography of the West African Craton and bordering belts during the Neoproterozoic, *Precambrian Res.*, *69*, 307–326.

Weijermars, R. (1993), Estimation of paleostress orientation within deformation zones between two mobile plates, *Geol. Soc. Am. Bull.*, *105*, 1491–1510.

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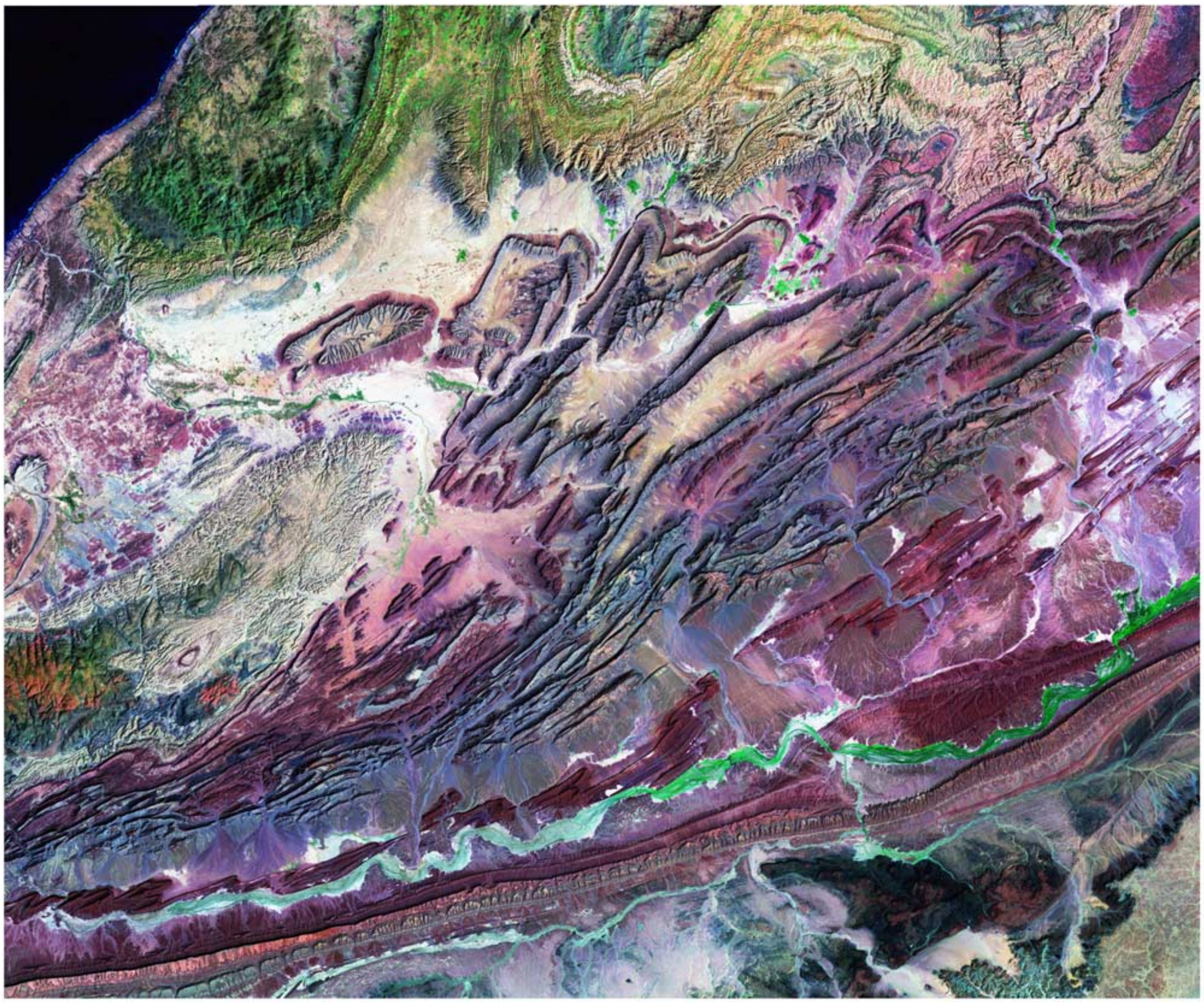

Figure 2. Landsat image of the westernmost Anti-Atlas chain. Compare with Figure 1 for major landmark features such as the basement inliers (dark green) with their ''autochthonous'' Cambrian cover in light green and tan colors. Quartz-rich lithologies of the Jbel Bani, the Jbel Rich, and Jbel Ouarkziz appear in dark purple. Image courtesy of NASA and Earth Satellite Corporation (available at http://zulu.ssc.nasa.gov/mrsid/mrsid.pl).

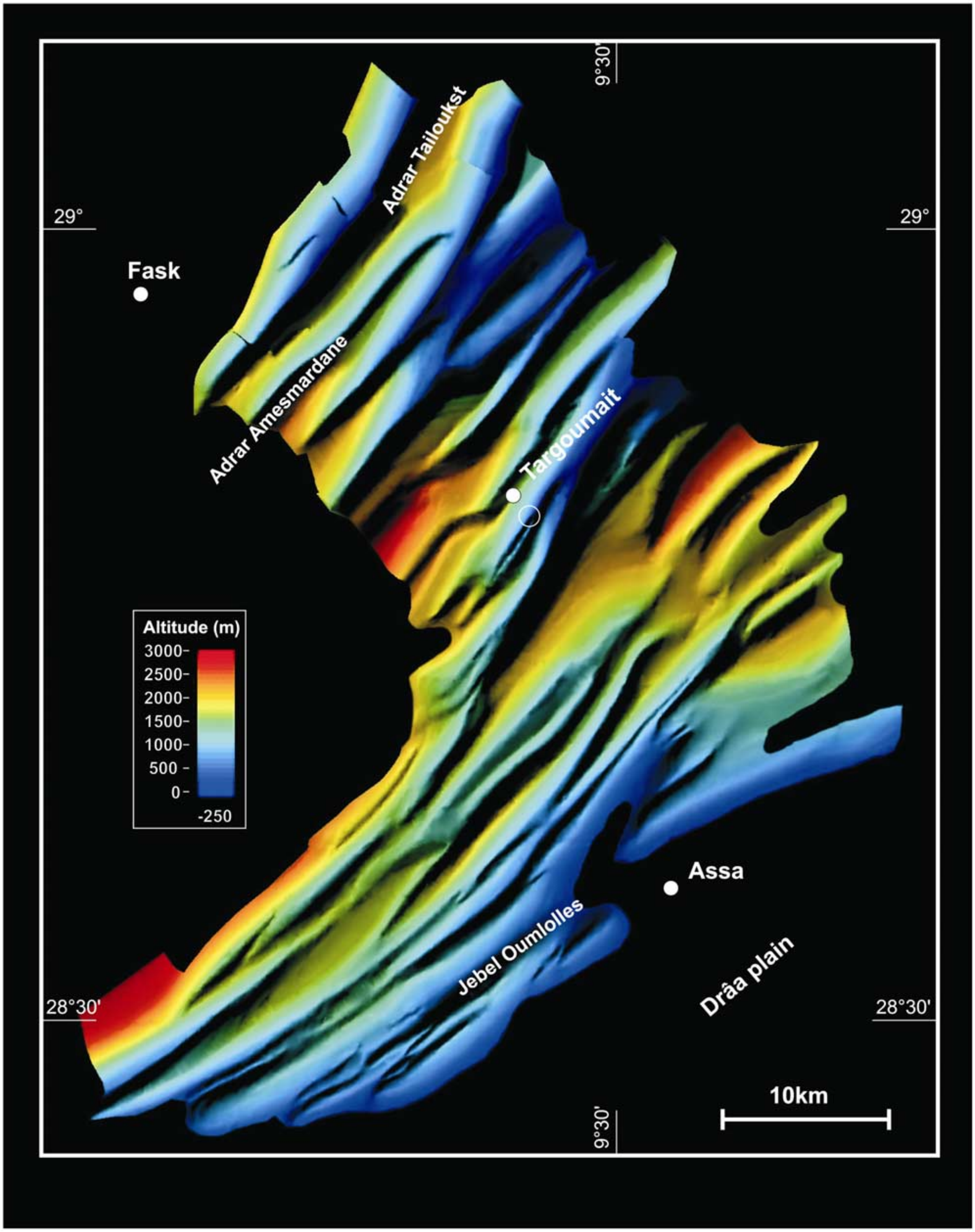

Figure 6. Three-dimensional rendering of a structure contour map constructed for the top of the or6c marker horizon in the western part of the Jbel Bani. The small white circle shows the location of the photograph of Figure 8. Note the decreasing amplitude and wavelength toward the south.