# Maximum extent of ice sheets in Morocco during the Late Ordovician glaciation

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**Abstract**

New field data demonstrate that during the Late Ordovician (Hirnantian) glaciation, an ice sheet expanding northwestwards over the Anti-Atlas range reached into the southern Meseta of northern Morocco. Its growth to a glacial maximum position resulted in extensive subglacial erosion and deformation including the development of soft-sediment striated surfaces and streamlined subglacial bedforms preserved between the High Atlas of Marrakech and Rehamna. These features imply that this ice mass extended >200 km further than previously thought, and increase its size by at least c.190, 000 km$^2$ (comparable in area to the UK). Correlation between a measured section in the High Atlas of Marrakech and that of the southern Meseta identifies sedimentary evolution within an ice-contact system common to both. These findings imply that the West African Craton and northern Morocco were in full glaciological communication during the latest Ordovician. Palaeogeographic reconstruction shows that beyond the ice sheet, south and southeastward palaeoslopes persisted on the shelf. A palaeohigh beyond the main ice sheet was a major source for sand, feeding delta systems that grew along the shelf as far as the shelf break. This palaeohigh probably formed as a result of rift shoulder uplift and supported a satellite ice

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mass. In the eastern Meseta, a thick (350 m) underflow-dominated deep marine fan was fed both from this shelf delta system and from glaciogenic debris derived from the main ice sheet. The occurrence of this unexpected deep marine area in northern Morocco implies that continued northward advance of the ice sheet was hampered by a dramatic break in bathymetry. Two depositional units are recognised across the Meseta, containing four distinct sedimentary cycles, each recognised as a glacioeustatic response to the waxing and waning of ice masses elsewhere in West Gondwana.

**Keywords**: Glaciation, Ordovician, Morocco, ice sheet, Gondwana.

## 1. Introduction

During the latest Ordovician glaciation, an ice sheet grew and decayed over West Gondwana in <1 Myr (Brenchley et al., 1994). During the Hirnantian (Destombes et al., 1985), a stage coeval with the *elongata* and following *oulebsiri* chitinozoan biozones (Bourahrouh et al., 2004), this glacier covered the main part of northern and western Africa (e.g. Beuf et al., 1971; Deynoux, 1985; Ghienne, 2003). The ice was complex and dynamic and characterised by migrating ice fronts during successive phases of ice advance and recession (Ghienne, 2003). The number of glacial cycles preserved in West Gondwana is controversial and varies between 2 and 5; see additional reviews in Rust (1981), Sutcliffe et al. (2000, 2001) and Le Heron et al. (2005).

The record of Hirnantian ice sheet growth is largely preserved in the form of glaciotectonic deformation structures within unconsolidated glacial sediments. They include complex subglacial shear zones, decametre-scale sediment diapirs, load-structures and pluri-kilometric composite thrust and fold systems (Beuf et al., 1971; Biju-Duval, 1974; Ghienne, 2003; Deynoux and Ghienne, 2004; Le Heron et al., 2005). Phases of ice-sheet decay were accompanied by the incision of c. 100-200 m deep, 2-5 km wide anastomosing tunnel valleys from Mauritania to Libya (Ghienne and Deynoux, 1998; Ghienne et al., 2003; Hirst et al., 2002; Le Heron et al., 2004). It is clear that repeated cannibalisation of these areas during successive phases of glaciation has made reconstructing a proximal-distal shelf profile somewhat difficult.

In Morocco (Fig. 1), Destombes (1968) showed that the ice sheet grew over the Tindouf Basin as far as the Anti-Atlas in the south of the country. Extension further northward, over the Southern Atlas Thrust and into the Meseta (Moroccan Hercynides) has never been demonstrated, partly because the tectonic relationship between the Meseta and the West Gondwana cratonic shelf remains highly controversial. It is debated whether these areas were in full palaeogeographic communication (e.g. Piqué and Michard, 1989; Paris et al., 1995; Sutcliffe et al., 2001). Solving these palaeogeographic problems will provide constraints on both the maximum size of the Late Ordovican ice sheet and the reconstruction of the Hercynides.

This paper attempts to define the late Ordovician ice maximum in Morocco, and presents sedimentological evidence that the Moroccan Hercynides (Meseta) were connected to the West Gondwana cratonic shelf during the Lower Palaeozoic. The late Ordovician glacial record in Morocco is unique in that it is one of the only regions in North Africa where the ancient glacial maximum can be determined at outcrop and not from unpublished well data (c.f. Beuf et al., 1971). In addition, in a manner analogous to the analysis of Pleistocene glaciated shelves, it is the first attempt to describe, at a basin scale, the architectural linkage between proglacial shallow shelf through shelf edge to slope systems and the direct response of these systems to this polyphased glaciation.

## 2. Regional geological setting

Throughout the Phanerozoic, central parts of North Africa were characterised by tectonic stability with only relatively minor, intraplate deformation (Selley, 1997). In contrast, the evolution of the West Gondwana margin (e.g. Spain, Turkey, Morocco) was complex and characterised by phases of rifting, basin development, inversion and orogenesis during the Phanerozoic (e.g. Piqué and Michard, 1989; Piqué, 2001).

From north to south, four distinct tectonostratigraphic domains are recognised in Morocco, namely the Rif, Meseta, High Atlas and Anti-Atlas (Piqué, 2001) (Fig. 1). The Meseta comprises a belt of Hercynian deformation that runs between the alpine Rif and High Atlas domains (Piqué and Michard, 1989). In the south of the country, a major structural discontinuity (the Tizi N' Test fault and its associated splay) separate the High Atlas from the Anti-Atlas domain, which comprises a large antiform corresponding to a thick-skinned inversion belt related to the Hercynian uplift of the Precambrian basement of the West African craton (Fabre, 1971; Soulaimani, 1998; Caritg et al., 2004) (Fig. 1).

Some regional palaeogeographic reconstructions have shown northern Morocco as an allochthonous terrane (e.g. Sutcliffe et al., 2001; Monod et al., 2003), despite its Lower Palaeozoic stratigraphic similarity with the Anti-Atlas (Destombes, 1971; Destombes et al., 1985), and the common NE-SW striking Hercynian grain that these regions share (Piqué and Michard, 1989). Intense controversy remains, and compromise models depicting rift margins have also been developed. Extensive Cambrian graben formation in the western part of Meseta (Bernardin et al., 1988; El Attari et al., 1997) and in the western High Atlas (Cornée et al., 1987) was mirrored in the Anti-Atlas (Piqué et al., 1990, Piqué et al., 1995), lending support for failed rifting in NW Africa during the earliest Palaeozoic. For the Late Ordovician, sedimentary models show a large NE-SW oriented trough or basinal depression in the Meseta (Hamoumi, 1988; Khoukhi and Hamoumi, 2001) subsequently amplified during the Devonian (Piqué, 2001). In addition, during the mid-late Ordovician, around 455-445 Ma, a thermal event (Clauer et al.,

1995) is identified and the eruption of Silurian alkali basalts occurred in the coastal (north-western) Meseta (El Kamel et al., 1998). These data lead Stampfli and Borel (2002) to interpret a renewed phase of rifting by the beginning of the Silurian.

A clear evaluation of the tectonic setting of the northern Morocco is vital for the reconstruction of Late Ordovician ice sheets because it could reasonably be expected to have had a controlling influence on the distribution of grounded ice. If northern Morocco was an allochthonous terrane, evidence for late Ordovician grounded ice sheets should not be present. Conversely, evidence for the presence of ice sheets could present a strong case for its palaeogeographic continuity with Gondwana.

## 3. Structural evolution

In northern Morocco, upper Ordovician rocks crop out within large but disconnected Lower Palaeozoic inliers (Fig. 1). During the course of this study, data were collected from the Tazzeka Massif (Middle Atlas, eastern Meseta), the coastal Meseta, from three major Palaeozoic massifs (Massif Central, Rehamna, and Jbilet areas), and from the High Atlas of Marrakech (central High Atlas) (Fig. 1). The inliers were significantly deformed during the Hercynian Orogeny (Devonian to Carboniferous), which is partly concealed by post-orogenic Mesozoic cover. The Meseta was then subject to Alpine tectonism (Piqué and Michard, 1989; Piqué et al., 1993). Therefore, a brief synopsis of its Hercynian tectonic history is necessary to explain the present distribution of Lower Palaeozoic inliers.

In the Meseta, pre-Hercynian deformation is locally intense but probably had little influence on the distribution of upper Ordovician rocks. Caledonian folding and metamorphism occurred at its northern margin in the Sehoul zone (Piqué et al., 1993), and was followed by the emplacement of small granitoids at 430 ± 2 Ma (El Hassani et al., 1991). In contrast, Hercynian deformation, associated with the closure of the Proto-Tethys in the “middle” Palaeozoic and collision of Gondwana with Laurussia (Badalini et al., 2002) was severe. This orogeny was the primary control on the distribution of upper Ordovician strata (Piqué and Michard, 1989). Deformation was polyphased and heterogeneous, propagating westward during the Devonian to the Carboniferous. On the basis of the timing of Hercynian deformation (Piqué and Michard, 1989; Laville et al., 1991), two major tectonic zones can be distinguished within Lower Palaeozoic basement of the High Atlas and Meseta.

Zone 1 includes the eastern Massif Central, the eastern Jbilet, Tazzeka Massif, as well as those outcrops in the eastern Meseta not considered here (Midelt, Debdou, Mekam). Deformation in this zone was characterised by an early phase of westward-directed folding and/or thrusting, during the late Devonian to late Visean (Eovariscan, Breton or Sudete phases: Allary et al., 1976; Hoepffner, 1987; Bouabdelli, 1989). This deformation resulted in significant stratigraphic repetition of the Cambro-Ordovician succession, most clearly demonstrated in the western Tazzeka Massif (Hoepffner, 1987), but also in the eastern Massif Central, eastern Jbilet and Ait Tamlil (Jenny et al., 1983; Bouabdelli, 1989; Bamoumem, 1987; El Houicha, 1994). Zone 1 was then affected by transtension, whereby a basin bounded by NE-SW trending dextral faults, accompanied by andesitic volcanism, formed at the western margin of the Eastern Meseta (Bouabdelli, 1989). Renewed compression formed syn-sedimentary and gravitational nappes in Ordovician to Devonian rocks (Allary et al., 1972; Huvelin, 1977; Hollard et al., 1977; Jenny et al., 1983; Bouabdelli, 1989), thrust west or northwest during the late Visean-Namurian (Bouabdelli, 1989).

Zone 2 lacks Eovariscan structures and deformation was restricted to the late Carboniferous. The western part of this zone, including the extreme west of the Massif Central, western Rehamna, western Jbilet and Isk N’Tazzoult in the High Atlas of Marrakech, was weakly deformed, except at its eastern extremity near the NE-SW oriented Western Meseta Shear Zone (Piqué et al., 1980). Deformation in this zone is characterised by upright, pluri-kilometric open folds associated with nearly vertical wrench faults or reverse faults (El Attari, 2001), variously dipping shear zones and syn- to late kinematic granites (Piqué, 1979; Lagarde and Michard, 1986). In the westernmost Massif Central, the fold structures consist of NE-SW oriented anticlinoria cored by Lower Palaeozoic strata, whereas southward, in the central Rehamna, central Jbilet and High Atlas of Marrakech, NNE-SSW striking thrusts formed during 3-4 phases of deformation (Piqué and Michard, 1989). This was followed by extension and gravity collapse in the Rehamna (Aghzer, 1994; Razin et al., 2001; Baudin et al., 2001), western High Atlas (Cornée, 1989), and probably in the western Massif Central (Baudin et al., 2001). Within this zone, the relative absence of deformation in the Lower Palaeozoic block of the High Atlas of Marrakech compares to the Anti Atlas against which it is juxtaposed (Piqué and Michard, 1989). Structurally, it is thus considered to belong to the Anti Atlas domain (Ouanaimi and Petit, 1992).

Despite the overall intensity of deformation, the consistent orientation of metamorphic isograds, cleavage, and thrust faults between the Coastal Meseta, Massif Central, Rehamna and Jbilet massifs implies that the Lower Palaeozoic inliers are, at a regional scale, mainly autochthonous (Piqué and Michard, 1989). However, it should be emphasised that regional thrusting involving the basement may result in an underestimation of the pre-tectonic distances between the localities described in this paper (particularly in the Jbilet; Huvelin, 1977). These problems must be borne in mind during palaeogeographic reconstruction.

Following the Hercynian orogeny, a period of WSW-ENE oriented crustal extension ensued with the onset of mid-Atlantic rifting in the Triassic (Piqué et al., 2002). Rift basins were infilled from the early Liassic, then inverted by transpressional deformation during the middle Jurassic. This deformation was then overprinted during the Atlas Orogeny in the Cenozoic (Piqué et al., 2002). These post-Hercynian deformation events did not influence the distribution of Lower Palaeozoic inliers (Piqué and Michard, 1989).

**4. Late Ordovician glaciation in northern Morocco**

Hirnantian (syn-glacial) strata are generally devoid of macrofossils in northern Morocco. However, Ashgill fauna are identified in the underlying strata (Razin et al., 2001 and references therein) and lowermost Silurian graptolite-bearing shale is generally found above the syn-glacial strata (Destombes and Willefert, 1988) (Table 1). This shale is thus a useful lithostratigraphic guide, and thus correlation with the syn-glacial Upper Second Bani Formation of the Anti-Atlas is thus readily facilitated (Table 1; Destombes et al., 1985). Formal formation names have thus far only been given to the syn-glacial succession in the Rehamna, Massif Central and Tazzeka sections in northern Morocco (Table 1). In the Massif Central, transitional uppermost Ordovician-lowermost Silurian beds (*persculptus?* and *ascensus-acuminatus* biozones, determination by P. Storch, Prague) have been found by us directly above the glacially-related succession. Similar fauna have been found in the Jbilet (Destombes and Willefert, 1988), suggesting that syn-glacial sedimentation ended before the Silurian in the Meseta.

Below, a detailed stratigraphic description and interpretation of the Rehamna inlier is provided. Because this locality contains the best evidence for the presence of Hirnantian ice sheets, together with a readily interpretable stratigraphy, it is thus used as a benchmark section for the rest of the paper, in which basin-scale stratigraphic architecture is evaluated from outcrops in the High Atlas of Marrakech, Jbilet, Massif Central, Coastal Meseta and Tazzeka inliers (Fig. 1). These data allow palaeogeographic changes over >400 km to be assessed. A correlation scheme is presented, the maximum extent of ice sheets determined, and the temporal relationship between syn-glacial rocks in each inlier is evaluated. All directional data (palaeocurrents, ice flows) has been corrected for tectonic tilt.

### *4.1. Rehamna*

In the Rehamna Massif, upper Ordovician pre-glacial strata form the El Mechach Formation, syn-glacial strata form the Goulibet Formation, and those of the Silurian belong to the El Mesrane Formation (according to the ongoing regional geological mapping programme, P. Razin, pers. comm.). Exposures were studied near the village of Amar Ben Nouti, beneath the Jbel Kharrou, c. 25 km east of the town of Skour des Rehamna. Here, the Goulibet Formation is vertical to overturned and crops out along ridges that mark the crests of stacked, westward-transported and refolded thrust sheets (Hoepffner, 1974). Pre-glacial sediments are characterised by poorly exposed, green-weathering interstratified siltstone and fine-grained sandstone (storm-dominated deposits) of at least 400 m in thickness.

#### 4.1.1. Description

Syn-glacial strata are divisible into two units that are bounded by a zone of soft-sediment deformation. Both units are characterised by 20-25 m thick fining-then-coarsening up motifs (Fig. 2 B).

Unit 1 rests unconformably on the pre-glacial storm-dominated succession. Its lowermost strata comprise channel fills (up to 6 m thick), showing crudely developed cross-stratification, and fining up from coarse to medium-grained sandstone. The latter passes upward into silty shales, containing one isolated bioclastic (*Lingula*-bearing) bed. These are overlain by a coarsening-upward motif, characterised by the progressive appearance in laminated mudrock of tabular, 5-15 cm thick, well bedded, current-rippled sandstone beds with occasional load casts. A sharp-based, coarse-grained sandstone bed (20-40 cm thick), and overlying thick-bedded (20-40 cm), fine-to-medium grained, climbing-rippled sandstone with southward directed palaeocurrents forms the upper part of unit 1 in the study section. The latter sediments are involved in the zone of soft-sediment deformation. However, laterally, a more complete record has been preserved (Fig. 3). Above the rippled sandstones, a several metre-thick mudstone succession is cut by thick channelised sandstone bodies with pervasive parallel lamination or trough cross-strata. They are in turn capped/ truncated by a distinctive sandy diamictite corresponding to the base of unit 2 (Figs. 3, 4 A-C).

The clast-poor, red-brown, sandy diamictite is widespread and traceable over several kilometres. It is intimately associated with the zone of soft-sediment deformation affecting up to 15 m of sediment. Detailed sections (Fig. 3) reveal that this deformation zone contains from base to top: 1) chaotic and north to northwestward-verging, decametre-scale folds affecting the underlying sandstones, 2) soft-sediment striae preserved along intrastratal detachments within the diamictite (Fig. 4 C), and 3) downward-tapering sedimentary dykes, penetrating the diamictite and sourced from an overlying coarse-grained rippled sandstone (Fig. 4 B). This sandstone, of several decimetres thickness, contains a pervasive shear fabric (Fig. 4 D) and locally rests angularly on the diamictite, which is undeformed.

Above the deformation zone and associated sediments, the lower half of unit 2 differs from that of unit 1 in the importance of poorly sorted sediments (clast-poor, muddy to sandy diamictites), punctuated by minor sand-filled channels. The middle to upper coarsening-upward part comprises from base to top 1) laminated very fine-grained sandstones, 2) thin and tabular-bedded sandstones with occasional ball-and-pillow structures, load casts and flute casts in the lower beds, hummocky cross-stratification in the upper beds, 3) a sharp-based, crudely cross-stratified, coarse- to very coarse-grained sandstone bed, 4) fine-grained pure-quartz sandstone beds, bearing south-eastward dipping sigmoidal cross-beds with cyclic thickness variations (Fig.

4 E). The latter are capped by a final bioturbated horizon. The contact with the overlying Silurian shale is tectonic.

4.1.2. Interpretation

The key point in the Rehamna section is the preservation of a restricted zone of soft-sediment deformation. In a manner consistent with similar deformation structures in the late Ordovician record of Libya (Le Heron et al., 2005), the folded zone is interpreted as a suite of gravitational deformation structures induced by the weight and/or shear of an overriding ice sheet. The truncation of these folds by the distinctive red-brown diamictite and the occurrence of a soft-sediment striated surface (Deynoux and Ghienne, 2004) add further credence to this interpretation. The structures thus formed within a subglacial shear zone (Hart and Boulton, 1991; Van der Wateren et al. 2000). Downward-injection of sedimentary dykes is not unusual in subglacial settings (e.g. Hyam et al., 1997). In the Rehamna, a mechanism involving ice sheet-induced intrusion of the dykes is favoured by a) their downward tapering nature and b) the position of the source bed above them. The pervasive shear fabric in the coarse-grained topmost bed suggests a glacial outwash deposit directly overridden by flowing ice. Therefore, its upper surface probably formed at the ice-sediment interface.

The presence of a subglacial deformation zone in the Rehamna implies that sediment supply and depositional facies are at least partly related to the waxing and waning of this ice mass in the Meseta. These structures reflect a glacial erosion surface (GES). This provides guidelines for the interpretation of the sedimentary record in the Rehamna.

The sharp-based and coarse-grained lithofacies at the base of unit 1 are interpreted as an ice-proximal outwash deposit, in the form of glaciofluvial or glaciomarine bars on the formerly storm-dominated shelf. They imply a first significant ice advance, with well-developed fining-up of overlying sediments indicating rapid retreat. The concentration of *Lingula* at a discrete interval above these sediments could correspond to opportunist colonization within the marine setting of an interglacial. Above, the succession of rippled sandstone beds with finer-grained interbeds, sharp-based coarse-grained sandstone beds and overlying climbing rippled lithofacies corresponds to regressive-transgressive fluctuation reflected successively by shoreface, fluvial and tidal flat deposits. A glacio-eustatic cycle is inferred for these deposits since they bear no evidence of glacially fed depositional environments. Therefore, mudstones that cap these sediments were probably deposited during renewed interglacial conditions. Overlying rocks of the uppermost part of unit 1 were thus deformed during a second, significant glacial advance. The lower part of unit 2 is considered to reflect 1) erosion and deposition of high-energy sandstones reflecting ice-proximal outwash deposits, 2) diamictite deposition (either glaciomarine or subglacial?), 3) subglacial deformation and glacio-fluvial deposition. The overlying clast-poor, muddy to sandy diamictites and sand-filled channels reflects glaciomarine sedimentation in an underflow-fed submarine outwash fan system during ice sheet recession.

Overlying shoreface sediments, punctuated first by turbiditic (load and flute casts) and then storm (Hummocky cross-stratification) events, may have been deposited in a prodelta setting. As in the upper part of unit 1, the succession of shoreface deposits, a thin fluvial horizon and overlying well developed, tidally bundled, migrating bedforms (Nio and Yang, 1991) reflects an additional regressive-transgressive fluctuation. The tidal sediments may represent transgressive facies below the overlying Silurian shales. However, comparison between the upper part of unit 2 with that of unit 1 identifies a further glacio-eustatic cycle operational prior to post-glacial flooding in the earliest Silurian.

4.1.3. Regional implications

Vergence of fold structures in the subglacial deformation zone indicate a top to the north sense of shear, indicating the growth of ice sheets from ice centres in the Anti-Atlas and further southward (Algeria, Mauritania). The presence of this subglacial deformation zone in the Rehamna implies that the Hirnantian glaciers extended into the Meseta at least 200 km over the Southern Atlas Thrust.

Depositional units correspond to two first-order glacial cycles (e.g. Sutcliffe et al., 2000), each of them subdivided into two second-order cycles. Thus, event stratigraphy identifies four sedimentary cycles in total, each potentially controlled by regional advance and retreat cycles of the Hirnantian ice sheet (Ghienne, 2003). In each unit, the first (lower) second-order cycle records ice-proximal to subglacial deposition, and subsequent retreat following a major ice sheet advance into the basin. In contrast, the second (upper) second-order cycle only records a glacio-eustatic fluctuation, lacking any evidence for even local glaciers in the basin. An alternative is that the uppermost cycle in each unit was driven by glacioisostacy (isostatic rebound) following glacial retreat as suggested by Sutcliffe et al. (2000) elsewhere in North Africa. However, this explanation is not considered appropriate because no evidence of uplift, hardground development or erosion is preserved at the base of these cycles in Rehamna. These conclusions are fully supported by the record of the Hirnantian glaciation in the other Palaeozoic inliers of Northern Morocco.

### *4.2. High Atlas of Marrakech*

Exposures at Isk N' Tazzoult, 2.5 km west of the Tizi N' Tichka pass (Fig. 1), occur in a Lower Palaeozoic inlier that preserves a thick succession of Cambrian to Silurian strata (Ouanaimi, 1989, 1998). This locality belongs to the Anti-Atlas tectonostratigraphic domain rather than to the Meseta (Allary et al., 1972; Ouanaimi and Petit, 1992). Therefore, comparison between upper

Ordovician deposits in this locality with those in the Meseta will yield affirmative data on the palaeo-tectonic relationship of these two regions.

4.2.1. Sedimentary succession

The syn-glacial strata rest with sharp disconformity upon pre-glacial, bioturbated, storm-dominated fine-grained deposits. Large (100 m wide) crosscutting channels (Fig. 5), filled with sand-grade sediment are interstratified with heterolithic lithofacies including well-stratified shale, siltstone, and fine sandstone or massive sandy clast-poor diamictite. Although complex lateral and vertical substitutions occur between these lithofacies, the basal part of the succession comprises coarse-grained sand-filled channels (Fig. 2A, 17-25 m) overlain sharply by sandy diamictite that bears large (1-5 m) ball and pillow structures (Fig. 2A, 25-37 m). The diamictite is capped by a well bedded, coarsening and thickening upward rippled siltstone and sandstone succession, with bioturbation (Fig. 2A, 37-43 m), passing upward into shale (Fig. 2A, 45 m). Coarse-grained sand-filled channels with scattered boulder-sized clasts (Fig. 6 A) (Fig. 2A, 45-65 m) cut down into these sediments. These sediments, which show flute casts, tool marks and conglomerate lenses at their base, are reworked up section into a soft-sediment deformation zone comprising large, polished, streamlined bedforms (Fig. 6 B) with superimposed hairpin structures (Fig. 6C) (Fig. 2A, 65 m). This zone, which is sealed by a thin and discontinuous diamictite horizon, is in turn overlain by a thin coarsening-then-fining upward succession with lenticular coarse-grained sandstone beds in the middle part (Fig. 2A, 66-70 m). The topmost part of the succession comprises finely laminated siltstone and black shale (the so-called Silurian shale), which attains at least 30 m thick. Rare lonestones occur within the lowermost part of this shale.

The large channels were cut and filled in a high energy setting, most probably in a sub- or proglacial environment during the rapid release of glacial meltwater (e.g. by "jökulhlaup" outburst floods). The heterolithic lithofacies variably indicate suspension settling (shale), traction (well bedded siltstone and fine sandstone) and slurry/mass-flow processes (diamictites). Bioturbation implies colonisation of the substrate by opportunistic fauna in environments far from the glacial sediment input points. Polished streamlined bedforms result from glacial fluting, and hairpin structures resemble spindle flutes found on Pleistocene deglaciated surfaces (e.g. Kor et al., 1991). These structures, interpreted as subglacial bedforms and attributed to a grounded ice sheet, indicate a subglacial deformation zone in the upper part of the succession in the High Atlas of Marrakech.

4.2.2. Event stratigraphy and correlation

The following sequence of events is interpreted from the syn-glacial succession: 1) incision of channels into pre-glacial substrate and deposition of ice proximal lithofacies, 2) glacial retreat (diamictite fining up to non-glacial mudrock), 3) a distinct regressive-transgressive cycle remote from an ice front, 4) following a second phase of channel incision, truncation of ice proximal lithofacies by a GES, 5) glacial retreat (thin diamictite drape), 6) a regressive-transgressive cycle deposited under a glaciomarine influence (channels with coarse-grained sand fill, occasional dropstones). Comparison to the Rehamna section suggests that most of the succession belongs to unit 1 in the High Atlas (below the major subglacial surface); topmost sediments, bearing the uppermost glacially related facies, therefore correspond to unit 2. As in other parts of northwest Africa such as the Taoudeni Basin (Underwood et al., 1998; Paris et al., 1998), these sediments are transitional into postglacial (probably late Hirnantian and earliest Silurian) shale with no hiatus inferred.

*4.3. Jbilet*

Lying between the High Atlas and the Rehamna (Fig. 1), the Jbilet inlier probably occupied an intermediate palaeogeographic position. Like the Rehamna, it consists of westward-directed imbricate thrust slices (Huvelin, 1977). Recognition of syn-glacial deposits includes the disappearance of the ubiquitous bioturbation present in pre-glacial mudrocks, or the sudden appearance of diamictites. Uppermost sandstones of the succession have yielded a terminal Ordovician graptolite fauna (*persculptus* biozone, Destombes and Willefert, 1988).

Syn-glacial deposits attain >160 m thickness and comprise structureless, homogeneous, unstratified clast-poor sandy diamictites. Two poorly defined coarsening-up cycles are recognised, the lower of which is punctuated by >10 m wide, sandstone-filled channels (Fig. 7), whilst the upper one is capped by current rippled and cross-bedded (northeastward-directed palaeocurrents) siltstones and fine-grained sandstones.

The homogeneity and thickness of the sandy diamictites implies that the Jbilet lay close to an ice front during deposition. Their unstratified nature, coarsening-up profile and interruption by channelised sandstone implies the rapid progradation of a thick, subaqueous diamicton fan whose poorly expressed foresets were dominated by glaciogenic debris flows and topsets by tractive processes, rather than an ice-rafting mechanism.

### *4.4. Massif Central*

Detailed study was undertaken in the western part of the Massif Central at Mchmech El Diab, 10 km SE of the village of Ezziligha (Fig. 2 C) and additional outcrops were examined in the central, northern and eastern Massif Central. A synthetic overview is given below. Syn-glacial strata of the mud-dominated Ould Akra (early-late Ashgill) and overlying sand-dominated Ezzhiliga formations (late Ashgill), are overlain by shales of the lower Silurian Sidi M'Bellej Formation (Baudin et al., 2001; Chèvremont et al., 2001; Razin et al., 2001).

4.4.1. Sedimentary succession

The lower contact of the syn-glacial strata (upper Ould Akra Formation) is placed at the first appearance of clast-poor muddy diamictites above bioturbated siltstones and fine-grained, current rippled to hummocky cross-stratified sandstones. Diamictites, either massive or crudely stratified, make up the lower part of unit 1 (Fig. 8 A). These become thicker and coarser-grained up section at Mchmech El Diab (Fig. 2 C, 3-40 m), but in the eastern Massif Central (Khenifra area) fine upwards. Lonestones are pebble to cobble-sized, and comprise quartzite or sandstone (Fig. 8 B). Discontinuous, cross-stratified sandstone horizons are locally interstratified within the diamictites. Above, sand-dominated lithofacies (Ezzhiliga Formation) are widespread and traceable over the whole Massif Central (Figs. 8 C-E, Fig. 2 C, 40-85 m). Locally, they rest on the underlying diamictites with a sharp and loaded contact. This sandstone comprises two conspicuous coarsening and thickening upward motifs, comprising moderately to well sorted, fine to coarse-grained sandstone (Fig. 8 E). Wave ripples and rare amalgamated hummocky cross-stratification (HCS) occurs in the lowermost sandstone beds of the lower thickening upward motifs with metre-scale megaripples becoming more common upwards. The sharp surface bounding the two thickening upward motifs is regionally developed dividing unit 1 from unit 2 (Figs. 2 C, 8 E). Poorly exposed sediments at the base of unit 2 comprise interstratified siltstones and fine-grained sandstones. Its upper part compares closely to that of unit 1, but characterised by more abundant climbing ripple cross-stratification (Fig. 2 C). At Mchmech El Diab, a green siltstone unit, several metres thick, caps the sandstone (Ezzhiliga Formation) and fines up into graptolite-bearing shale with a thin, ferruginous horizon. In the eastern Massif Central, graptolite-bearing shale directly overlies the sandstone.

Fluid escape structures are locally abundant in the sandstone lithofacies (Razin et al., 2001). However, no deformation structures related to soft-sediment subglacial deformation were found. Regionally, the sand-dominated Ezzhiliga Formation tends to be finer grained and thinner southward and eastward, an observation consistent with south-eastward directed palaeocurrents in rippled or cross-stratified facies.

HCS-bearing and current rippled sandstones in the latest pre-glacial deposits imply storm-deposition in a dominantly lower shoreface setting. The diamictites are interpreted as glaciomarine facies, deposited from glaciogenic debris flows with gravity re-sedimentation. However, their variable motif (variably coarsening and fining up depending on the locality), coarse granulometry and sharp upper contact with the overlying sandstones suggest that they do not belong to the lower part of a coarsening-upward succession to which the Ezzhiliga Formation forms the upper part. The two regionally extensive overlying coarsening and thickening up motifs comprising relatively well-sorted sandstones formed through efficient mud winnowing on delta mouth bars. The predominance of the sharp to irregular-based sandstone beds implies low accommodation and extensive, high-energy re-distribution. Storms contributed to mouth bar reworking, as evidenced by HCS, but the predominance of cross strata highlights the importance of traction currents upsection. The sharp surface separating the two motifs is interpreted as a major abandonment surface. A second phase of delta mouth bar build-out is interpreted on the basis of a renewed coarsening-up profile following abandonment of the former delta and the occurrence of similar lithofacies to those in the lower part of the Ezzhiliga Formation. The overlying green siltstones most probably represent prodelta deposits, grading up into shale reflecting offshore conditions. As siltstones and then shale overly abruptly delta sandstones, a rapid backstepping of the delta system is envisaged.

#### 4.4.2. Event stratigraphy and correlation

Compared with the High Atlas and Rehamna domains, the sedimentary record in the Massif Central is correlated in the following way: diamictites (uppermost Ould Akra Formation) are correlative to the first ice-proximal deposits (Fig. 2). Coarsening-upwards sand-dominated motifs (delta mouth bars) most likely represent proximal counterparts of the two regressive-transgressive, glacio-eustatically driven second order cycles (Fig. 2). Under this model, the subglacial erosion surface (GES) identified in the High Atlas and Rehamna domains, probably synchronous with a major sea-level fall, corresponds to the abandonment surface between the two delta systems. The ferruginous horizon in graptolite-bearing shale above the delta should thus be correlated with the postglacial maximum flooding surface within Silurian hot shales elsewhere on the Gondwana platform (Lüning et al, 2000).

*4.5. Coastal Meseta*

Syn-glacial strata of Ben Slimane (Oued Nfifkh), which await formal stratigraphic recognition, are affected by 3-4 stages of Hercynian deformation, and large-scale conjugate faulting (El Attari, 2001). They comprise two well-defined coarsening and thickening upward motifs, overlain by a diamictite succession (Fig. 2 D). The lower motif (unit 1) comprises mudstones with thin current-rippled sandstones (>30 m) overlain by coarsening-up, well-sorted, siltstone to medium-grained sandstone that bears eastward-directed, decametre-scale cross-beds and linguoid ripples. A decimetre-thick, erosionally based pebbly conglomerate caps this succession (Fig. 2 D). The upper coarsening upward motif (unit 2) sharply overlies the lower one and comprises an identical, coarsening-up lithofacies evolution capped by a second conglomerate. In contrast with the Massif Central, topmost deposits are abruptly overlain by 10 m of clast-poor sandy diamictites, passing transitionally into graptolite-bearing shale (Fig. 2 D). No striated surfaces or other glaciotectonic structures are preserved in this area.

The coarsening-up motifs and progressive upward appearance of large bedforms implies two phases of deltaic progradation. The sharp surface bounding the two motifs is interpreted as an abandonment surface prior to a renewed phase of build-out. The thin, sharp-based pebble beds capping each unit suggests scour on the upper surface of the mouth bar, potentially by fluvial processes. These data would imply that the delta topsets were extremely thin and characterised by bypass rather than deposition. However, a 20 m thick succession of pebbly sandstone with large-scale cross-bedding occurs northward (40 km west of Settat, Sutcliffe et al., 2001) and may suggest a thickening of the pebbly facies, either locally (channel, incised valley ?) or regionally (in a more proximal setting). Textural characteristics of the overlying diamictite suggest direct glacial derivation. The two successive coarsening-up motifs in both the Coastal Meseta and Massif Central are clearly correlative. Reasons for the occurrence of a diamictite in the upper synglacial succession are discussed below.

### *4.6. Eastern Meseta*

In the Tazzeka Massif, pre-glacial strata (Tehar El Brehl Formation), syn-glacial strata (Tifarouine Formation) and the overlying Silurian shales crop out immediately south of the Rif in a 15 km radius of Tahala, with the best exposure at Zerarda (Fig. 9) (Destombes et al., 1985, Khoukhi, 1993; Khoukhi and Hamoumi, 2001). These exposures are presently located c.200 km north-east of those in the Massif Central (Fig. 1) but prior to Hercynian deformation may have been more distant during the Ordovician.

#### 4.6.1. Sedimentary succession

Pre-glacial rocks mainly comprise rhythmically bedded, bioturbated, silty shale and siltstone. In the upper levels of pre-glacial strata, within which Ashgill palynomorphs have been identified (determination F. Paris, Rennes), fine-grained lithofacies interbedded with current-rippled sandstone beds prevail. Bioturbation becomes more subtle upward. The first stratigraphic occurrence of syn-glacial strata is considered to be defined by a sharp-based sandstone bed with flute casts (Fig. 9; 25 m).

Syn-glacial strata are organised into two well-defined stratigraphic units (Figs. 9, 10), traceable over 20 km. Subordinate smaller-scale coarsening-up cycles on two smaller orders of magnitude are also expressed (Fig. 11 A). Syn-glacial strata are made up of extremely well-differentiated shale, siltstone, sandstone and rare clast-poor, sandy diamictite. Each coarsening-up cycle attains 100-150 m thick, with an unexpected total section thickness (approaching 350 m; Figs. 9, 10). The diamictite facies are preserved intermittently in unit 1 (attaining several metres thick at c. 70m, Fig. 9) but rare in unit 2.

Two types of fine- to medium-grained sandstone lithofacies co-occur with mudstone, namely 1) thin, isolated, SE-directed current-rippled horizons that predominate at the base of each unit and 2) well-defined, structureless, sometimes graded, crudely horizontally laminated or rippled tabular beds prevailing mid-way up each unit (Fig. 11 A). These latter sandstone beds occur toward the top of metre-scale coarsening-up motifs (Fig. 11 A). They are sharp-based, irregular, or channelised with S to SE-directed (and rarely NW-directed) flute casts or small channels occasionally filled by coarse-grained sandstone. Aside from centimetre-scale load casts at the base of these beds, no soft-sediment deformation occurs in the Tazzeka Massif.

The upper part of both units comprises stacked, erosionally based, horizontally laminated sandstone beds, forming amalgamated packages up to 10 m in thickness (Figs. 9, 10). They rest directly on clast-poor muddy diamictite. At the top of unit 1 (135-170 m, Fig. 9), they are coarse-grained and contain abundant sheet dewatering structures. Metre-scale flute-like structures of consistent long-axis orientation toward the SE (Figs. 11 B-D) containing gravel lags are overlain by a single fining upward sandstone bed, 40-80 cm thick, with a wave-rippled top. A sharp contact with overlying shale and sandstone occurs.

Sandstones capping unit 2 are fine-to medium (and rarely coarse) grained (320-375 m, Fig. 9). Diffuse horizontal lamination and small flute casts occur. The transition into Silurian shale and well bedded phtanite is sharp and tectonic. Scattered quartz grains (< 1 mm) are preserved in the lowermost exposures of shale below phtanitic beds.

4.6.2. Interpretation

Together, the lithofacies described above characterise the cyclic progradation of a turbiditic deep marine fan (e.g. Stow and Mayall, 2000). Significant water depths during deposition are indicated by the great (and presently compacted) thickness of these sediments, by the absence of angular unconformities supporting continuous sedimentation and by the almost total absence of wave- or storm-induced sedimentary facies.

The three main types of sandstone lithofacies i.e. isolated and current rippled, the tabular beds within mudrock, and the amalgamated, stacked examples, are interpreted to belong to distal, medial and proximal portions of the fan system respectively. The first lithofacies were probably deposited by distal, low-density turbidity currents; the SE-oriented palaeocurrents obtained from these rocks thus provide a palaeoslope indicator. The second and third sandstone lithofacies are interpreted as sandy gravity flows, with the diffuse, horizontal lamination within the amalgamated sandstone beds testifying to non-Newtonian rheology, internal shearing, and to the emplacement of stacked debris flows (Shanmugam, 2000). The emplacement of these last flows as discrete “event beds” is supported by the frequent occurrence of fluted and gravely surfaces within these strata. Their lateral extent over at least 20 km, together with their position at the top of the first-order coarsening up cycles, indicates that they occupied an extensive upper part of this fan system. The surface with large-scale flutes is thought to reflect high-energy underflows, possibly linked with abrupt meltwater discharge during jökulhlaup glacial outburst events (e.g. Carrivick et al., 2004). A rapid end to sand sedimentation above this surface implies avulsion of the depositional lobe.

4.6.3. Event stratigraphy and correlation

Syn-glacial rocks in the Tazzeka Massif are notable for their significantly greater thickness (350 m) than anywhere else in the Meseta and, in common with the Massif Central and Coastal Meseta areas, the absence of glacially-related soft-sediment deformation structures. These observations suggest high sediment accommodation within a basin untouched by an ice sheet. Four cycles corresponding to coarsening and thickening upward succession bounded by diamictite horizons are recognized (Fig. 9). The proposed correlation, developed further below, suggests that these four cycles reflect the four glacial cycles identified in the Rehamna and High Atlas domains, the major glacial advance being reflected by the coarse-grained lithofacies of the upper third of unit 1 and its conspicuous fluted erosion surface.

## 5. Hirnantian sedimentary systems in northern Morocco

Our data demonstrate the presence of the late Ordovician ice sheet in the Meseta that reached as far north as the Rehamna Massif. Unequivocal evidence for subglacial deformation is restricted to the surface separating depositional units 1 and 2 (Figs. 2, 12). This surface formed during the third glacially-related cycle and is considered to reflect the maximum extent reached by the Gondwanan ice sheet during glaciation of Morocco (i.e. the glacial maximum). However, the coarse-grained and erosionally-based nature of sandstones at the base of unit 1 in both the Rehamna and High Atlas probably imply deposition during an earlier glacial advance prior to advance of the ice sheet toward its ice maximum position. No direct evidence for subglacial erosion (e.g. soft sediment striation) was found in or above these channel deposits, but in the High Atlas, the abundance of sandy diamictites above them strongly suggests the presence of a nearby ice front. Therefore, these data imply two advances of the Gondwanan ice sheet into northern Morocco.

This section investigates why the ice did not extend further north into the Massif Central or Coastal Meseta. It unravels complex stratal geometries (Fig. 12), proposing a palaeogeographic reconstruction (Fig. 13 A) and depositional model (Fig. 13 B).

### *5.1. Regional correlation*

A correlation is proposed between the glacial record of four of the Palaeozoic inliers (High Atlas of Marrakech, Rehamna, Massif Central and Coastal Meseta), which are approximately aligned in a palaeogeographic profile (Figs. 2, 12 and 13 A). Four chronostratigraphic boundaries are identified; the first and third corresponding to glacial erosion surfaces (GES).

Limited but consistent palaeocurrent data, palaeo-ice flow indicators and regional trends identify three palaeogeographic domains. The “ice-contact system” was characterised by a south-north proximal-distal trend in glacially-related sedimentary systems identified in the Rehamna and High Atlas (Fig. 12, blue colour). The second, a “shelf-delta system”, shows that a north-south (to NE-SW) proximal-distal trend persisted in the delta systems of the Massif Central and Coastal Meseta (Fig. 12, orange colour). Therefore, ice contact and shelf delta sediments were shed from different sources. The linkage and relationship between these systems is further investigated below.

The Jbilet and Tazzeka inliers are more difficult to place on the High Atlas- Coastal Meseta profile because the former, comprising only glacially-derived material and lacking delta deposits, is regarded as a distal glaciomarine equivalent of the High Atlas section (Fig. 13). The Tazzeka succession represents a third palaeogeographic domain ("deep-marine system") and the buildout of a submarine fan in the eastern, deeper portion of the platform connected to and fed by both the shelf-delta and ice-contact systems (Fig. 13).

Below, the linkage between ice-contact, shelf-delta and deep-marine systems is investigated, with emphasis on palaeogeographic setting at the glacial maximum (Fig. 13 A).

### *5.2. Ice-Contact system*

At the glacial maximum, the upward transition from outwash facies (diamictites and large sand-filled channels), into a conspicuous subglacial deformation zone, and finally distal glaciomarine facies (diamicton deltas?) is characteristic. These changes imply proximal proglacial deposition during ice-sheet advance, full glacial conditions, and rapid retreat of the grounded ice sheet. Lateral and vertical lithofacies substitution is common in the ice-contact system. However, since ice-contact systems are recognised in both the High Atlas and Rehamna sections, palaeogeographic linkage between these areas at the glacial maximum was likely. Sediment was probably transported onto the proglacial shelf by a combination of subglacial and proglacial channels (Fig. 13). The transition toward shelf delta systems is less certain because of the lack of continuous exposure between the Rehamna and Massif Central, but it is suggested that these systems interdigitated at their depositional termini (Fig. 12).

### *5.3. Shelf-delta system*

The lack of subglacial soft-sediment deformation structures north of the Rehamna inlier suggests that syn-glacial sediments north of it were deposited in a paraglacial setting untouched by an ice sheet (Fig. 13 A). The growth of deltas in this area is consistent with earlier interpretations (e.g.

Khoukhi and Hamoumi, 2001; Baudin et al., 2001; Razin et al., 2001). The similarity in the vertical lithofacies successions in both the Massif Central and Coastal Meseta, and lateral extent of mouth bar sands over at least 100 km between them, implies delta formation over a vast, topographically subdued shelf area. Growth of these deltas during a significant phase of sea-level fall is supported by the sand-prone nature of mouth bar facies and evidence for their extensive reworking. Syn-glacial sediments of the Massif Central and Coastal Meseta compare to shelf margin deltas perched at or near a break in slope on the continental shelf (Porębski and Steel, 2003). During glacio-eustatic sea level fall, bypassing of sediment onto the shelf edge is promoted, and sandy mouth-barred deltas offlap inner shelf sand and mud (Porębski and Steel, 2003). Such a palaeogeographic position is supported by the lateral extent and tabular geometry of their mouth bar sands. Thin fluvial sediments (pebbly horizons) were deposited either as delta topsets or during late lowstand conditions. As tidal facies interpreted as transgressive deposits occur directly above the mouth bars, the second hypothesis is preferred.

Shelf margin deltas probably merged landward (northward) into a network of fluvial channels (Fig. 13 B). The local occurrence of clast poor, sandy diamictites in the Coastal Meseta could suggest that at least some sediment was ice-derived from a local ice cap (see below). Basinward (southward), deltas merged into inner shelf environments (High Atlas, Rehamna) (Fig. 12), dominated by storm deposition, with subordinate turbiditic deposition.

At the glacial maximum, accommodation space was minimal. The disconformity separating the two coarsening-upward motifs in the Massif Central and Coastal Meseta is considered to be time-equivalent to subglacial erosion southward, forming as the ice advanced to its glacial maximum position (Fig. 12). It resulted in erosion of the mouth bar as a result of sea-level fall during ice sheet expansion. At that time, the fluvial system would have connected directly with an underflow-dominated deep-marine system (Tazzeka domain). Whilst no canyon systems were observed by us, probably as a result of their very limited lateral extent, they are likely to exist and probably formed an integral sediment pathway to the deep marine system (Fig. 13 B).

### *5.4. Deep-marine system*

During the late Quaternary glacial maximum, large trough-mouth fan-systems formed beyond the shelf break in the Barents Sea (e.g. Taylor et al., 2002). However, these deposits are a poor analogue to those of the Tazzeka Massif because they are built largely of unevolved glaciogenic debris flows. In contrast, Wilkes Land turbidite systems fed by palaeo-ice streams in Antarctica are a better analogue (e.g. Escutia et al., 2000). Whereas the main part of the Tazzeka succession represent deposition on distal to medial part of a deep marine fan, the stacked, amalgamated packages of sandstone towards the top of unit 1, interpreted as stacked debris flows, are best developed in the proximal part of the turbidite fan and imply heightened rates of sediment supply, sediment remobilisation, and fan-lobe growth in the deep basin. These conditions prevailed at the glacial maximum.

Knowledge of the causes of sediment remobilisation is essential to understand the linkage between ice-contact, shallow and deep-marine settings. In modern settings, mobilisation of shelf-margin sediments to the deep marine system occurs via mass-movement/slope instability (Stow and Johansson, 2000), “seafloor polishing” from wave and tide agitation (Armishaw et al., 2000; Stow and Mayall, 2000), or direct throughput by fluvial floods (Mutti et al., 1996, 1999). These processes generate gravity flows (Shanmugam, 2000).

In the Tazzeka Massif, turbidites evolved downdip from mouth bar sands at the foot of a S to SE-ward slope from the Massif Central and Coastal Meseta. These conditions are thought to have persisted during the first, second and fourth glacially related cycles. It is thus suggested that bypass wedges resulted from periodic instability of the shelf-margin delta mouth bars, in combination with direct throughput by fluvial floods. Distal glaciomarine facies were deposited during ice retreat episodes and correlative sea-level rise.

This deep-water area lay beyond both the shelf-delta and ice-contact systems (Fig. 13 A, B). It was the final repository for glacially derived sediments, sourced from the NNW advancing ice sheet that covered the Tindouf Basin (Destombes et al., 1985). The latter interpretation is supported by earlier work (Sutcliffe et al., 2001), raising the possibility that sediments in the deeper parts of the basin were derived from two sediment sources, one from the ice sheet, and the other from an upland area/ palaeohigh beyond the ice front (Fig. 12). Such a mixed sediment source is envisaged for the amalgamated sandstones at the top of unit 1 (deposited during the glacial maximum), both by their coarse-grained nature and their depositional processes (stacked debris flows, high-energy underflows). The dramatic increase in sediment thickness between the Massif Central/ Coastal Meseta and the Tazzeka Massif (Fig. 2, inset) may thus indicate the first description of a late Ordovician trough-mouth fan in North Africa, though further work is required to confirm this interpretation.

## 6. Palaeogeographic Implications

### *6.1. Gondwana ice-sheet extent*

The lateral extent of subglacial bedforms is a powerful means of constraining the dimensions of former ice sheets (e.g. Jansson and Glasser, 2005). The occurrence of subglacial deformation structures in the High Atlas of Marrakech and the Rehamna inliers extends the ice sheet at least 200 km beyond the northernmost occurrences of striated surfaces reported by previous authors (Central Anti Atlas, Destombes, 1969; Hamoumi, 1988), or 150 km north of the most distal glaciotectonic folds and thrusts (Alnif, Sutcliffe et al., 2001) or glacial valleys (Destombes et al., 1985; Alvaro et al., 2004). Taking into account directional data in the Anti-Atlas, and assuming a lobate ice front geometry, similar to that of the late Pleistocene in Fennoscandia (e.g. Punkari, 1997), then the ice covered a semi-elliptical area of at least 200 km X 400 km. The area of this hitherto unknown ice cover is approximately 0.2 million km$^2$, of comparable size to the UK (Fig. 13 A).

The new data presented in this article give a lucid illustration of the palaeogeographic changes that occur at and beyond the terminus of an ancient ice sheet at its glacial maximum. Both the geometry of this ice front and the character of sediments debouched from it were probably closely dependant on the character of ice sheet flow and processes of erosion further up-glacier, south into the Anti-Atlas of Morocco and the Algerian Sahara. Recent work on the Pleistocene glacial record has focussed on the location of palaeo-ice streams, fast-flowing corridors of ice up to hundreds of kilometres long and tens of kilometres wide (e.g. Stokes and Clark, 2001). Their description as the arteries of ice sheets reflects their phenomenal drainage capacity and ability to substantially modify their topographic profile and lateral extent (Bennett, 2003). In the Anti-Atlas and the Tindouf Basin, isopach maps show 1) a pre-glacial depocentre dominated by clay-prone sediments that is centered over Central Anti-Atlas (Destombes et al., 1985) and 2) a syn-glacial depocentre oriented SSE-NNW. As the axis of both the pre-glacial and synglacial depocentres are aligned, these may preferentially have been occupied by a palaeo-ice stream at the glacial maximum (Fig. 13 A). For these reasons a palaeo-ice stream is tentatively placed over the Anti-Atlas at the height of the glaciation (Fig. 13). New data are needed to support this hypothesis, which for the moment can only be regarded as speculative.

### 6.2 *Location of the Meseta relative to the West African Craton*

Close correlation between the High Atlas and Rehamna is extremely significant because, structurally, the former area belongs to the West African Craton (Allary et al., 1972). Therefore, as the ice-sheet straddling the Meseta originated from the south, our data not only show the presence of ice in the Meseta (Rehamna) but demonstrate that it was in full palaeogeographic communication with the Anti-Atlas. These data imply that during the Late Ordovician, no large oceanic domain existed between the Meseta and Gondwana, supporting the conclusions of Piqué and Michard (1989). However, lateral displacement along the South Atlas Thrust may have occurred since the Ordovician, and the Late Ordovician location of the Meseta relative to the Anti-Atlas may have been shifted to some extent eastward or westward.

6.3 *Palaeohighs on the West Gondwana margin*

As discussed above, shelf margin deltas prograding S or SE-ward developed beyond the NW margin of the Gondwana ice-sheet. Deep marine systems occurred eastward in the Tazzeka domain, supporting the existence of an eastward deepening basin/ embayment between the Anti-Atlas and the northern Meseta, as proposed by previous workers (Hamoumi, 1988; Khoukhi and Hamoumi, 2001). To the west and/or north of the Coastal Meseta lay an upland area (Fig. 13). Addition data to support the occurrence of such a palaeohigh includes Middle to Late Ordovician hiatus accompanied by magmatic activity (Charlot et al., 1973; El Hassani et al., 1991; El Kamel et al. 1998). In the Sehoul zone, Silurian strata lap directly onto deformed Cambrian strata that are intruded by Middle Ordovician granite (Piqué et al., 1993). In addition, in Nova Scotia, which was a part of the West Gondwana margin in the Late Ordovician (the Meguma zone; e.g. Schenk, 1997), up to 10 km of volcanogenic deposits dated at $438^{+3/-2}$ Ma were deposited on lower Ordovician clastics after a 60 Ma hiatus (MacDonald et al., 2001). The alkaline affinity of these rocks implies an intraplate rift (MacDonald et al., 2001) and of comparable geochemistry to rarely occurring ash horizons in the Coastal Meseta (El Kamel et al. 1998). It is possible that the thermal event known in the Meseta (Clauer at al., 1995) could account for palaeohighs to the west of this area. Taken together, these observations suggest that upland areas in the westernmost and/ or northern Meseta might relate to rift shoulder uplift. However, the thickness of underflow deposits in the Tazzeka Massif aside, no evidence is found for a rift propagating directly into the Meseta is identified as suggested previously (e.g. Stampfli and Borel, 2002). Instead, the locus for rifting probably lay beyond the present-day Meseta.

### *6.4 Satellite ice caps*

In the context of uplands beyond the ice sheet, the occurrence of relatively proximal glaciomarine deposits in the uppermost part of the Late Ordovician succession in the Coastal Meseta is noteworthy. Here, glaciogenic lithofacies were not deposited during the main part of the glaciation because glacio-eustatic sea-level fall exposed the shelf leading to the development of a fluvial network that fed the shelf margin deltas (see above). However, following melting of the Hirnantian ice sheet, high sea levels were restored and backstepping shorelines may have allowed a marine ice front to form from individual ice caps centred on upland area that were probably more stable and long-lived than the huge, low-lying Gondwana ice sheet.

In Western Europe (Portugal, Spain, France), which lay at the fringes of Gondwana, widespread glaciomarine lithofacies are recognised (Robardet and Doré, 1988). To analyse these successions in detail is beyond the scope of this paper. However, the prevalence of sandy diamictite facies merits consideration because of their similarity to the topmost diamictites of the Coastal Meseta (Fig. 2). Earlier workers suggested glacier-ice or sea-ice rafting for the deposition of these rocks (Fortuin, 1984; Robardet and Doré, 1988). An alternative explanation could be the existence of several ice caps sited on upland areas during the Hirnantian. The recognition of widespread subglacial structures in these areas is unlikely as glaciers would have populated palaeohighs, leaving no sedimentary record on the uplands. However, marine ice fronts imply that valley glaciers or glacial lobes reached sea level. Localised subglacial deformation may have been preserved, but while striated pavements have been reported by previous workers (Arbey and Tamain, 1971), later authors prefer a tectonic interpretation for these structures (Robardet and Doré, 1988). Nevertheless, the existence of small ice caps is also inferred from the Late Ordovician glacial record in Newfoundland, an area that was located at notably lower palaeolatitudes but corresponding to island arcs in the Iapetus Ocean (MacNiocall et al., 1997).

**7. Conclusions**

1. Firm evidence that the Moroccan Meseta was in palaeogeographic communication with the Anti-Atlas during the Late Ordovician is provided by the discovery of glaciogenic deposits north of the Southern Atlas Thrust. Correlation between the Rehamna and the High Atlas of Marrakech demonstrates that the Anti-Atlas and Meseta were subject to a mutual phase of glaciation during the glacial maximum;
2. Evidence for the presence of former ice sheets includes soft-sediment striations, streamlined bedforms, downward-injected sedimentary dykes, shear zones and chaotic fold zones, capped, at least in the Rehamna inlier, by a distinctive diamictite. High-energy meltwater channels (jökulhlaup outburst channels) occur in the High Atlas;
3. A palaeogeographic transect over >400 km recognises ice-contact, shelf-margin and deep-marine proglacial depositional environments. The transition between these zones is not defined by a simple proximal-distal profile because an upland area persisted beyond the Coastal Meseta, potentially occupied by a satellite ice cap and possibly caused by rift-

shoulder uplift;

4. A deep marine system, represented by the Tazzeka Massif, lay in the eastern Meseta beyond the ice front. It was not overridden by an ice sheet as a result of the significant break in bathymetry that would have occurred in a north-eastward direction, and was probably fed from two sediment sources- namely the ice-contact and shelf-delta systems;
5. Two depositional units contain the record of four glacioeustatic sea level variations, thought to relate to the waxing and waning of other parts of the Gondwana ice sheet.

## Acknowledgements

Dan Le Heron was supported by TOTAL during his tenure as a research scientist. He was employed by the CNRS, France and based at École et Observatoire des Sciences de la Terre, Université Louis Pasteur, Strasbourg (UMR CNRS 7517). Age determinations and analyses of fossil material were kindly undertaken by Petr Storch (graptolites) of the Institute of Geology, Academy of Sciences, Czech Republic and Florentin Paris (chitinozoa) of Géosciences Rennes. We are especially grateful to Dr Philippe Razin for providing basemaps and key information on the best localities in the Massif Central and Rehamna, and to Professor Hassan El Chellai for his generous spirit and warm hospitality in Marrakech. This paper is a contribution to IGCP 503.

## References

Aghzer, A.M., Arenas, R., 1995. Détachement et tectonique extensive dans le massif hercynien des Rehamna, (Maroc). Journal of African Earth Sciences, 21 (3), 383-393.

Allary, A., Andrieux, J., Lavenu, A., Ribeyrolles, M., 1972. Les nappes hercyniennes de la

Meseta sud-orientale (Maroc central). Comptes Rendus de l'Academie des Sciences, Paris 274, 2284-2287.

Allary, A., Lavenu A., Ribeyrolles, M., 1976. Etude tectonique et microtectonique d'un segment de chaîne hercynienne dans la partie sud-orientale du Maroc central, Notes et Mémoires de la Service Géologique du Maroc 276, 169p.

Alvaro, J-J., Vennin, E., Villas, E., Destombes, J., Vizcaïno, D., 2004. Hirnantian valley-glacier sedimentation in the eastern Anti-Atlas, Morocco. International Symposium on Early Palaeozoic Palaeogeography and Palaeoclimate. Abstract volume, 18.

Arbey, F., Tamain., G., 1971. Existence d'une glaciation siluro-ordovicienne en Sierra Morena (Espagne). Comptes Rendus de l'Academie des Sciences, Paris 272, 1721-1723.

Armishaw, J., Holmes, R., Stow, D.A.V., 2000. The Barra Fan: A bottom-current reworked, glacially-fed submarine fan system. Marine and Petroleum Geology 17, 219-238.

Badalini, G., Redfern, J., Carr, I., 2002. Introduction: A Synthesis of current understanding of the Structural Evolution of North Africa. Journal of Petroleum Geology 25, 1-15.

Baudin, T., Chevremont, P., Razin, P., Theiblemont, D., Rachdi, H., Roger, J., Benhaourch, R., Winckel, A., 2001. Carte Géologique du Maroc au 1: 50,000 Feuille d'Oulmès- Mémoire Explicatif. Éditions du service géologique du Maroc, Rabat.

Bamoumen, H., 1988. Géométrie et cinématique de la déformation dans les nappes hercyniennes des Jebilet centre-orientales, Maroc. Thèse de 3ème Cycle, Université de Marrakech, 189p.

Bennett, M.R., 2003. Ice streams as the arteries of an ice sheet: their mechanics, stability and significance. Earth-Science Reviews 61 (3-4), 309-339.

Biju-Duval, B., 1974. Exemples de depots fluvioglaciares dans l'Ordovicien superieur du Sahara central. Bulletin Centre de Recherche de Pau. 8, 209-236.

Bernardin, C., Cornée J-J., Corsini, M., Mayol, S., Muller, J., Tayebi, M., 1988. Les variations d'épaisseurs du Cambrien moyen en Meseta occidentale. Données de surface et de subsurfaces. Signification géodynamique. Canadian Journal of Earth. Sciences 25, 2104-2117.

Beuf, S., Biju-Duval, B., de Charpal, O., Rognon, P., Gariel, O., Bennacef, A., 1971. Les Grés du Paléozoïque inférieur au Sahara. Paris, Editions Technip, 464p.

Bouabdelli, M., 1989. Tectonique et sédimentation dans un bassin orogénique, le sillon viséen d'Azrou-Khénifra (Est du Massif hercynien central du Maroc). Thèse d'Etat, Université Louis-Pasteur, Strasbourg, 262p.

Bourahrouh, A, Paris, F., Elaouad-Debbaj, Z., 2004. Biostratigraphy, biodiversity and palaeoenvironments of the chitinozoans and associated palynomorphs from the Upper Ordovician of the Central Anti-Atlas, Morocco. Review of Palaeobotony and Palynology 130, 17-40.

Brenchley, P.J., Marshall, J.D., Carden, G.A.F., Robertson, D.B.R., Long, D.G.F., Meidla, T., Hints, L., Anderson, T.F., 1994. Bathymetric and isotopic evidence for a short-lived Ordovician glaciation in a greenhouse period. Geology 22, 295-298.

Caritg, S., Burkhard, M., Ducommun, R., Helg, U., Kopp, L., Sue, C., 2004. Fold interference patterns in the Late Palaeozoic Anti-Atlas belt of Morocco. Terra Nova 16, 27-37.

Carrivick, J.L., Russell, A.J., Tweed, F.S., 2004. Sedimentology and palaeohydrology of jökulhlaups from Kverkfjöll, Iceland. Sedimentary Geology 172, 19-40.

Charlot, R., Rhalib, M., Tisserant, D., 1973. Étude géochronologique préliminaire des granites de la region du Rabat- Tiflet (Maroc occidental). Notes et mémoires du service géologique du

Maroc 249, 55-58.

Chèvremont, P., Cailleux, Y., Baudin, T., Razin, P., Thiéblemont, D., Hoepffner, C., Bensahal, A., Benhaourch, R., 2001. Carte Géologique du Maroc au 1: 50,000 Feuille de d'Ezzhiliga- Mémoire Explicatif. Éditions du service géologique du Maroc, Rabat.

Clauer, N., Rais, N., Schaltegger, U., Piqué, A., 1995. K-Ar systematics of clay-to-mica minerals in a multi-stage low-grade metamorphic evolution. Chemical Geology 124, 305-316.

Cornée, J-J., Destombes, J., Willefert, S., 1987. Stratigraphie du Paléozoïque de l'extrémité nord-ouest du Haut-Atlas occidental (Maroc hercynien); interprétation du cadre sédimentaire du Maroc occidental. Bulletin de la Societé geologique de France 8, 327-335.

Destombes, J., 1968. Sur la presence d'une discordance generale de ravinement d'age Ashgill superieur dans l'Ordovicien terminal de l'Anti-Atlas (Maroc). Comptes Rendus de l'Academie des Sciences, Paris 267, 565-567.

Destombes, J., 1971. L'Ordovicien au Maroc. Essai de synthèse stratigraphique. Mémoires de la B.R.G.M, Paris 73, 237-263.

Destombes, J., Holland, H., Willefert, S., 1985. Lower Paleozoic rocks of Morocco. In: Holland, C. H. (ed) Lower Palaeozoic rocks of the World. Wiley, London, pp.91-336.

Destombes, J., Willefert, S., 1988. The Ordovician-Silurian boundary in Morocco. Bulletin of the British Museum of Natural History (Geology) 43, 165-170.

Deynoux, M., 1985. Terrestrial or waterlain glacial diamictites? Three case studies from the Late Precambrian and Late Ordovician glacial drifts in West Africa. Palaeogeography, Palaeoclimatology, Palaeoecology 51, 97-141.

Deynoux, M. and Ghienne, J-F., 2004. Late Ordovician glacial pavements revisited: a reappraisal of the origin of striated surfaces. Terra Nova 16, 95-101.

El Attari, A., 2001. Etude lithostratigraphique et tectonique des terrains paléozoïques du Môle Côtier (Meseta Orientale, Maroc). Thèse d'Etat, Université Mohammed V Agdal, Rabat, 394p.

El Attari, A., Hoepffner, C., Jouhari, A. 1997. Nouvelles données magmatiques et structurales en relation avec la cinématique de l'ouverture du bassin cambrien de la Meseta occidentale (Maroc). Gaia 14, 11-21.

El Hassani, A., Huon, S., Hoepfnner, C., Whitechurch, H., Piqué, A., 1991. Une déformation d'âge ordovicien moyen dans la zone des Sehoul (Meseta marocaine septentrionale). Regard sur les segments "calédoniens" au NW de l'Afrique. Comptes Rendus de l'Academie des Sciences, Paris 1027-1032.

El Houicha, M., 1994. Dynamique d'un bassin orogénique et processus gravitaires syn-à-tadi-sédimentaires: le sillon viséo-namurien de Khénifra (SE du Massif hercynien central du Maroc). Thèse de 3ème Cycle, Université de Marrakech, 170p.

El Kamel, F., Remmal, T., Mohsine, A., 1998. Evidence of a post Caledonian withinplate alkaline magmatism in the Ouled Abbou Silurian basin (coastal Meseta, Morocco). Comptes Rendus de l'Academie des Sciences, Paris 327, 309-314.

Escutia, C., Eittreim, A.K., Cooper, A.K., Nelson, C.H., 2000. Morphology and acoustic character of the Antarctic Wilkes Land turbidite systems: ice-sheet-sourced versus river-sourced fans. Journal of Sedimentary Research 70, 84-93.

Fabre, J., 1971. Structure de l'Algérie saharienne. Eléments pour une carte tectonique. Bulletin du Service géologique d'Algérie 41, 161-185.

Fortuin, A.P., 1984. Late Ordovician glaciomarine deposits (Orea Shale) in the Sierra de

Albarracin, Spain. Palaeogeography, Palaeoclimatology, Palaeoecology 48, 245-261.

Ghienne, J-F., 2003. Late Ordovician sedimentary environments, glacial cycles, and post-glacial transgression in the Taoudeni Basin, West Africa. Palaeogeography, Palaeoclimatology, Palaeoecology 189, 117-145.

Ghienne, J.F., Deynoux, M., 1998. Large-scale channel fill structures in Late Ordovician glacial deposits in Mauritania, western Sahara. Sedimentary Geology 119, 141-159.

Ghienne, J-F., Deynoux, M., Manatschal, G., Rubino, J-L., 2003. Palaeovalleys and fault-controlled depocentres in the Late-Ordovician glacial record of the Murzuq Basin (central Libya). Comptes Rendus Geosciences 335, 1091-1100.

Hamoumi, N., 1988. La plate-forme ordovicienne du Maroc: dynamique des ensembles sédimentaires. Thèse de 3ème Cycle, Université Louis Pasteur, Strasbourg, 237p.

Hart, J.K., Boulton, G.S., 1991. The interrelation of glaciotectonic and glaciodepositional processes within the glacial environment. Quaternary Science Reviews 10, 335-350.

Hirst, J.P.P., Benbakir, A., Payne, D.F., Westlake, I.R., 2002. Tunnel Valleys and Density Flow Processes in the upper Ordovician glacial succession, Illizi Basin, Algeria: influence on reservoir quality. Journal of Petroleum Geology 25, 297-324.

Hoepffner, C., 1974. Contribution a la géologie structurale des Rehamna (Meseta marocaine méridionale): le matériel Paléozoïque et son évolution Hercynienne dans l'est du massif. Thèse de 3ème Cycle, Université Louis Pasteur, Strasbourg, 91p.

Hoepffner, C., 1987. La tectonique hercynienne dans l'Est du Maroc. Thèse d'Etat, Université Louis Pasteur, Strasbourg, 280p.

Hollard H., Huvelin P., Mamet B., 1977. Stratigraphie du Viséen supérieur des Jebilets et âge de la mise en place de la nappe des Jebilets orientales (Maroc). Notes et Mémoires du Service Géologique du Maroc 37 (267), 7-22.

Huvelin, P., 1977. Etude Geologique et Gitologique du Massif Hercynien des Jebilet (Maroc Occidental). Editions du service géologique du Maroc 232, 307p. Rabat.

Hyam, D.M., Marshall, J.E.A., Sanderson, D.J., 1997. Carboniferous dykes in the Falkland Islands. Journal of African Earth Sciences 25, 505-517.

Jansson, K.N., Glasser, N., 2005. Palaeoglaciology of the Welsh sector of the British-Irish Ice Sheet. Journal of the Geological Society, London 162, 25-37.

Jenny, J., 1983. L'avant-pays présaharien (Anti Atlas central) et la partie méridionale de la chaîne hercynienne du Maroc. In: Caledonian Orogeny IGCP n°27, Rabat symposium. Excursion guide B5, 82-99. Service Géologique du Maroc.

Khoukhi, Y., 1993, L'Ordovicien et le Cambro-ordovicien de la Meseta orientale, Maroc: pétrologie, dynamique sédimentaire et stratigraphie séquentielle. Thèse de 3ème Cycle, Faculté des Sciences d'Oujda, Morocco.

Khoukhi, Y., Hamoumi, N., 2001. L'Ordovicien de la Meseta orientale (Maroc): stratigraphie génétique- contrôle géodynamique, climatique et eustatique. Africa Geoscience Review 8, 289-302.

Kor, P.S.G., Shaw, J., Sharpe, D.R., 1991. Erosion of bedrock by subglacial meltwater, Georgian Bay, Ontario: a regional view. Canadian Journal of Earth Sciences 28, 623-642.

Lagarde, J.-L., Michard, A., 1986. Stretching normal to the regional thrust displacement in a thrust wrench shear zone, Rehamna massif, Morocco. Journal of Structural Geology 8 (3/4), 483-492.

Laville, E., Piqué, A., 1991. La distension crustale atlantique et atlasique au Maroc au début du Mésozoïque: le rejeu des structures hercyniennes. Bulletin de la Société géologique de France 162, 1161–1171.

Le Heron, D., Sutcliffe, O., Bourgig, K., Craig, J., Visentin, C., Whittington, R., 2004. Sedimentary architecture of Upper Ordovician tunnel valleys, Gargaf Arch, Libya: Implications for the genesis of a hydrocarbon reservoir. GeoArabia 9, 137-160.

Le Heron, D.P., Sutcliffe, O.E., Whittington, R.J., Craig, J., 2005. The origins of glacially related soft-sediment deformation structures in Upper Ordovician glaciogenic rocks: implication for ice sheet dynamics. Palaeogeography, Palaeoclimatology, Palaeoecology 218, 75-103.

Lüning, S., Craig, J., Loydell, D.K., Storch, P., Fitches, W., 2000. Lowermost Silurian "hot shales" in north Africa and Arabia: regional distribution and depositional model. Earth Science Reviews 49, 121-200.

MacDonald, L.A., Barr, S.M., White, C.E., Ketchum, J.W.F., 2001. Petrology, age, and tectonic setting of the White Rock Formation, Meguma terrane, Nova Scotia: evidence for Silurian continental rifting. Canadian Journal of Earth Sciences 39, 259-277.

MacNiocaill, C., van der Pluijm, B.A., van der Voo, R., 1997. Ordovician paleogeography and the evolution of the Iapetus Ocean. Geology 25, 159-162.

Moreau, J., Ghienne, J-F., Le Heron, D.P., Rubino, J-L., Deynoux, M., Accepted. A 440 Million year old ice stream in North Africa. Geology.

Monod, O., Kozlu, H., Ghienne, J-F., Dean, W.T., Günay, Y., Le Hérissé, A., Paris, F., Robardet, M., 2003. Late Ordovician glaciation in southern Turkey. Terra Nova 15 (4), 249-257.

Mutti, E., Davoli, G., Tinterri, R., Zavala, C., 1996. The importance of Ancient Fluvio-deltaic

Systems dominated by Catastrophic Flooding in Tectonically Active Basins. Memorie di scienze geologiche, Padova 48, 233-291.

Mutti, E., Tinterri, R., Remacha, E., Mavilla, N., Angella, S., Fava, L., 1999. An Introduction to the Analysis of Ancient Turbidite Basins from an Outcrop Perspective. American Association of Petroleum Geologists Continuing Education Course notes series 39. Oklahoma, USA.

Nio, S-D., Yang, C-S., 1991. Diagnostic attributes of clastic tidal deposits: a review. In: Smith, D.G., Reinson, B., Zaitlin, B.A., Rahmani, R.A. (Eds.) Clastic Tidal Sedimentology. Canadian Society of Petroleum Geologists Memoir 16, 3-28.

Ouanami, H., 1989. Évolution sédimentaire et tectonique de la partie orientale du Massif Ancien du Haut Atlas (Maroc). Thèse de 3ème Cycle, Université de Montpelier-II.

Ouanami, H., 1998. Le passage Ordovicien-Silurien à Tizi n'Tichka (Haut-Atlas, Maroc): variations du niveau marin. Comptes Rendus de l'Académie des Sciences, Paris 326, 65-70.

Ouanami, H., Petit, J-P., 1992. La limite sud de la chaîne hercynienne dans le Haut Atlas marocain: reconstitution d'un saillant non déforme. Bulletin de la Société Géologique de France 162, 63-72.

Paris, F., Elaouad-Debbaj, Z., Jaglin, J.C., Massa, D., Oulebsir, L., 1995. Chitinozoans and late Ordovician glacial events on Gondwana. In: Cooper, J.D., Droser, M.L., Finney, S.C. (Eds). Ordovician Odyssey: Short Papers for the Seventh International Symposium on the Ordovician System, 171-177. Fullerton, California: The Pacific Section Society for Sedimentary Geology (SEPM).

Paris, F., Deynoux, M., and Ghienne, J.F., 1998. Découverte de chitinozoaires à la limite Ordovicien-Silurien en Mauritanie); implications paléogéographiques. Comptes Rendus de l'Académie des Sciences, Paris 326, 499-504.

Piqué, A., 1979. Evolution structurale d'un segment de la chaîne hercynienne: la Meseta marocaine nord-occidentale. Thèse d'Etat, Université Louis Pasteur, Strasbourg, 253p.

Piqué, A., 2001. Geology of Northwest Africa. Gebrueder Borntraeger Verlagsbuchhandlung. Berlin-Stuttgart, Federal Republic of Germany, 324p.

Piqué, A., Bouabdelli, M., Darboux, R. 1995. Le rift cambrien du Maroc occidental. Comptes Rendus de l'Académie des Sciences, Paris 320, 1017-1024.

Piqué, A., El Hassani, A., Hoepffner, C., 1993. Les déformations ordoviciennes dans la zone des Sehoul (Maroc septentrional): une orogenèse calédonienne en Afrique du Nord. Canadian Journal of Earth Sciences 30, 1332-1337.

Piqué, A., Jeannette, D., Michard, A., 1980. The western Meseta shear zone, a major and permanent feature of the Hercynian belt in Morocco. Journal of Structural Geology 2 (1/2), 55-61.

Piqué, A., O'Brien, S., King, A.F., Schenk, P.E., Skehan, J.W., Hon, R., 1990. La marge nord-occidentale du Paleo-Gondwana (Maroc occidental et zones orientales des Appalaches); Rifting au Précambrien terminal et au Paléozoïque inférieur, et compression hercynienne et alleghanienne au Paléozoique supérieur. Comptes Rendus de l'Académie des Sciences, Paris 310, 411-416.

Piqué, A., Michard, A., 1989. Moroccan Hercynides: A Synopsis. The Paleozoic sedimentary and tectonic evolution at the northern margin of West Africa. American Journal of Science 289, 286-330.

Piqué, A., Tricart, P., Guiraud, R., Laville, E., Bouaziz, S., Amrhar, M., Aït Ouali, R., 2002. The

Mesozoic–Cenozoic Atlas belt (North Africa): an overview. Geodinamica Acta 15, 185–208.

Porębski, S.J., Steel, R., 2003. Shelf-margin deltas: their stratigraphic significance and relation to deepwater sands. Earth Science Reviews 62, 283-326.

Punkari, M., 1997. Glacial and glaciofluvial deposits in the interlobate areas of the Scandinavian Ice Sheet. Quaternary Science Reviews, Special Issue 16, 12-26.

Razin, P., Janjou, D., Baudin., T., Benschal., A., Hoepffner, C., Thiéblemont, D., Chèvremont, P., Benhaourch, R., 2001. Carte Géologique du Maroc au 1: 50, 000 Feuille de Sidi Matla' Ech Chams- Mémoire Explicatif. Éditions du service géologique du Maroc, Rabat.

Robardet, M., Doré, F., 1988. The late Ordovician diamictic formations from southwestern Europe: North-Gondwana glaciomarine deposits. Palaeogeography, Palaeoclimatology, Palaeoecology 66, 19-31.

Schenk, P.E. 1997. Sequence stratigraphy and provenance on Gondwana's margin: the Meguma Zone (Cambrian to Devonian) of Nova Scotia, Canada. Geological Society of America Bulletin 109 (4), 395-409.

Selley, R.C., 1997. The Basins of NW Africa: Structural Evolution. In: Selley, R.C. (Ed). African Basins: Sedimentary Basins of the World 3. Elsevier, Amsterdam, 17-26.

Shanmugam, G., 2000. 50 years of the turbidite paradigm (1950s—1990s): deep-water processes and facies models-a critical perspective. Marine and Petroleum Geology 17, 285-342.

Soulaimani, A., 1998. Interactions socle/couverture dans l'Anti Atlas occidental (Maroc): rifting fini-protérozoïque et orogenèse hercynienne. Thèse d'Etat, Université de Marrakech, Morocco, 189p.

Stampfli, G.M., Borel, G.D., 2002. A plate tectonic model for the Palaeozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrons. Earth and Planetary Science Letters 196, 17-33.

Stokes, C.R., Clark, C.D., 2001. Palaeo-ice streams. Quaternary Science Reviews 20, 1437-1457.

Stow, D.A.V. & Johansson, M., 2000.Deep-water massive sands: nature, origin and hydrocarbon implications. Marine and Petroleum Geology 17, 145-174.

Stow, D.A.V., Mayall, M., 2000. Deep-water sedimentary systems: New models for the 21st century. Marine and Petroleum Geology 17, 125-135.

Sutcliffe, O.E., Dowdeswell., J.A., Whittington., R.J., Theron, J.N., Craig, J., 2000. Calibrating the Late Ordovician glaciation and mass extinction by the eccentricity cycles of the Earth's orbit. Geology 23, 967-970.

Sutcliffe, O.E., Harper, D.A.T., Aït Salem, A., Whittington, R.J., Craig, J., 2001. The development of an atypical Hirnantia-brachiopod Fauna and the onset of glaciation in the late Ordovician of Gondwana. Transactions of Royal Society of Edinburgh, Earth Sciences 92, 1-14.

Taylor, J., Dowdeswell, J.A., Kenyon, N.H., O'Cofaigh., 2002. Late Quaternary architecture of trough-mouth fans: debris flows and suspended sediments on the Norwegian margin. In: Dowdeswell, J.A., O'Cofaigh, C. (Eds). Glacier-Influenced Sedimentation on High-Latitude Continental Margins. Geological Society, London, Special Publication 203, 55-71.

Underwood, C.J., Deynoux, M., Ghienne, J.-F., 1998. High palaeolatitude (Hodh, Mauritania) recovery of graptolite fauna after the Hirnantian (end Ordovician) extinction event. Palaeogeography, Palaeoclimatology, Palaeoecology 142, 91-105.

Van der Wateren, F.M., Kluiving, S.J., Bartek, L.R., 2000. Kinematic indicators of subglacial shearing. In Maltman, A.J., Hubbard, B. and Hambrey, M.J. (Eds). Deformation of Glacial materials. Geological Society, London, Special Publication 176, 259-278.

**Figure Captions:**

Figure 1. Location map of Morocco. Lower Palaeozoic inliers in the Meseta and High Atlas regions are shown in black, whereas those in the Anti-Atlas are shown in grey. Numbers indicate inliers examined during the course of this study. Contours are isopachs drawn for the pre-glacial upper Ordovician (Ashgill) Upper Ktaoua Formation in the Anti-Atlas (from Destombes et al., 1985). These data give a good impression of topography on the pre-glacial substrate (axis of depositional trough is marked). Previous estimation of the late Ordovician glacial maximum in this region put the ice front in the Anti Atlas (stippled line; Destombes, 1968). Control data points in the Anti Atlas are the Zagora/ Tagounite areas (A) and the Erfoud area (B).

Figure 2. Correlation panel between sedimentary logs for A) the High Atlas of Marrakech (Isk N'Tazzoult; GPS 31°18.299'N 07°20.729'W), B) Rehamna (Amar Ben Nouti; GPS 32°25.577'N 07°31.072'W), C) Massif Central (Mchmech El Diab; 33°16.045'N 06°26.01'W) and D) Coastal Meseta (Oued Nfifkh; 33°33.703'N 07°11.474'W) areas.

Figure 3. Detailed section of syn-glacial deposits in the Rehamna inlier, approximately 1 km north-west of the log shown in Fig. 2 B, to highlight stratigraphic position of soft-sediment deformation structures and their relationship to the distinctive red-brown diamictite. This section corresponds to the interval indicated on Fig. 2 B.

Figure 4. Glaciogenic facies at the top of unit 1 in the Rehamna inlier at Amar Ben Nouti (taken from one-kilometre radius centred on 32°25.577'N 07°31.072'W). A: Distinctive brown-red clast-poor sandy diamictite, preserving the counterpart to a pair of soft-sediment striae (C). B: Complex and irregular white sandstone dykes sourced from directly above the diamictite and intruded down into it. D: Detail of centimetre-scale shear zones developed within the white sandstone bed that sources the dykes. Together, the soft-sediment striae, downward-injected sedimentary dykes, and massive diamictite provide strong evidence of a subglacial erosion surface (GES) in the Rehamna; see text for details. E: Sigmoidal cross-bedding in tidal sandstones at the top of unit 2.

Figure 5. Large-scale channel systems and soft-sediment deformation at Isk N'Tazzoult, High Atlas of Marrakech. Two glacial erosion surfaces (GES) can be recognised in this section. The subglacial bedforms (Fig. 6) occur along the second, higher, surface. Note chaotic internal organisation of lithofacies. Significant lateral facies variations occur over short distances.

Figure 6. A: Large boulder of rippled sandstone within a pro- or subglacial channel at Isk N'Tazzoult, High Atlas of Marrakech. B: Polished drumlin-like bedform. C: Large "hairpin-like" structure on the polished surface of a folded drumlin. These structures are interpreted to have formed subglacially, with the last structure characteristic of the hydroplastic deformation of unconsolidated sediment.

Figure 7: Large sandstone channels within clast-poor sandy diamictites at Larissa (31°56.194'N 07°39.980'W), Jbilet inlier, constituting the thickest uninterrupted example of these facies (150

m) observed in Morocco.

Figure 8: Syn-glacial lithofacies in the Massif Central. A: Diamictites in depositional unit 1 at Mchmech El Diab (see Fig. 2 C for GPS co-ordinates), showing interbedding of massive and thinly bedded variants of this lithofacies. B: Examples of lonestones within massive, clast-poor sandy diamictites considered to have been deposited by glaciogenic debris flows. C: Section view at Mchmech El Diab from poorly exposed diamictite to overlying sandstone. D (and E, close-up): Well defined subdivision of the Ezzhiliga Formation (top of depositional unit 1 to unit 2) into two coarsening-up motifs developed exclusively in sandstone at El Harcha. This subdivision and the sharp surface separating these motifs is developed widely in the Massif Central. The sharp contact between the Ezzhiliga Formation and underlying diamictite implies that the former does not simply cap a coarsening-upward unit. GPS co-ordinates for El Harcha: 33º30.830'N 06º08.690'W.

Figure 9: Measured section of latest Ordovician strata at Zerarda, Tazzeka Massif (Eastern Meseta; 34º 01. 545'N  04º20.148'W). As in other parts of the Meseta, two units are recognised. Numbers on the log correspond to major phases in the evolution of the Tazzeka fan, which are directly attributed to ice sheet waxing and waning on the shelf. Minor coarsening and fining-upward motifs are probably autocyclic in nature. The package marked "GES" is thought to be coeval with glacial erosion next to the ice sheet, resulting from bypass of the shelf margin (see Fig. 13).

Figure 10: Panoramic photograph of the logged Zerarda section showing generally high quality of exposure and stratigraphic continuity, annotated for direct comparison with Fig. 9. The

interpreted pre-glacial to syn-glacial transition is shown, together with transition of these rocks into lowermost Silurian shale.

Figure 11: Sedimentary facies at Zerarda, Tazzeka Massif. A: Well-defined, sharp based sandstones interbedded with mudrocks form metre to tens of metre-scale coarsening upward motifs. These are probably due to allocyclic phases of fan lobe growth and abandonment. B: View looking down onto the top of unit 1, represented by an amalgamated package of sandstone (Fig. 9, 172 m). Clearly visible on this surface (illustrated in C) are large flutes. D: Close-up of these features, with hammer for scale, showing gravel lag in the central depressed part of a flute.

Figure 12: Synthesis of stratigraphic architecture shown on correlation panel (Fig. 2), emphasising the occurrence of four glacially-related cycles in northern Morocco. The ice maximum is represented by the GES at the top of unit 1.

Figure 13: A: Palaeogeographic reconstruction of the northwestern fringe of Africa during the late Ordovician glacial maximum, showing extension of the ice front across the Anti-Atlas and into the Moroccan Meseta. A shallow marine area, occupied by shelf margin deltas, occurred just in front of the ice sheet. Limited palaeocurrent data imply a palaeohigh towards the northwest, potentially an extension of Nova Scotian upland. These two sediment sources probably commingled in the deep basin, which received a thick succession of underflow deposits. B: Block diagram/ depositional model for northern Morocco at the late Ordovician glacial maximum. The linkage and inter-relationship of both the ice-contact and shelf delta environments and the deep marine system is emphasised in the text.

Table 1. Simplified stratigraphic comparison of formation names for pre-glacial, syn-glacial and post-glacial strata in Morocco. Formalised stratigraphies exist in the Anti Atlas (Destombes et al., 1985), Massif Central (e.g. Razin et al., 2001) and Tazzeka Massif (Destombes et al., 1985; Khoukhi and Hamoumi, 2001). A stratigraphic scheme awaits formalisation in the Rehamna where an active mapping programme is currently underway (P. Razin, pers. comm.). An informal formation name (Douar-Zrahna formation) for syn-glacial strata was suggested for the Coastal Meseta region by Sutcliffe et al. (2001). No formalised stratigraphy exists for either the High Atlas of Marrakech or Jbilet regions.

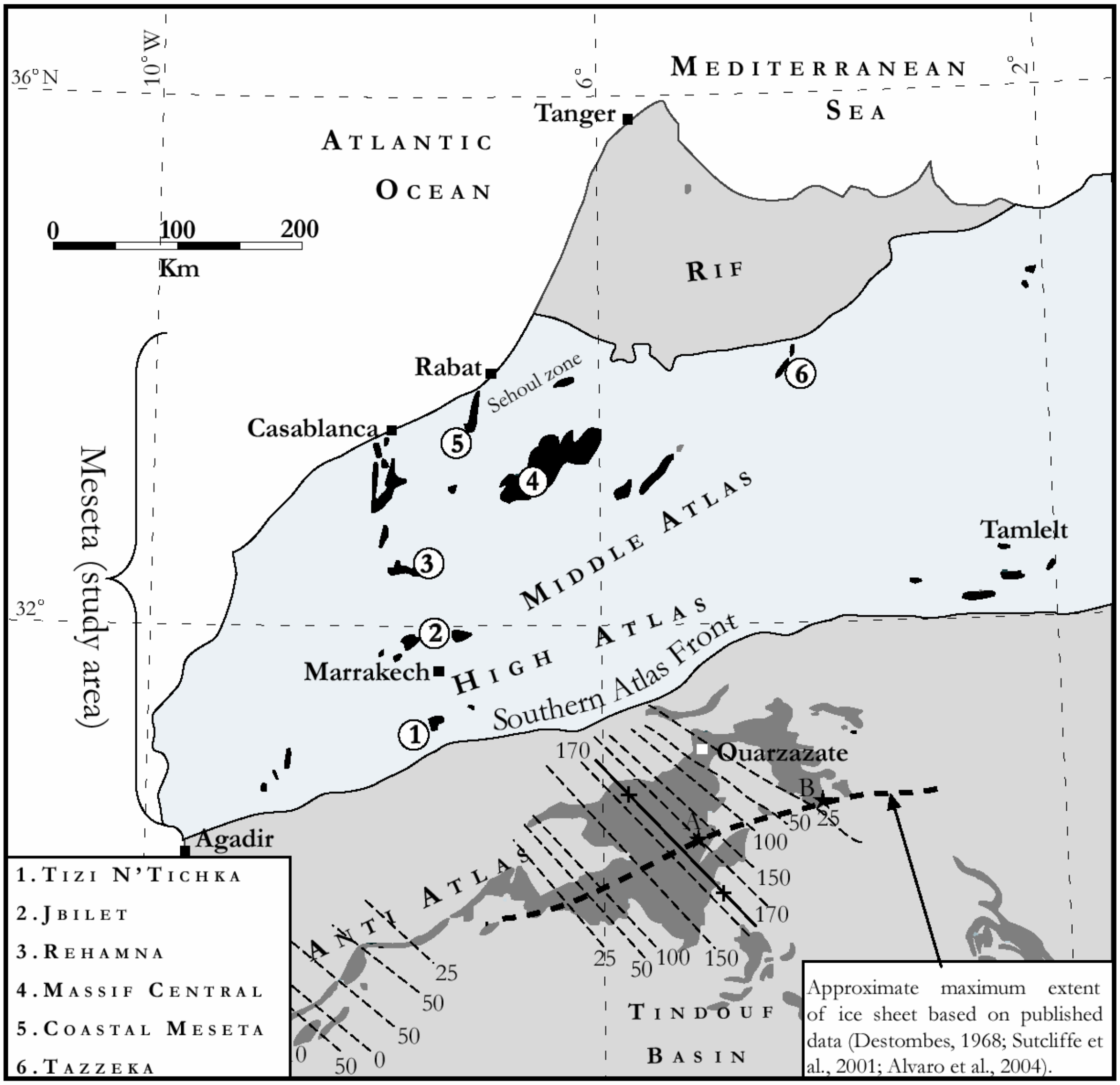

36° N
10° W
6°
2°
32°
MEDITERRANEAN SEA
ATLANTIC OCEAN
Tanger
0 100 200
Km
RIF
Rabat
Sehoul zone
Casablanca
Meseta (study area)
MIDDLE ATLAS
Tamlelt
HIGH ATLAS
Marrakech
Southern Atlas Front
Quarzazate
Agadir
ANTI ATLAS
TINDOUF BASIN
1. TIZI N'TICHKA
2. JBILET
3. REHAMNA
4. MASSIF CENTRAL
5. COASTAL MESETA
6. TAZZEKA
Approximate maximum extent of ice sheet based on published data (Destombes, 1968; Sutcliffe et al., 2001; Alvaro et al., 2004).

**Figure 1**

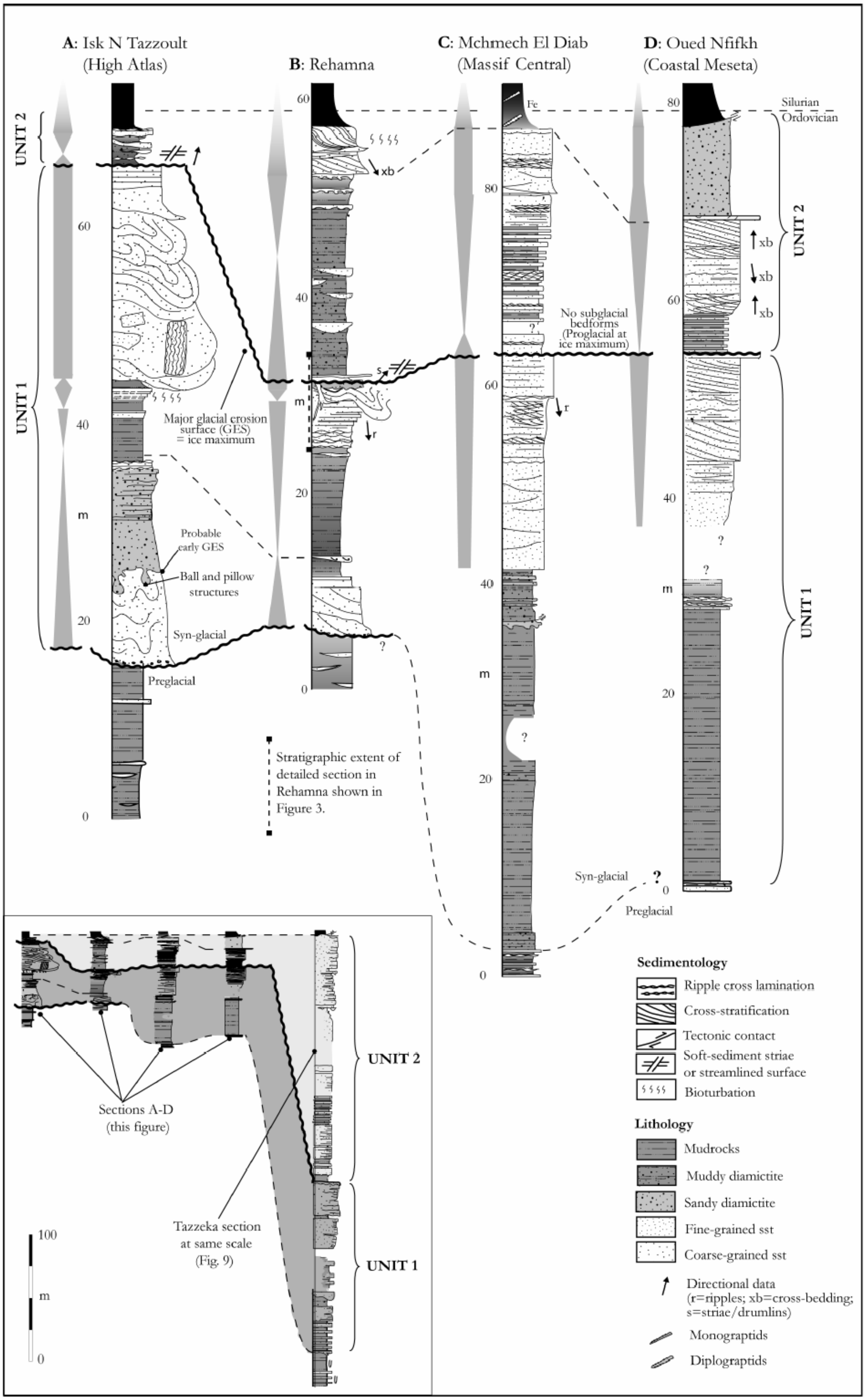

A: Isk N Tazzoult (High Atlas)
B: Rehamna
C: Mchmech El Diab (Massif Central)
D: Oued Nfifkh (Coastal Meseta)
UNIT 2
UNIT 1
Silurian
Ordovician
Major glacial erosion surface (GES) = ice maximum
Probable early GES
Ball and pillow structures
Syn-glacial
Preglacial
No subglacial bedforms (Proglacial at ice maximum)
Stratigraphic extent of detailed section in Rehamna shown in Figure 3.
Sections A-D (this figure)
Tazzeka section at same scale (Fig. 9)
Sedimentology
Ripple cross lamination
Cross-stratification
Tectonic contact
Soft-sediment striae or streamlined surface
Bioturbation
Lithology
Mudrocks
Muddy diamictite
Sandy diamictite
Fine-grained sst
Coarse-grained sst
Directional data (r=ripples; xb=cross-bedding; s=striae/drumlins)
Monograptids
Diplograptids

**Figure 2**

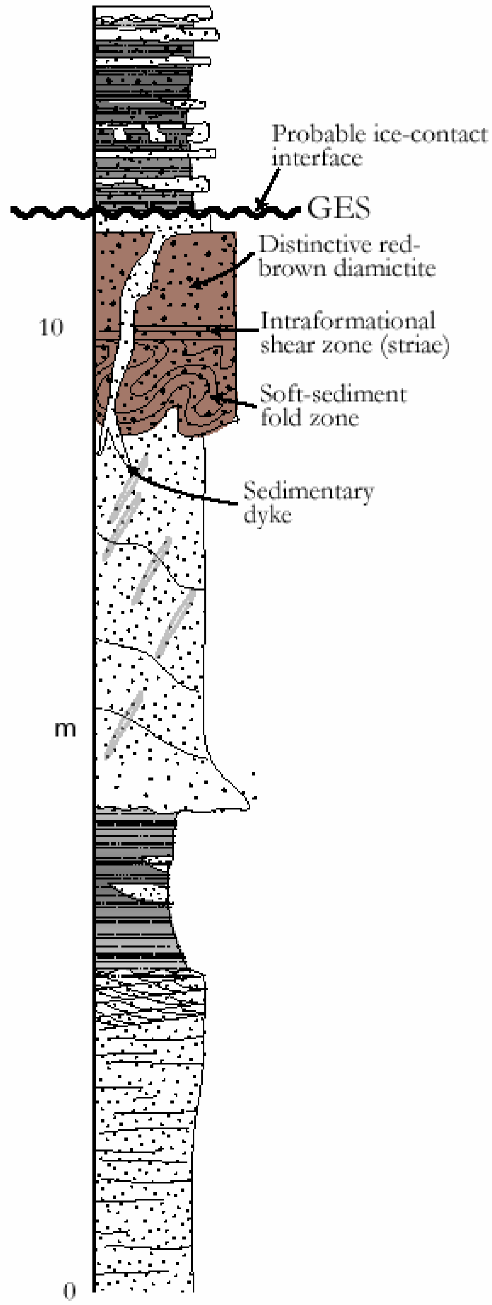

Probable ice-contact interface
GES
Distinctive red-brown diamictite
Intraformational shear zone (striae)
Soft-sediment fold zone
Sedimentary dyke
10
m
0

**Figure 3**


A
Overturned bedding
B
C
B
Sandy diamictite (host sediment)
Overturned bedding
Intrusion
Source bed
C
E
Way up
D

**Figure 4**

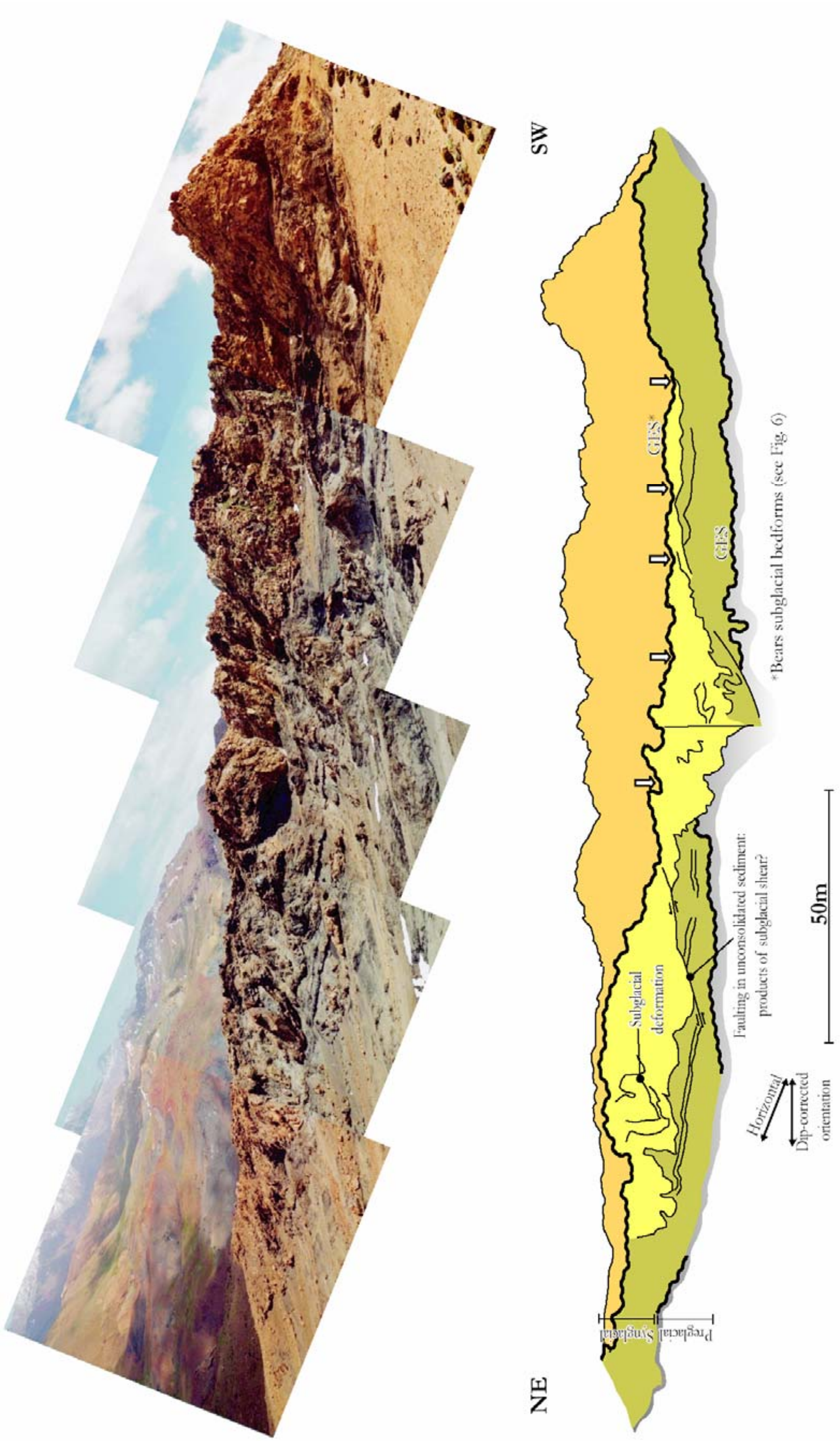

SW
NE
GES*
GES
*Bears subglacial bedforms (see Fig. 6)
Faulting in unconsolidated sediment:
products of subglacial shear?
Subglacial
deformation
50m
Horizontal
Dip-corrected
orientation
Preglacial
Synglacial

**Figure 5**


A
Sandstone block
B
Long axis of streamlined bedform
C
Direction of ice advance

**Figure 6**

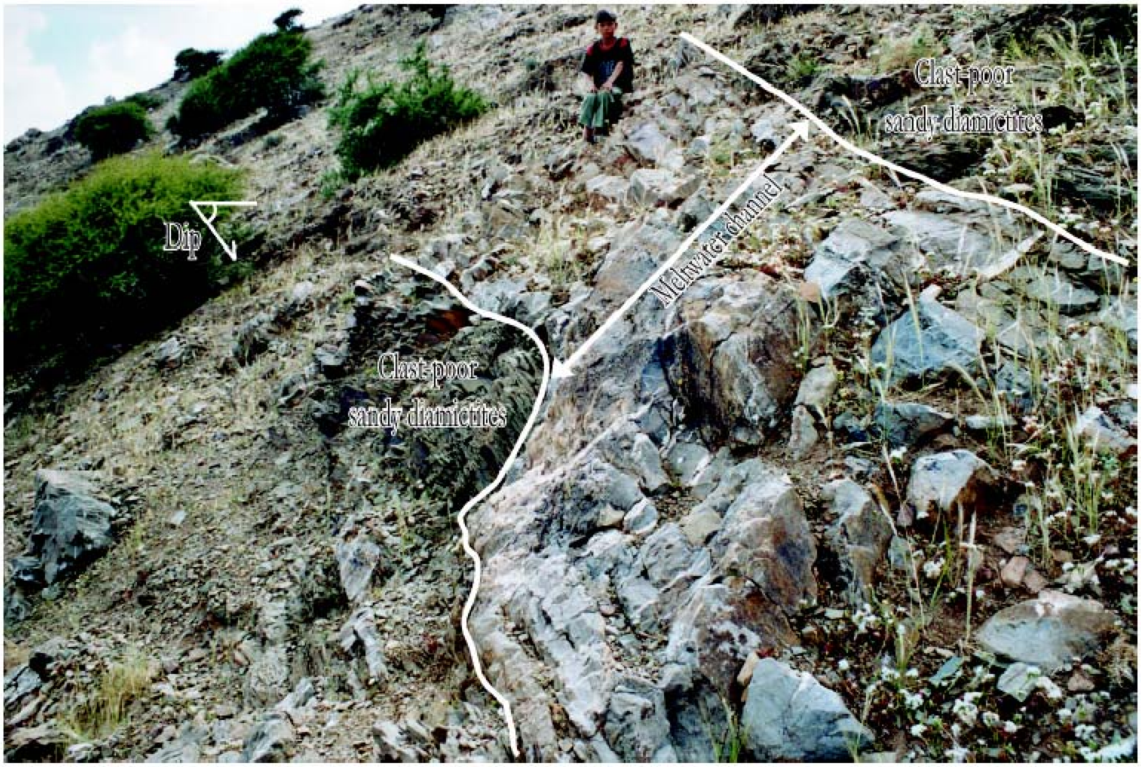

Clast-poor
sandy diamictites
Dip
Meltwater channel
Clast-poor
sandy diamictites

**Figure 7**

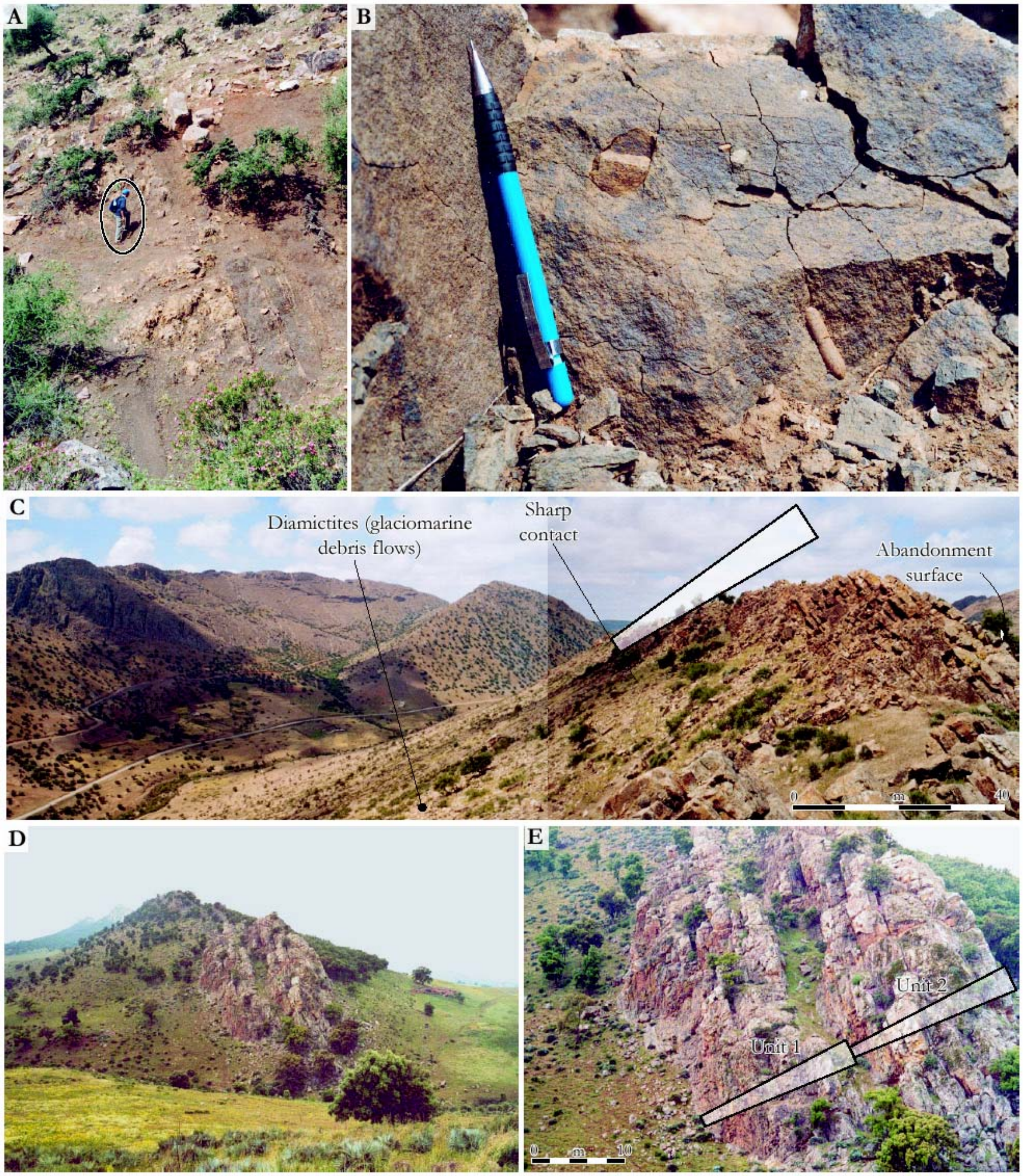
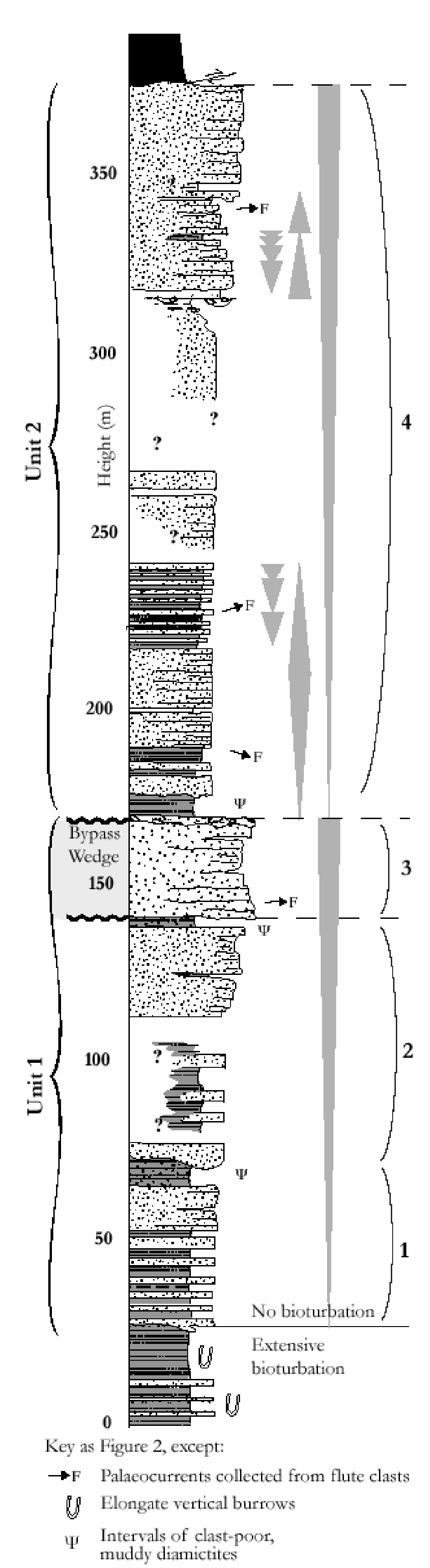

A
B
C
Diamictites (glaciomarine
debris flows)
Sharp
contact
Abandonment
surface
0
m
40
D
E
Unit 2
Unit 1
0
m
10

**Figure 8**



Unit 2
Unit 1
Height (m)
350
300
250
200
150
100
50
0
Bypass
Wedge
F
?
Ψ
4
3
2
1
No bioturbation
Extensive
bioturbation
Key as Figure 2, except:
F Palaeocurrents collected from flute clasts
Elongate vertical burrows
Ψ Intervals of clast-poor,
muddy diamictites


**Figure 9**

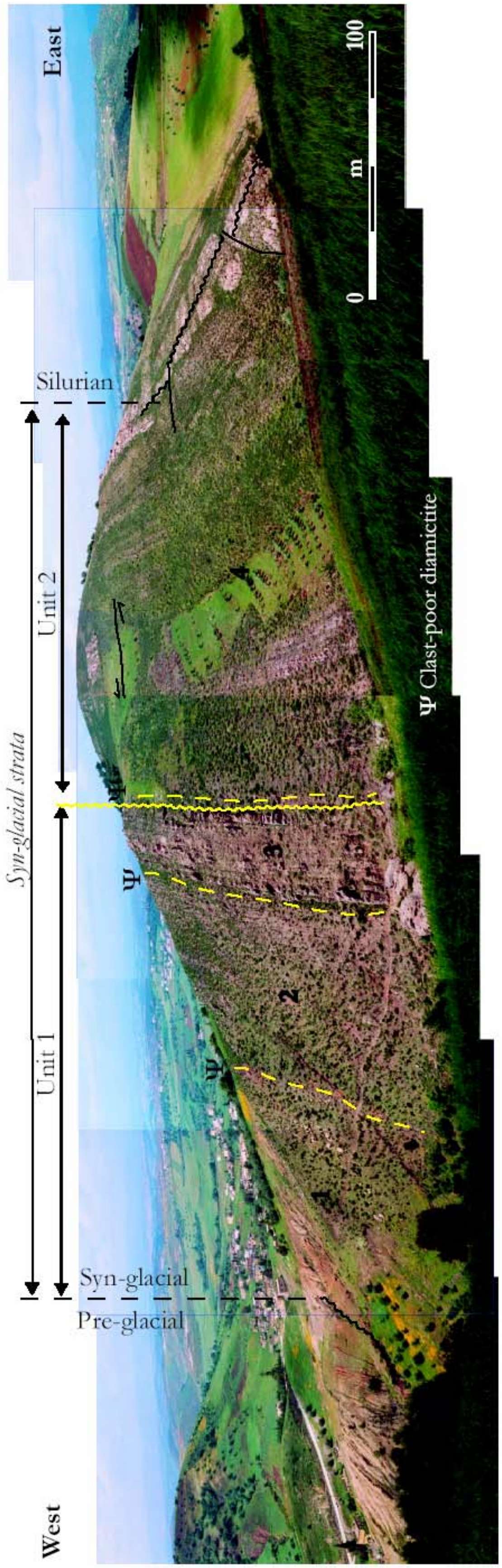

East
West
Silurian
Unit 2
Unit 1
*Syn-glacial strata*
Syn-glacial
Pre-glacial
Ψ Clast-poor diamictite
0
m
100

**Figure 10**

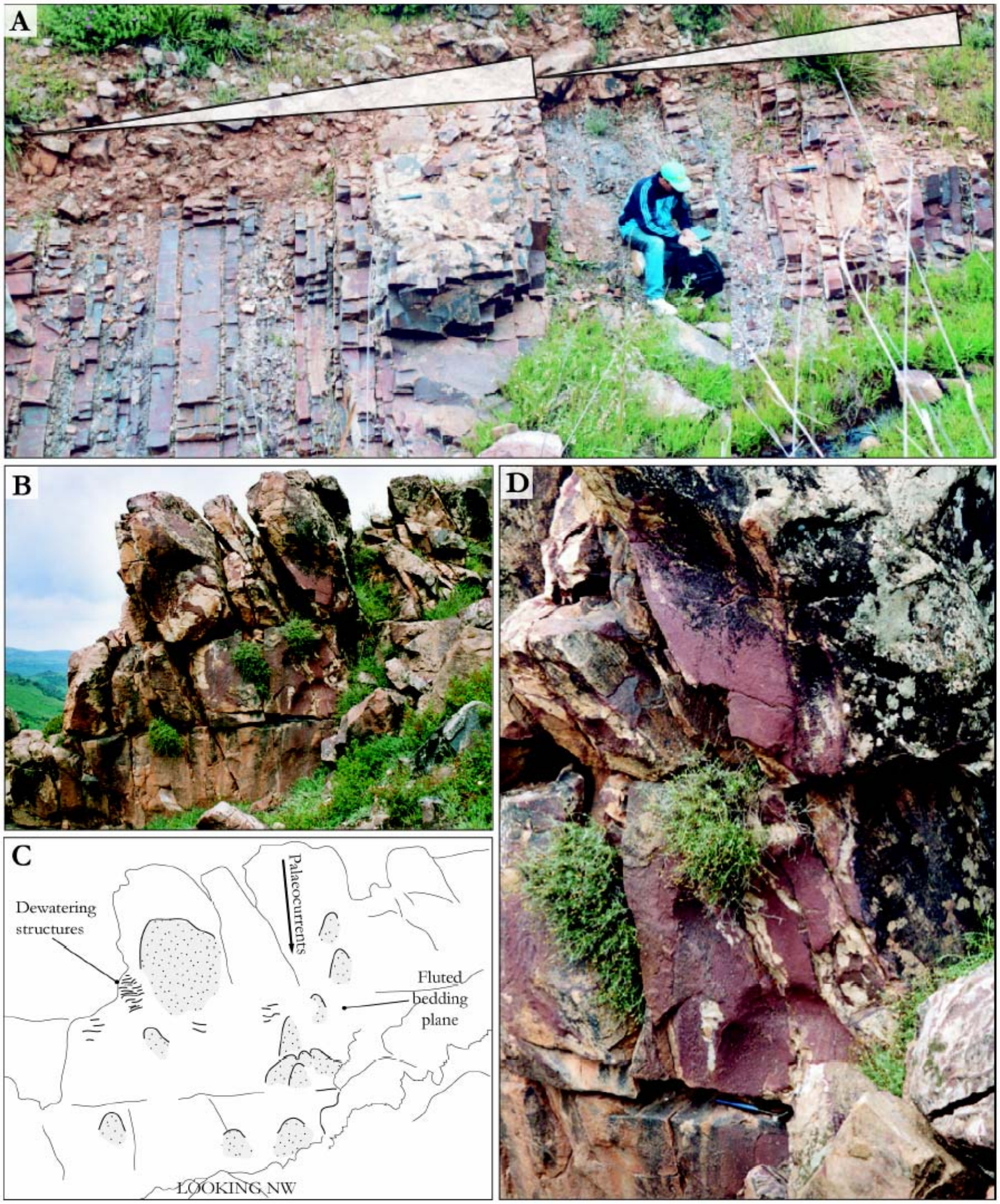

A
B
C
D
Dewatering structures
Palaeocurrents
Fluted bedding plane
LOOKING NW

**Figure 11**

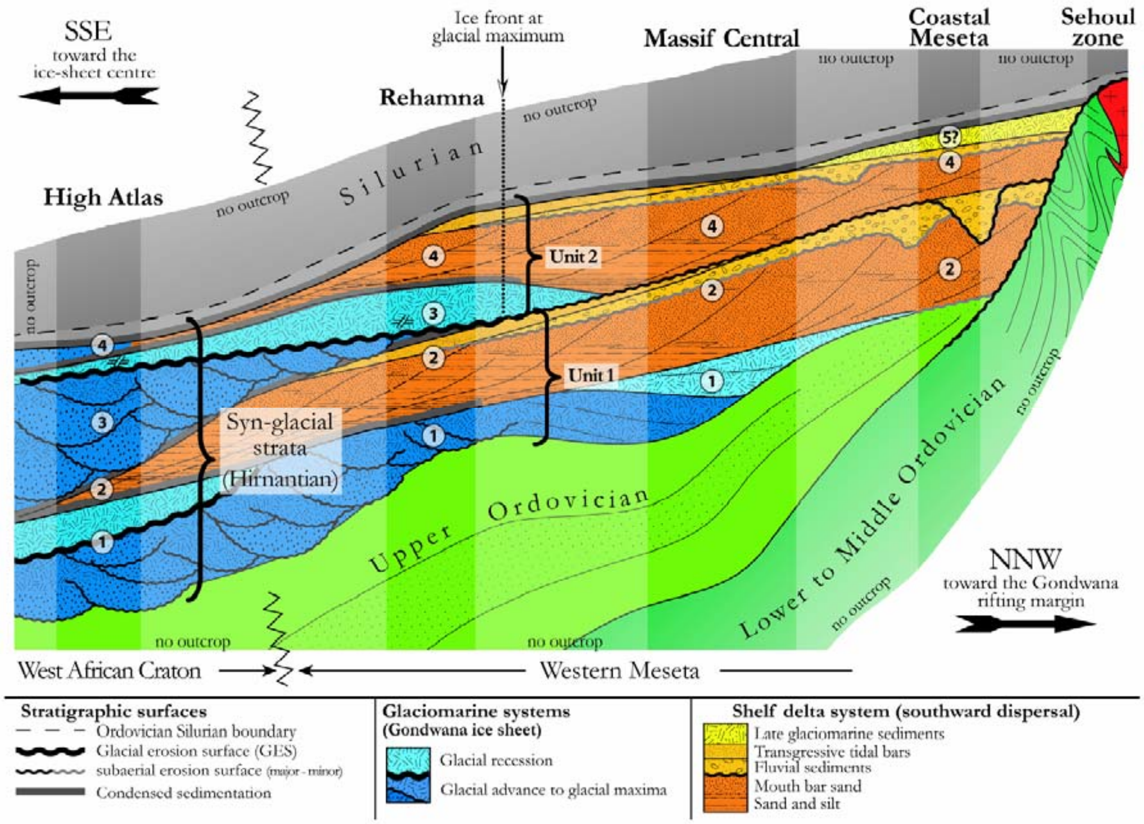

SSE
toward the ice-sheet centre
Ice front at glacial maximum
Massif Central
Coastal Meseta
Sehoul zone
Rehamna
High Atlas
Silurian
no outcrop
Unit 2
Unit 1
Syn-glacial strata (Hirnantian)
Upper Ordovician
Lower to Middle Ordovician
NNW
toward the Gondwana rifting margin
West African Craton
Western Meseta
Stratigraphic surfaces
Ordovician Silurian boundary
Glacial erosion surface (GES)
subaerial erosion surface (major - minor)
Condensed sedimentation
Glaciomarine systems (Gondwana ice sheet)
Glacial recession
Glacial advance to glacial maxima
Shelf delta system (southward dispersal)
Late glaciomarine sediments
Transgressive tidal bars
Fluvial sediments
Mouth bar sand
Sand and silt

**Figure 12**

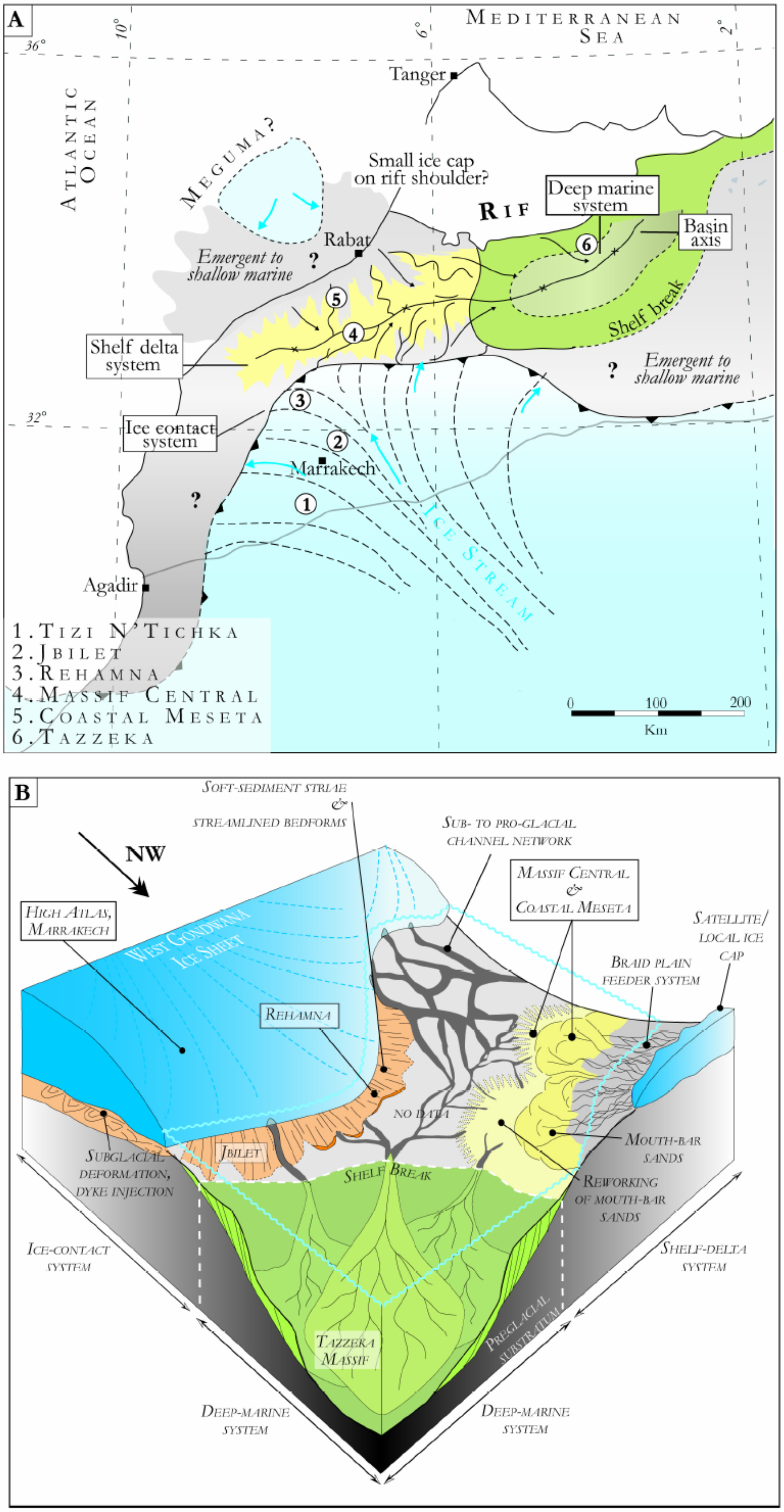

A
Mediterranean Sea
Atlantic Ocean
36°
32°
10°
6°
2°
Tanger
Meguma?
Small ice cap
on rift shoulder?
Rif
Deep marine
system
Basin
axis
Rabat
Emergent to
shallow marine
Shelf break
Shelf delta
system
Emergent to
shallow marine
Ice contact
system
Marrakech
Ice Stream
Agadir
1. Tizi N'Tichka
2. Jbilet
3. Rehamna
4. Massif Central
5. Coastal Meseta
6. Tazzeka
0
100
200
Km
B
Soft-sediment striae
&
streamlined bedforms
Sub- to pro-glacial
channel network
NW
Massif Central
&
Coastal Meseta
High Atlas,
Marrakech
West Gondwana
Ice Sheet
Satellite/
local ice
cap
Braid plain
feeder system
Rehamna
No data
Jbilet
Shelf Break
Subglacial
deformation,
dyke injection
Mouth-bar
sands
Reworking
of mouth-bar
sands
Ice-contact
system
Shelf-delta
system
Preglacial
substratum
Tazzeka
Massif
Deep-marine
system
Deep-marine
system

**Figure 13**