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Stratigraphy and sedimentology of the Sardona unit, Glarus Alps: Upper Cretaceous/middle Eocene deep-marine flysch sediments from the Ultrahelvetic realm

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Key words: Sardona unit, Ultrahelvetic, flysch, turbidites, Late Cretaceous, Eocene

ZUSAMMENFASSUNG

In der vorliegenden Arbeit wird eine revidierte Litho- und Chronostratigraphie für die Sardona-Einheit präsentiert. Ferner wird aufgrund der Sedimentologie das Ablagerungsmilieu interpretiert. Die Sardona-Einheit besteht aus ca. 200 m pelagischen und hemipelagischen Globotruncanenkalken und -mergeln von Cenomanian-Campanian Alter. Darüber folgen ca. 650 m Flysch von Maastrichtian-Bartonian Alter aus dem Ultrahelvetischen paläogeographischen Raum. Dieser Sardona Flysch wurde vermutlich in einem Becken mit eingeschränkter Zirkulation zwischen der Hochzonen des Südhelvetikums und des nördlichen Prättigau abgelagert. Er wird in drei Formationen unterteilt, von denen zwei vorwiegend aus kalksandigen Turbiditen bestehen. Sie werden Infraquarzit-Flysch und Supraquarzit-Flysch genannt. Getrennt werden diese durch die überwiegend siliziklastische Sardona-Quarzit-Formation, welche Paläozän bis Ypresian Alter hat. Diese Formation kann in verschiedene Faziesassoziationen unterteilt werden: Ölquarzit, sandiger Sardona Quarzit und konglomeratischer Sardona Quarzit. Zusammen bilden diese einen tiefmarinen Sedimentationsfächer, welcher während eines Meeresspiegel-Tiefstandes vom Norden her ins Becken progradierte. Der vierte Faziestyp besteht aus einem sehr grobkörnigen Kristallin-Konglomerat, welches aus intraformationellen und exotischen sedimentären und kristallinen Klasten zusammengesetzt ist. Dieser Sedimenttyp wird als Ablagerung von Schlamm- und Schuttströmen (debris-flow) gedeutet, welche vom Süden her ins Becken gelangten. Der Wechsel zurück zu Kalk-dominierten Turbiditen zu Beginn des Lutetian wurde durch erneute Transgression verursacht, welche ehemals freigelegte Flachwasserbereiche überflutete. Die Sedimentation innerhalb des Sardona-Flysch-Beckens wurde durch Überschiebungen vom Süden her beendet.

ABSTRACT

This paper presents a revised lithostratigraphy and chronostratigraphy for the Sardona unit, plus new sedimentological and environmental interpretations for its sediments. The Sardona unit consists of ~200 m of Cenomanian-Campanian pelagic and hemipelagic Globotruncana Limestone and Marl, plus ~650 m of Maastrichtian-Bartonian flysch deposits from the Ultrahelvetic paleogeographic realm. The Sardona Flysch was probably deposited in a structurally restricted, starved basin enclosed between the South Helvetic and North Prättigau Swells; it can be subdivided into three formations, two of which are predominantly composed of calcareous sandy turbidites, termed the Infraquartzite and Supraquartzite Flysch, that are separated by the predominantly siliceous Sardona Quartzite Formation of Paleocene-Ypresian age. The latter can be further subdivided into facies associations: the Ölquartzite, sandy Sardona Quartzite and pebbly Sardona Quartzite units, which together represent a highly efficient deep-marine fan system that prograded into the basin from its northern margin dur-

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ing a relative sea-level lowstand. The fourth facies association is the cobble- to boulder-grade Crystalline Conglomerate, which is composed of intraformational and exotic, sedimentary and crystalline clasts and has been interpreted as a series of cohesive debris flows derived from the southern basin margin. The switch back to calcareous turbidite sedimentation at the beginning of the Lutetian was caused by renewed transgression, drowning previously exposed shelf areas. Deposition within the Sardona flysch basin was terminated by overthrusting at its southern margin.

1. Introduction

The Sardona unit is located in the Glarus Alps of eastern Switzerland between the Sernf and Rhine Valleys, where it crops out in three main areas: Calfeisental, Weisstannental and the district of Elm within Sernftal (Fig. 1), extending over an area of 25 x 10 km, at elevations of 500 to 2,700 m. The Sardona unit has a WSW-ENE structural grain, is bounded by thrusts and forms part of the Infrahelvetic nappe pile that structurally underlies the Helvetic nappes (Pfiffner 1986; Lihou 1996). It is exposed in tectonic windows through the Helvetic nappes, from which it is separated by a major thrust, the Glarus Overthrust. The Sardona unit is structurally underlain by successively younger tectonic units, the Blattengrat and North Helvetic Flysch (NHF) units, respectively. A structural inversion has therefore been created, in which the oldest unit appears at the top of the nappe pile and the youngest at the base. Whereas the Sardona and Blattengrat units are entirely detached from their original basement (i.e., are allochthonous), the NHF unit is parautochthonous, being partially attached to the Aar massif in the south of the area.

The Sardona unit contains an apparently continuous record of deep-marine sedimentation from the Late Cretaceous to Early Tertiary. By contrast, Lower Tertiary shallow-marine and shelf deposits of the Blattengrat and NHF units were deposited on an unconformity (Herb 1988; Crampton 1992; Lihou 1995b), that has been interpreted as a 'forebulge unconformity', developed in an early, 'underfilled' North Alpine Foreland Basin (NAFB) (Allen et al. 1991; Crampton & Allen 1995; Lihou 1995b; Lihou & Allen in press). The structural inversion of the Infrahelvetic units may have been brought about by telescoping of the NAFB during northward translation of the Helvetic and Penninic nappes (Lihou 1996), such that the conformable sequence within the Sardona unit was also originally deposited in the NAFB, but closer to the advancing orogenic thrust wedge composed of the Penninic and Austroalpine nappes. Correlations between the Infrahelvetic units and the possible evolution of the NAFB have been discussed by Lihou (1995a) and Lihou and Allen (in press); this contribution presents the revised lithostratigraphy and chronostratigraphy for the Sardona unit that were necessary for such correlations, as well as providing new sedimentological and environmental interpretations for its sediments.

2. History of research

The Sardona unit was assigned to the Ultrahelvetic paleogeographic realm by Swiss workers (Leupold 1937, 1938, 1943; Bisig 1957; Rüefli 1959; Wegmann 1961). In contrast to other flysch units, it has not been studied since the 1960's because the sediments are extensively deformed and it was thought that little more could be learnt from them. A lithostratigraphy was erected between 1940 and 1960 and micropaleontological ages established for the various formations, but no attempt was made to interpret the results in

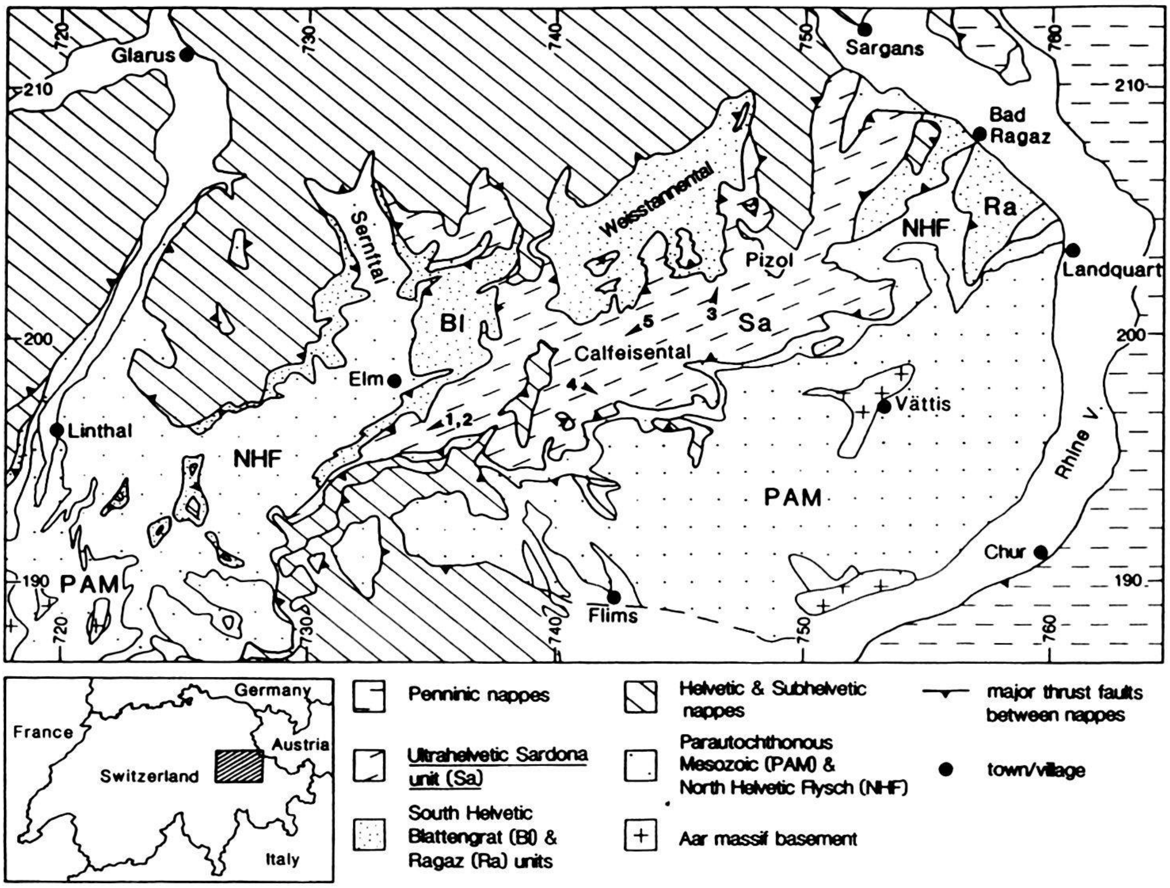


Fig. 1. Tectonic map of the Glarus Alps, adapted from Trümpy (1967) and Spicher (1980), showing the approximate localities of the type sections for the Sardona unit; marginal numbers refer to the Swiss national coordinate system and are marked at intervals of 10 km.
Type localities: Globotruncana Limestone and Marl: (1) Rufiberg (733.6/196.1); Infraquartzite Flysch: (2) Rindermättli (733.3/195.4), (3) Altsäss (746.0/202.2); Sardona Quartzite Formation: (4) Jakobiweid/Troseggtobel (741.8/198.0); Supraquartzite Flysch: (5) Plattenkopf (741.8/199.8).

terms of process-related sedimentology. In addition, the nature of the existing stratigraphic nomenclature for the Sardona unit was confusing, since different formation names were assigned by the earlier investigators to similar sediments in each of the three valleys where the Sardona unit is exposed. Therefore, a simplified stratigraphy which can be universally applied was adopted. In addition, the faunal ranges of the microfauna identified by Bisig (1957), Rüefli (1959) and Wegmann (1961) were revised in accordance with more modern treatises (Wagner 1964; Schaub 1981; Caron 1985; Harland et al. 1989). Consequently, the youngest flysch deposits of the Sardona unit probably reach Bartonian, not just Ypresian, age (Fig. 2).

3. Stratigraphy and sedimentology

Further sedimentological and petrographic details can be found in Lihou (1995a), a copy of which is available for reference at the University of Oxford.

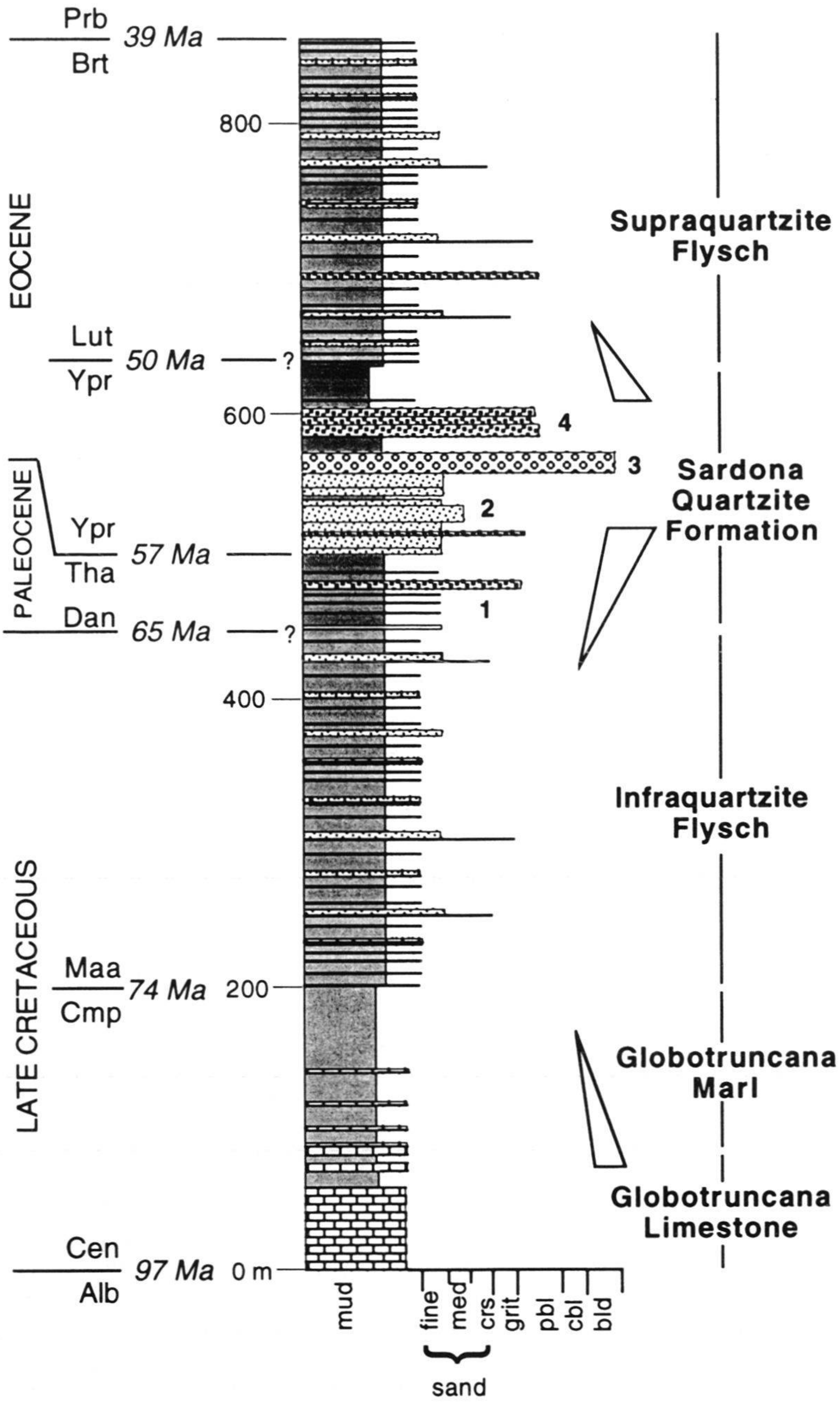


Fig. 2. Generalised stratigraphy of the Sardona unit, showing revised thickness estimates and micropaleontological dates; timescale according to Harland et al. (1989).
Facies associations within the Sardona Quartzite Formation: (1) Ölquartzite, (2) sandy Sardona Quartzite, (3) Crystalline Conglomerate and (4) pebbly Sardona Quartzite.

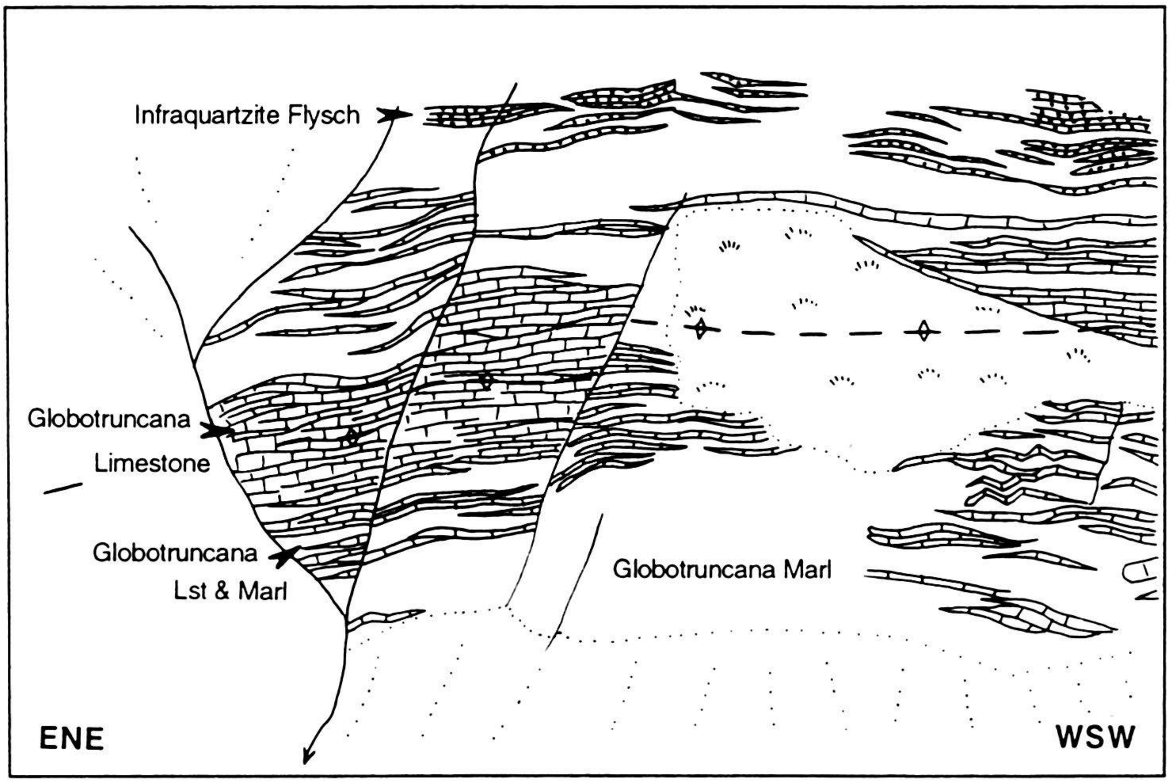


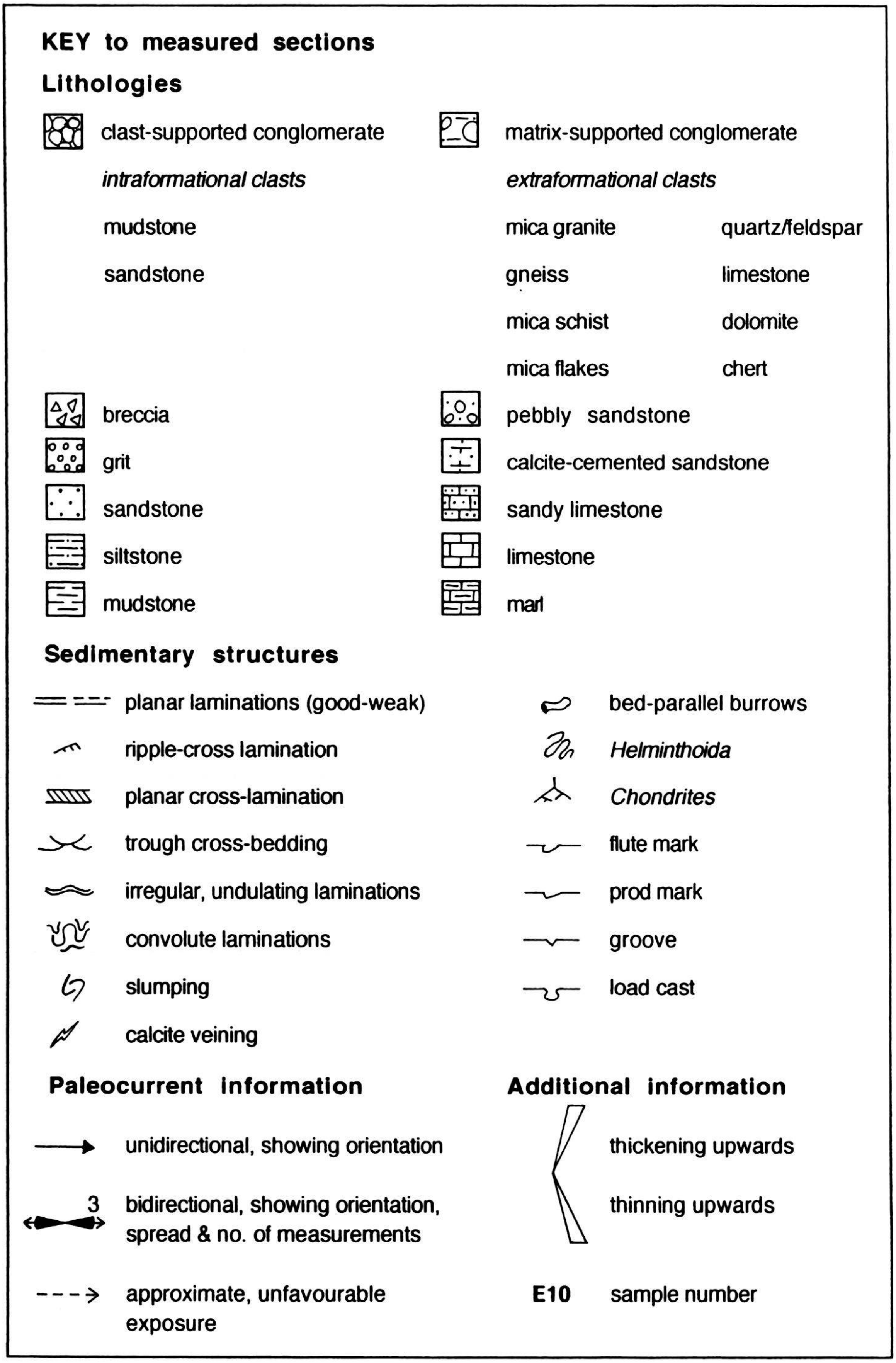
Fig. 3. Globotruncana Limestone and Marl exposed in the core of a recumbent anticline at Rufiberg (733.6/196.1) in the area south of Elm, with the transition to the Infraquartzite Flysch appearing in the upright limb at the top of the cliff; field of view ~200 m.

3.1 Globotruncana Limestone and Marl (Cenomanian-Campanian)

The oldest preserved deposit in the Sardona unit is the Globotruncana 'Seewer-like' Limestone, which is transitional up section into marl in complete exposures, usually found in the cores of anticlines (Fig. 3). The upward increase in marl content is associated with a thinning of individual limestone beds from an average of 15 to 5 cm (Fig. 4).

The majority of the foraminifera recovered from this unit are planktonic genera, namely *Globotruncana, Globigerina* and *Rotalipora* (Bisig 1957; Rüefli 1959; Wegmann 1961). The common occurrence of *Globotruncana lapparenti* together with the zonal fossils *Rotalipora montsalvensis* and *turonica/cushmani*, plus *Globotruncana helvetica, schneegansi* and *renzi*, indicate a probable age of Cenomanian-Campanian (Caron 1985; Harland et al. 1989). No older strata are preserved, as the base of the Sardona unit is bounded by a thrust. The total thickness of the limestone and marl is only about 200 m (Wegmann 1961), yet they represent ~23 Myr of sedimentation, so are likely to be a condensed sequence. I therefore interpret these thin-bedded carbonates as mainly pelagic limestones and hemipelagic marls deposited in a starved deep-marine environment.

The few fragmented, benthic foraminifera recovered from the Globotruncana Marl are typical of lagoonal and inner-shelf settings in modern environments (Murray 1991). Rüefli (1959) noted that influxes of fine quartz sometimes appeared to be connected with an abundance of sandy, agglutinating foraminifera, so the Cretaceous counterparts of the


KEY to measured sections
Lithologies
clast-supported conglomerate
matrix-supported conglomerate
intraformational clasts
extraformational clasts
mudstone
sandstone
mica granite
quartz/feldspar
gneiss
limestone
mica schist
dolomite
mica flakes
chert
breccia
pebbly sandstone
grit
calcite-cemented sandstone
sandstone
sandy limestone
siltstone
limestone
mudstone
marl
Sedimentary structures
planar laminations (good-weak)
ripple-cross lamination
planar cross-lamination
trough cross-bedding
irregular, undulating laminations
convolute laminations
slumping
calcite veining
bed-parallel burrows
Helminthoida
Chondrites
flute mark
prod mark
groove
load cast
Paleocurrent information
unidirectional, showing orientation
3
bidirectional, showing orientation, spread & no. of measurements
approximate, unfavourable exposure
Additional information
thickening upwards
thinning upwards
E10
sample number

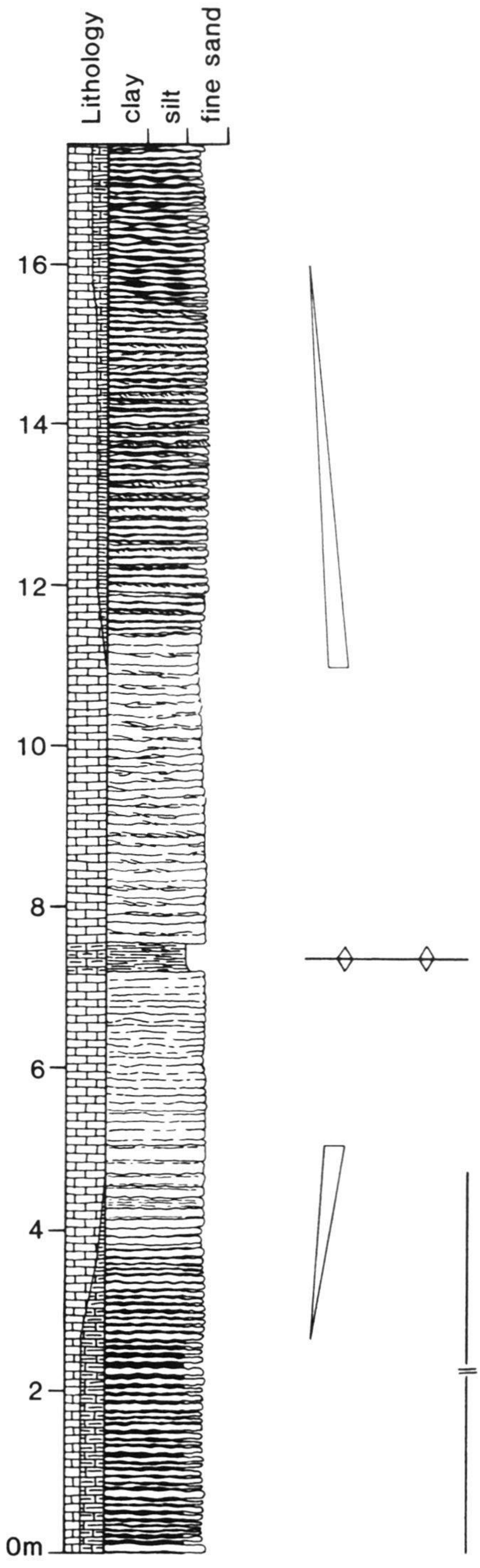


Fig. 4. Measured section through the Globotruncana Limestone and Marl at their type section at Rufiberg (733.6/196.1; 1560 m), as illustrated in Fig. 3.

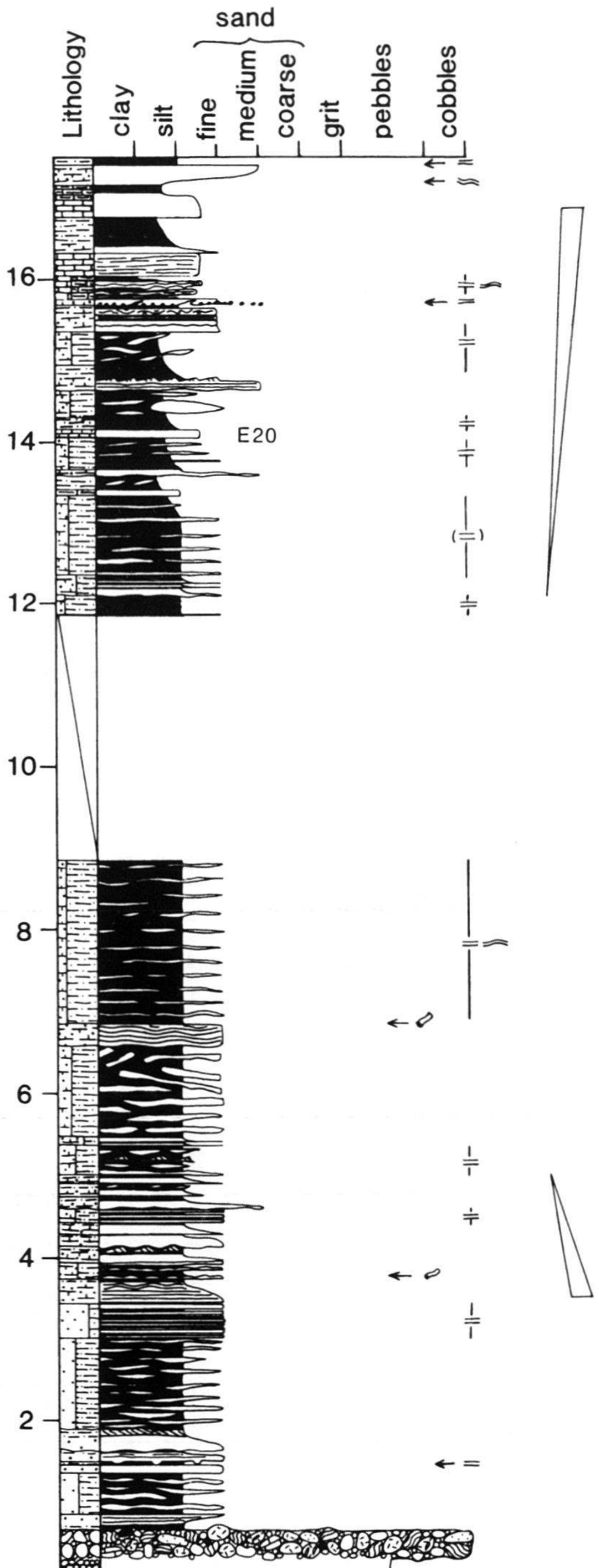


Fig. 5. Measured section through the 'Rindermättli Formation' of Wegmann (1961), which forms the basal part of the Infraquartzite Flysch Formation in the Elm district (733.3/195.4; 1750 m).

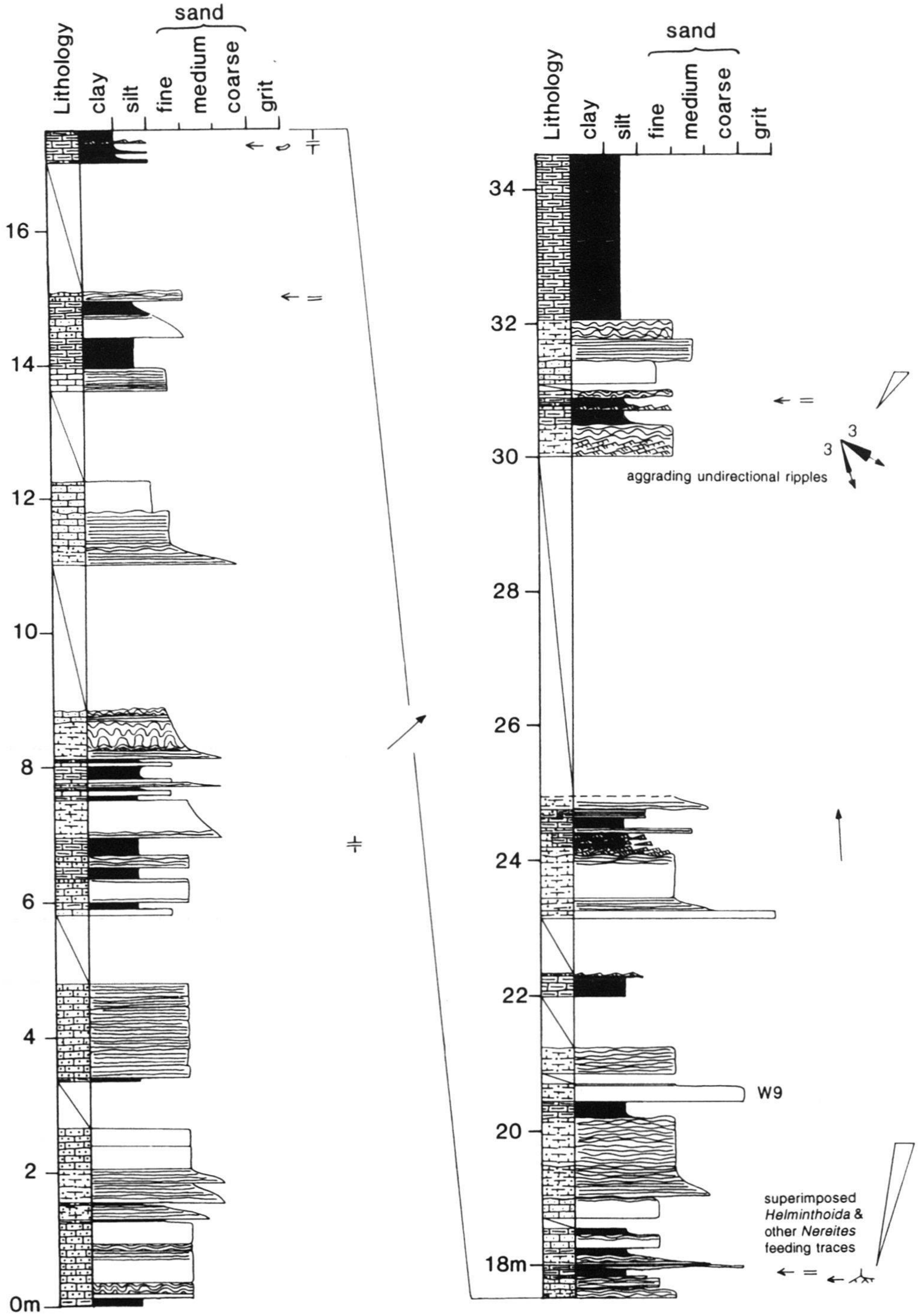
modern benthic communities may have been reworked into an outer-shelf environment by weak bottom currents.

3.2 Infraquartzite Flysch (Maastrichtian-?Paleocene)

The onset of clastic sedimentation across the basin was marked by the influx of graded, turbiditic calcarenites. The probability of reworking older fauna into a turbidite is extremely high (Homewood & Caron 1982), so it is most reliable to use the age of the youngest preserved microfauna, and in particular, the narrow age range of zonal fossils. On this basis, the timing of the change to clastic sedimentation in the Sardona basin is relatively well constrained by zonal *Globotruncana* species of Maastrichtian age, namely *Globotruncana leupoldi, stuarti, lugeoni* and *aegyptica* (Bisig 1957; Rüefli 1959; Wegmann 1961); these were recovered from the top of sandy flysch beds, where they probably accumulated during the waning of the turbidity currents. As with the older carbonate sediments, the microfauna in the Infraquartzite Flysch also includes a number of benthic foraminiferal genera, two of which, *Operculina* and *Amphistegina*, only appear in the Tertiary, therefore implying that the Infraquartzite Flysch continued to be deposited into the Paleocene. Its preserved thickness is 150–300 m, the uncertainty in this estimate arising from the intensity of the structural deformation.

Wegmann (1961) named the lowermost Infraquartzite Flysch that was transitional from the Globotruncana Marl the 'Rindermättli Formation', after his chosen type locality in the Elm district (733.3/195.4) (Fig. 1). Most of the beds at Rindermättli are lenticular to continuous, ungraded and graded, thin-bedded, micaceous fine sandstones and sandy limestones (facies D_3 of Mutti 1979), which are often broken, or disrupted by minor folding (Fig. 5). In places, there is clear evidence of soft-sediment deformation, e.g., convolute-laminated sandstones, or water-escape structures. Only 3% of this section consists of fine sandstone beds thicker than 10 cm (facies D_2 of Mutti 1979), and these thicker graded beds only rarely contain a thin carpet of medium to coarse sand. This suggests that either the parent flows had a very small volume and contained little material other than fine sand, or else that the basal Infraquartzite Flysch represents distal turbidites that had already deposited coarser sand in more proximal regions.

According to Rüefli (1959) and Wegmann (1961), both the bed thickness and the grain-size of the Infraquartzite Flysch increases southwards, implying that the source area was to the south. This suggestion could not be corroborated by direct measurement due to a lack of paleocurrent indicators and inconsistent results. The unit also apparently coarsens-upwards, at the same time becoming more micaceous and quartzose, suggesting that siliciclastic input increased over time either due to external changes in the source area, or to increasing relative proximity to the source region, or to a lateral facies change. Much thicker and coarser sandy turbidites can be found at Altsäss (746.0/202.2) (Fig. 6) and Sässplanggen (745.7/201.3) south of Weisstannen village, for example (Fig. 1). Here, Infraquartzite Flysch beds, or the 'brecciose sandy flysch' of Rüefli (1959), sometimes reach more than 50 cm thickness (facies C_2 of Mutti 1979) and 55% of the section consists of sandstones thicker than 10 cm (facies D_2 of Mutti 1979). Amalgamation of sandstone beds is commonplace (Fig. 6). The repeated appearance of coarse material or thicker turbidites in the Infraquartzite Flysch is not cyclic, suggesting that random, catastrophic events were responsible for their generation.


sand
Lithology
clay
silt
fine
medium
coarse
grit
16
14
12
10
8
6
4
2
0m
34
32
30
28
26
24
22
20
18m
3
3
aggrading undirectional ripples
W9
superimposed
Helminthoida &
other Nereites
feeding traces

The top of mudstone laminae are locally burrowed, indicating that more or less continuous burrowing by soft-bodied animals was punctuated by brief interludes of turbidite deposition. Frequent *Chondrites* and occasional *Helminthoida* traces are found within fine sandstone or mudstone facies (T_{de} interval of the Bouma sequence). These trace fossils form part of the *Nereites* ichnofacies of Seilacher (1967), which he associated with 'deep-water' pelagic and turbiditic sediments (i.e., bathyal-abyssal zone, from 600 m to more than 2000 m water depth). However, *Chondrites* and *Helminthoida* appear to be facies-independent genera (Crimes et al. 1981), so their presence is not diagnostic of any particular 'deep-water' depositional environment.

3.3 Sardona Quartzite Formation (?Paleocene-Ypresian)

At some time in the Paleocene, there was a change in composition of the turbidites entering the basin, from calcareous to siliceous and a predominantly arenitic unit, the Sardona Quartzite Formation, was deposited. Its age was relatively poorly constrained by earlier workers, due to a paucity of preserved microfauna. A few Tertiary benthic foraminifera were found in the upper part of the Sardona Quartzite Formation (Rüefli 1959), whose ages all lie within the Paleocene to mid Eocene. Improvements with the dating were attempted by Shell U.K. Exploration and Production, but no original microfauna were found owing to alteration of the calcite tests (pers. comm.). However, a few dissolution-resistant, diagnostic, *in-situ* palynomorphs were recovered from one quartzite sample: *Areoligera coronata, Areoligera medusettiformis* and *Areoligera senonensis*, which suggested a minimum age of middle Ypresian (late NP12/early NP13). Future attempts to improve the dating of the Sardona unit as a whole, by isolating dissolution-resistant palynomorphs, would be highly desirable.

The Sardona Quartzite Formation can be subdivided into four facies associations (Fig. 2, 7): (1) the thin-bedded 'Ölquartzite' unit; (2) the thicker-bedded, sheet-like, 'sandy Sardona Quartzite' unit, composed of graded and ungraded, massive quartzites, sometimes with thin, heterolithic interbeds; (3) the 'pebbly Sardona Quartzite' unit, consisting of stacked, poorly-sorted, quartzose pebbly sandstones and conglomerates; and (4) an unsorted, pebble to boulder conglomerate called the 'Crystalline Conglomerate' unit, that usually appears between the sandy and pebbly Sardona Quartzite units. These four subunits of the Sardona Quartzite Formation are spatially variable but can be traced regionally (Fig. 7), although individual flows cannot usually be correlated over distances of more than 1 km. I propose that the type section for the Sardona Quartzite Formation should be on Alp Sardona, in Calfeisental, between Jakobiweid (742.1/198.1) and Troseggtobel (741.8/198.0) (Fig. 1) because this is the most continuous conformable sequence available (being 155 m thick) (Pl. 1, 2).

The **Ölquartzite** always occurs at the base of the Sardona Quartzite Formation, but a similar facies association can also appear between the other units of this formation

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Fig. 6. Measured section through the 'brecciose sandy flysch' of Rüefli (1959) at Altsäss (746.0/202.2; 1860 m), which represents the type section for the Infraquartzite Flysch Formation in Weisstannental.

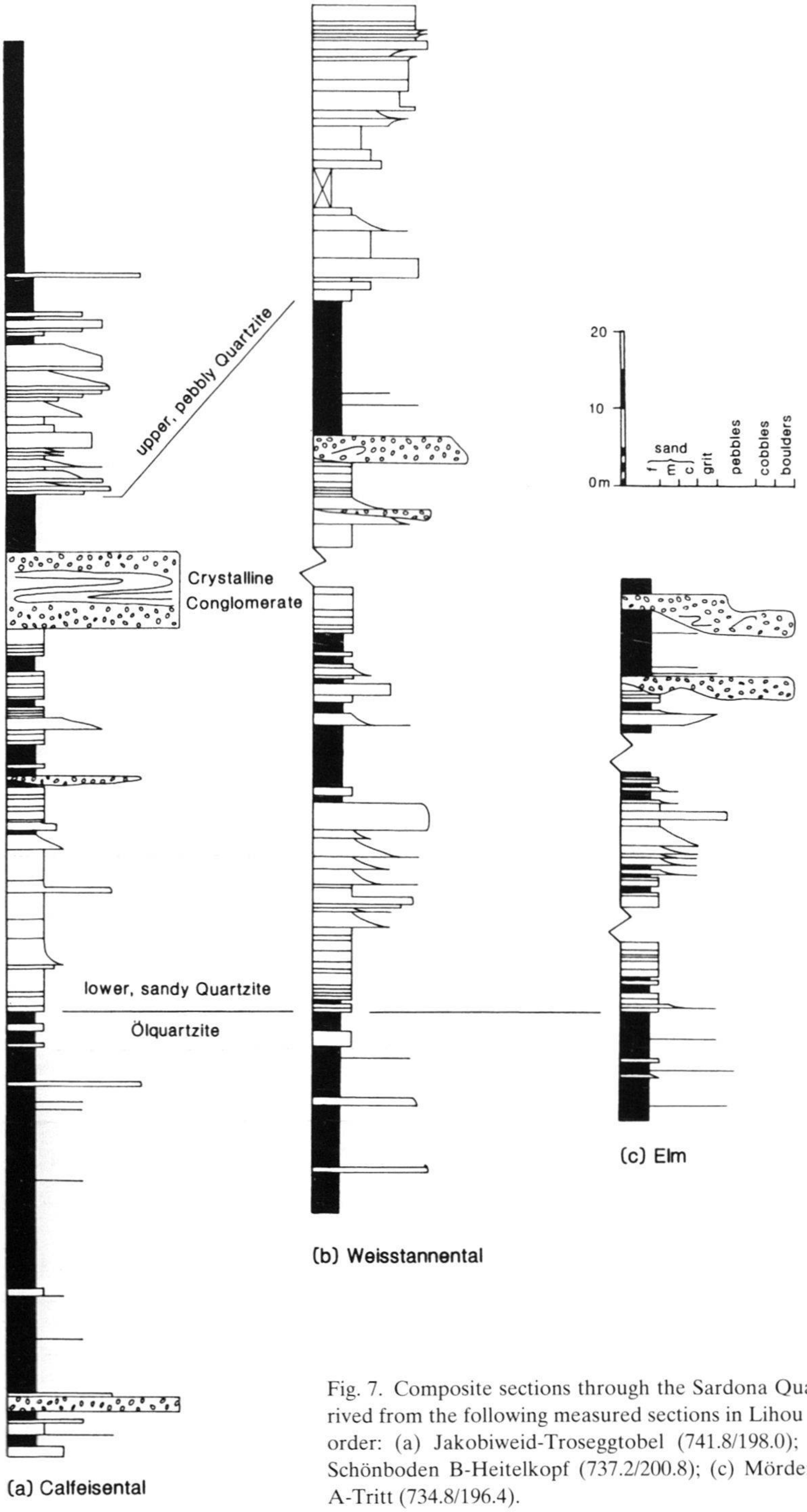

Fig. 7. Composite sections through the Sardona Quartzite Formation derived from the following measured sections in Lihou (1995a), in ascending order: (a) Jakobiweid-Troseggtobel (741.8/198.0); (b) Schönboden A-Schönboden B-Heitelkopf (737.2/200.8); (c) Mörderhorn B-Mörderhorn A-Tritt (734.8/196.4).

Fig. 8. Appearance of the Ölquartzite at its type locality of Jakobiweid, Alp Sardona, Calfeisental (742.1/198.1), illustrating the periodic appearance of thicker sandy turbidites within the section, which is not associated with cyclic thickening and thinning.

(Fig. 7). At its type locality of Jakobiweid, it reaches a maximum preserved thickness of 62 m (Pl. 1). Its brown-grey to rusty colour on weathered sandstone surfaces makes it readily distinguishable from the yellow-grey-weathering Infraquartzite and Supraquartzite Flysch. Interbedded silts and mudstones are light to medium grey and black, respectively, the colour probably reflecting their primary organic content.

Bioturbation is commonplace; thin fine sandstones are locally thoroughly bioturbated, so that their primary sedimentary structures are completely destroyed. The ichnogenera found most commonly in the Ölquartzite, and in positive relief on the base of other thin sandstone beds within the Sardona Quartzite Formation, are *Planolites* and *Palaeophycus*. Crimes (1976) considered *Planolites* to be a facies-independent 'deep-water' genus. Although the ichnogenera in these siliceous turbidites are different from those in the underlying calcareous turbidites, the depocentre probably remained at bathyal-abyssal depths and a change in environmental factors may have been partially responsible for reforming the ichnofauna.

In general, the Ölquartzite consists of thin-bedded, fine sandy turbidites (facies D of Mutti 1979), thicker-bedded fine to coarse, micaceous sandy turbidites or ungraded sandstones (facies C of Mutti 1979), and rare, non-channellised, clast-supported, medium-coarse pebbly conglomerates (facies R of Lowe 1982) (Pl. 1). In broad terms, the

Ölquartzite exhibits a series of coarsening-upwards cycles, but the periodic appearance of coarse material or thicker turbidites (>10 cm thickness) in the section is not associated with cycles of thickening and thinning (Fig. 8). The thicker and coarser turbidites may have originated as larger-volume flows, produced by gravelly and sandy high-density turbidity currents (HDTC) (Lowe 1982), rather than the low-density turbidity currents (LDTC) responsible for the more frequent, thinner, graded sandstones. In order for the coarse material to be deposited beyond the shelf-slope break in the deeper parts of the Sardona basin, it would have had to be fed into a steep canyon, in order to sustain the flow thickness required to maintain the flow.

Sole marks from Ölquartzite beds in the Elm district and Weisstannental suggest that the sand flowed across the basin principally from the northeast (Fig. 9a), while the sand in Calfeisental apparently came from the southeast and east (Fig. 9b). Ripple directions appear to be parallel to the modes in the corresponding sole marks. However, the data from Calfeisental include a significant number of ripples that are directed towards the southeast, in the opposite direction to the regional sole marks (Fig. 9b).

Obliquely bimodal ripple directions were occasionally obtained from a single bed in Calfeisental, plotting in the west and southeast quadrants, but ripples from most beds fell into one or other of these quadrants (Fig. 9b). In beds where both ripple directions were present, the southeast quadrant ripples were parallel to their basal grooves, and therefore probably represent the primary flow direction, i.e., towards the southeast.

In some of the thicker Ölquartzite sandstones, three-dimensional ripples exhibiting stoss-side preservation during predominantly aggradational sedimentation may have given rise to an obliquely bimodal ripple-lamination pattern. Most thin Ölquartzite turbidites did not accumulate by predominantly aggradational sedimentation however. Reflection of the primary turbidites at the basin margins or some intrabasinal high can create tractional bedforms with an opposing or oblique paleocurrent direction (Pickering & Hiscott 1985; Kneller et al. 1991; Sinclair 1994). The major problems with this hypothesis for the Ölquartzite paleocurrent pattern is that the beds are thin, there is no discrepancy between the orientation of the sole mark and ripple modes, and the ripple directions themselves are bimodal (Fig. 9b).

On the other hand, longitudinal bottom currents, or contourites, flowing westward, may have been responsible for redistributing sandy material originally deposited by southeasterly- or southwesterly-directed turbidity currents. Acceleration of modern, contour-parallel bottom currents occurs in narrow passages where flow is constricted, so that flow velocities commonly reach 10–20 cm/s, which is sufficient to erode, transport and deposit fine sediment (Stow & Lovell 1979). One of the vital criteria in the recognition con-

Fig. 9. Rose diagrams showing the compilation of all paleocurrents measured in the Sardona unit, grouped according to their geographical location and/or their stratigraphic position, with sole marks shown separately from internal primary sedimentary structures.
Bipolar paleocurrent indicators (grooves, channels and parting current lineations) are plotted as bipolar roses, whereas unimodal paleocurrents are plotted as single roses; the length of each rose is therefore the sum of both the bipolar and unimodal measurements. Corrections for bed tilt alone were made because fold plunges were found to be shallow enough to have little effect upon the structural correction.

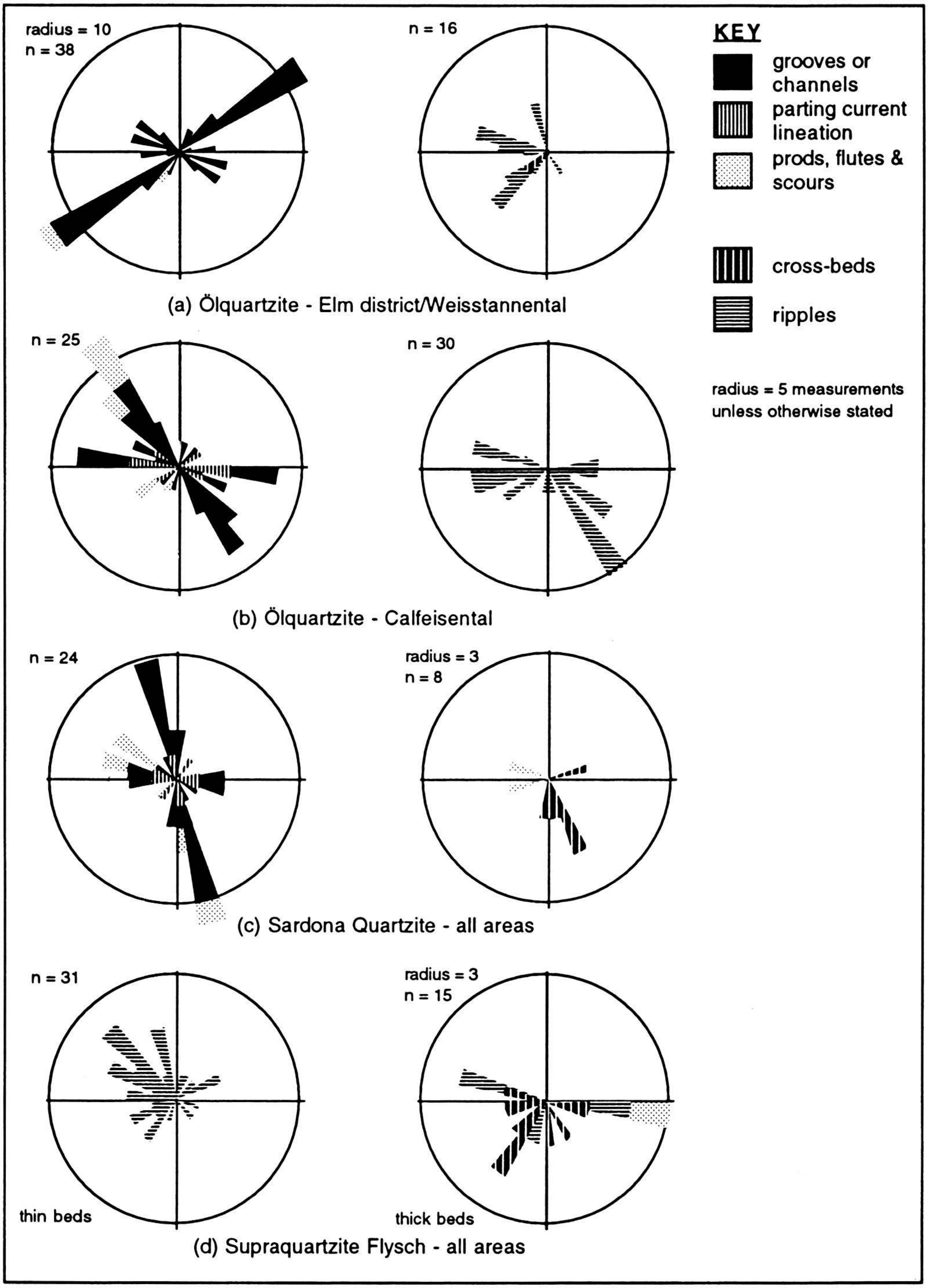
radius = 10
n = 38
n = 16
KEY
grooves or channels
parting current lineation
prods, flutes & scours
cross-beds
ripples
radius = 5 measurements unless otherwise stated
(a) Ölquartzite - Elm district/Weisstannental
n = 25
n = 30
(b) Ölquartzite - Calfeisental
n = 24
radius = 3
n = 8
(c) Sardona Quartzite - all areas
n = 31
radius = 3
n = 15
thin beds
thick beds
(d) Supraquartzite Flysch - all areas

Fig. 10. Example of the sharp interface between the thin-bedded Ölquartzite and the overlying thick-bedded sandy Sardona Quartzite at Jakobiweid, Alp Sardona, Calfeisental (742.1/198.1); height of cliff ~30 m.

tourites is paleocurrent directions that show bimodality at approximately 90°, with limited scatter (Stow & Lovell 1979). In the case of the Ölquartzite, the modes are moderately well clustered, differing by 60° and 110° in the two areas (Fig. 9a, b). There are additional, sedimentary characteristics of the Ölquartzite that are indicative of contourites: the sandstones are commonly thin (<3 cm thick), composed of well-sorted silt or fine sand, have parallel- or ripple-laminations, sharp upper and lower boundaries and are interbedded with black mudstone. Thin, contemporaneous Ölquartzite beds in the Tonstein Formation of the Schlieren Flysch, Central Switzerland, have also been interpreted as the product of reworking by contour currents, but in the Schlieren Flysch, the principal basin-parallel currents were eastward-directed (Winkler 1983).

There is an abrupt transition from the Ölquartzite into **the sandy Sardona Quartzite** unit (Fig. 3, 10), dated at the beginning of the Eocene, based upon the new palynological data. Individual sandstone beds are much thicker than in the Ölquartzite, are often amalgamated and there are fewer graded beds. The Sardona Quartzite is a dense, uniform light grey to white, micaceous quartzite, that weathers to a mottled brown-orange colour due to accessory amounts of hematite (Rüefli 1959; Wegmann 1961). It typically forms cliffs tens of metres high covered in a characteristic green lichen (Leupold 1943), thus forming a reliable marker horizon. At its type locality of Troseggtobel, it reaches a maximum preserved thickness of 48 m (Pl. 2).

The sandy Sardona Quartzite is composed of a series of ungraded or graded, massive, mostly fine sandstones, that are sometimes parallel-laminated at the top of the bed (facies C_1 of Mutti 1979; facies S of Lowe 1982), of which the ungraded sandstones are the most common. Thickening and thinning of the quartzite beds appears to be cyclic, some beds reaching several metres thickness (up to a maximum of 5 m), but there are no obvious coarsening- or fining-upwards cycles (Pl. 2). Thin, heterolithic interbeds locally appear between packages of thick sandstones. Amalgamation of the thicker sandstones, caused by removal of the intervening mudstones, is commonplace and provides evidence of the erosive nature of the currents responsible for their deposition.

The large number of graded, massive sandstone beds could be associated with flow stripping of channelised debris flows during spillover; if sandy HDTC travelling beyond their channellised debris flows reached unconfined regions of the basin and were then deposited rapidly, their high suspended sediment concentrations might dampen any turbulence required to give rise to stratified bedforms (Lowe 1982, 1988). It is possible that the lack of sedimentary structures in the ungraded, massive quartzites was also related to high sediment concentrations and sedimentation rates, if they were the product of large-volume turbidites with little or no variation in grain-size, so that grading could not develop.

Another reason for the occurrence of massive, metre-thick sandstones might be fluid escape following deposition, but no characteristic pillar or dish structures were observed and this probably cannot explain the occurrence of stacked, amalgamated massive beds that are a characteristic of the unit. A third method of producing massive sandstones is by intense, post-depositional bioturbation, but the few ichnogenera preserved in the sandy Sardona Quartzite were sediment grazers rather than infaunal deposit feeders likely to thoroughly bioturbate sandstones up to several metres thick. Hence, the only feasible explanation for the development of these massive sandstone beds is high suspended sediment concentrations and temporarily high sedimentation rates.

A similar facies association to the sandy Sardona Quartzite, called the **pebbly Sardona Quartzite**, forms the uppermost 20 m of the Sardona Quartzite Formation at its type section at Troseggtobel (Pl. 2). It has a limited outcrop and was only observed elsewhere at Heitelkopf in Weisstannental (738.3/200.7) (Fig. 7b) and at Heubützli Pass (741.6/200.4) in Calfeisental; Rüefli (1959) only found this facies association in the southern part of Weisstannental. It consists of a series of poorly-sorted, relatively immature, medium pebbly sandstones and conglomerates that are usually light grey when fresh, and brown when weathered. The sandstones are commonly ungraded or normal-graded and massive. Repeated cycles of thickening- and thinning-upwards can be recognised, within an overall thickening-upwards trend. These cycles do not coincide with fining- or coarsening-upwards trends, although such trends are present (Pl. 2). At Heitelkopf, there is also an upwards trend in the graded pebbly sandstones, from normal to reverse grading, with rare low-angle cross-bedding. The base of pebbly beds are occasionally erosive; scours filled with pebbly conglomerates can be traced over several metres, reaching depths of 100–120 cm.

A lack of grading in these pebbly sandstones is interpreted to have arisen from rapid sedimentation, during frictional freezing of cohesionless debris flows. Flow competence of cohesionless debris flows is dependent on the flow thickness (Nemec & Steel 1984) and channelling helps to maintain this; the depth of the scours may have influenced the

Fig. 11. Detail of the Crystalline Conglomerate, illustrating the lack of sorting of crystalline and sedimentary material enclosed within the black siltstone matrix. To the top right of the hammer is an unconsolidated, clast-supported pebbly conglomerate that has been reworked into the debris flow.

narrow range of bed thicknesses (10–70 cm) shown by the pebbly Sardona Quartzite. Pebbles would have been maintained in suspension owing to the dispersive pressure created by grain interactions, which can give rise to reverse grading. Minor turbulence caused by the incorporation of some water in more distal areas causes a normal-graded bed to be deposited. Hence, the trend from normal- to reverse-graded pebbly sandstones found at Heitelkopf may reflect increasing proximity to the source region over time. The fact that these sediments contain mechanically relatively unstable mica schist, gneiss, granitoid and feldspar fragments as well as quartz and chert, indirectly indicates that they were deposited relatively close to their source. Exceptional, isolated coal fragments, which were unlikely to survive significant transport distances, were found towards the base of a normal-graded pebbly sandstone at Heubützli Pass.

Far fewer paleocurrent indicators are preserved in the Sardona Quartzite than in the Ölquartzite (Fig. 9c). Contributing factors may have been the depositional mechanism for the thick-bedded, sandy Sardona Quartzite, which suppressed the development of internal sedimentary structures; the amalgamation of beds that arose from erosion of the upper part of the underlying bed, where ripple-lamination may have developed; and the uniform grain-size of the sandy quartzites, which was unlikely to encourage the formation of tool marks on the base of beds. The relatively poor exposure of cross-stratification

Table 1. Composition and provenance of clasts found in the Crystalline Conglomerate, compiled from Rüefli (1959), Wegmann (1961) and using new data collected in this study; clast compositions highlighted in bold letters are those that occur most commonly.

SEDIMENTARY INTRACLASTS	**quartzite**	**ölquartzite**
	black mudstone	*Globotruncana* lst
	clast-supported, polygenic cobble cgl	
UPPER CRETACEOUS	*Orbulinaria* lst	
LWR CRETACEOUS/UP JURASSIC	*Calpionella alpina* limestone with radiolaria	
LOWER & MIDDLE JURASSIC	oolitic limestone with *Cosconiconus alpinus*	
	siliceous limestone with sponge spicules	
	shale	
TRIASSIC	limestone	dolomite
VERRUCANO METASEDIMENTS	quartzite	
VERRUCANO METAVOLCANICS	diabase	spillite
	granophyre	andesite
PLUTONIC	**porphyritic white granite or granodiorite**	
	pink-red granite	aplite
	monzonite	**diorite**
	tonalite	**microgranite**
METAMORPHIC	**granitic gneiss**	
	augen or mylonitic gneiss	
	chloritic & other mica schists	
	vein quartz	dolomitic marble

within the pebbly Sardona Quartzite is probably the main reason for so few paleocurrents being identified in this lithology. Nevertheless, there is evidently a different paleocurrent pattern for the Sardona Quartzite than the Ölquartzite: sole marks from the Sardona Quartzite in all areas are principally oriented north-south. The few data available suggest that cross-bedded, pebbly and sandy quartzites were derived from the north (Fig. 9c). Nevertheless, the degree of scouring, the reverse grading and the presence of coal fragments in the more northern exposures of the pebbly Sardona Quartzite suggests that they are more proximal. Hence, it is tentatively concluded that the main source of the Sardona Quartzite was to the north of the basin.

An unsorted, matrix-supported, pebble to boulder conglomerate called the **Crystalline Conglomerate** (facies A_2 of Mutti 1979) usually appears above the sandy Sardona

Fig. 12. Typical appearance of the Crystalline Conglomerate, here seen at Schafälpli, Alp Sardona, Calfeisental (740.0/198.3); a brown-weathering channellised conglomerate, indicated by the arrow, underlies the Crystalline Conglomerate at the base of the cliff.

Quartzite unit (Fig. 7, 11). It does not represent a single event, either temporally or spatially confined, because it appears more than once in some measured sections through the Sardona Quartzite Formation (Fig. 7c) and in others not at all (e.g., at Heubützli Pass (741.6/200.4)). At its type section of Troseggtobel, it reaches a thickness of 10 m (Pl. 2). It was formerly placed stratigraphically below the Sardona Quartzite (Leupold 1943; Bisig 1957; Rüefli 1959; Wegmann 1961), but this is erroneous.

The Crystalline Conglomerate is a distinctive deposit because of its chaotic nature and the striking size of the largest, outsized clasts or megaclasts, several examples exceeding 1 m^3. However, most of the clasts are either small pebbles or cobbles (Fig. 11), consisting of well-rounded, exotic sedimentary and crystalline material, plus sedimentary intraclasts (Tab. 1). The intraclasts comprise individual beds, slumped blocks, rafts, rounded cobbles, boulders and mudstone rip-ups. Clasts can be found in concentrations of 5 to 30% by area, floating in a black siltstone matrix (Fig. 12). In general, the conglomerate is ungraded and unsorted. Lateral variability in the size, composition and concentration of clasts can be extreme: decimetre-thick lenses of small, angular pebbles may pass laterally over 10 m into a chaotic conglomerate several metres thick containing rounded boulders.

Contacts with the underlying beds may be either concordant or shallowly channelled (Fig. 7), but lateral changes in thickness owing to tectonic disruption may enhance the appearance of discordance in some cases. Scouring appears to be associated with rare examples of crude, normal grading of the conglomerate (facies A_1 of Mutti 1979), where there is an upward decrease in the size of the clasts brought about by slight dilution of the flow and large-scale turbulence.

Several independent observations led earlier workers to suggest that the Crystalline Conglomerate had a southerly source: Leupold (1943) discovered that the largest clasts were located in Calfeisental; Bisig (1957) found that the Crystalline Conglomerate was essentially absent north of Elm; Rüefli (1959) observed a dramatic southward increase in clast-size in Weisstannental; and Wegmann (1961) claimed that the proportion and size of the clasts south of Elm decreased towards the WSW. In addition, the Crystalline Conglomerate is generally thicker, appears more frequently (i.e., there is a greater volume of such material) and is more often scoured in Calfeisental, implying that this area was more proximal to its source region.

I interpret the Crystalline Conglomerate as the product of cohesive debris flows. Fine silt and clay mixed with < 50% water by volume, has sufficient cohesive strength to transport material up to pebble-grade (Rodine & Johnson 1976). The large boulders incorporated into the Crystalline Conglomerate would have been supported by buoyancy forces equivalent to the weight of the displaced volume of matrix, i.e., by 'Archimedes' Principle (Rodine & Johnson 1976). These boulders may also have been partially supported by excess pore-fluid pressure generated by their weight, which was slow to dissipate because of the poor sorting of the matrix (Pierson 1981).

Most megaclasts within the Crystalline Conglomerate were probably transported within a rigid plug, sliding or gliding along parallel to the base of the flow and supported by a cushion of over-pressurised matrix material. However, shearing at the base of some megaclasts causing fragmentation of neighbouring, smaller clasts and 'flow bands' of pebbles around the megaclasts may have been caused by internal shear or partial flow turbulence. Slump blocks, rafts and megaclasts of intraformational material in the Crystalline Conglomerate demonstrate that the debris flows had some erosive capacity since they were able to incorporate slabs of material from the surface over which they were flowing.

Scours at the base of the conglomerate need not have been created by the debris flows themselves, but could have been pre-existing morphological features that were exploited by them. Evidence that some of the debris flows initially followed the course of pre-existing channels is provided by a laterally impersistent clast-supported cobble conglomerate that locally underlies the debris flow conglomerate (Fig. 12). These clast-supported conglomerates probably represent channellised cohesionless debris flows which were often removed by the debris flow. The debris flows probably overspilled and overran these channels, passing into unconfined regions of the basin.

The presence of metre-sized boulders of exotic material within the Crystalline Conglomerate signifies a direct link between the source region and the depocentre on the basin slope and floor. This necessitates steep subaerial and submarine slopes to maintain the velocity required to carry such material, so they were probably introduced into the basin via steep submarine channels or canyons. Although debris flows usually initiate on steeper slopes, a recent submarine debris flow in British Columbia flowed 5 km across the bottom of a fjord which had an average slope of only 0.4° (Prior et al. 1984), therefore steep slopes need not be invoked for their transportation. Consequently, the Crystalline Conglomerate could have been deposited on the basin floor up to a few kilometres from the base of the slope.

The main part of the British Columbian flow removed and incorporated 4–5 m of muddy Holocene fjord-bottom sediments, which accounted for half of the total volume of the debris flow, the remainder coming from the delta-front where the flow initiated (Prior et al. 1984). The muddy matrix of the Crystalline Conglomerate may likewise have been derived from erosion and incorporation of a significant amount of unconsolidated material from a more proximal, muddy area. However, the coarse material in the Crystalline Conglomerate mainly came from extraformational sources and is well-rounded, suggesting that it had been mechanically abraded. These boulders were more likely to have been abraded by buffeting from smaller clasts in a fluvial or beach setting, prior to being introduced into the basin, than during transport within the debris flow.

3.4 Supraquartzite Flysch (?Lutetian-Bartonian)

The youngest preserved sediment in the Sardona unit, the Supraquartzite Flysch (Fig. 2), contains many nummulitids and assilinids of early to mid Eocene age (Rüefli 1959; Wegmann 1961; Schaub 1981), as well as the late Bartonian (P14) zonal fossil *Globorotalia spinulosa* (Caron 1985; Harland et al. 1989). Hence, its time range could span the Ypresian to Bartonian, but the Ypresian benthic foraminifera are likely to have been reworked into the turbiditic sandstones. Attempts to improve the dating of this unit again failed because the original planktonic foraminifera and nannofossils had been dissolved (Shell UK Exploration & Production, pers. comm.).

Leupold (1943) originally described the Supraquartzite Flysch as an ochre-yellow-weathering, black, velvety slate, full of 'fucoids' (*Chondrites*) and *Helminthoida*, with 0.5–2 m thick beds of micaceous, medium-grained, arkosic calcareous sandstones forming yellow bands that were identifiable from a distance. Wegmann (1961) observed that sedimentation of the coarse beds was 'cyclic' and that although this deposit superficially resembled the Infraquartzite Flysch, it was generally coarser. Leupold (in Wegmann 1961, p. 162) claimed that the base of the Supraquartzite Flysch in Calfeisental followed an interval of black slate. A 25 m thick interval of monotonous black shale was observed in conformable succession at the top of the Sardona Quartzite Formation type section at Troseggtobel (Fig. 7a) and was overlain by tectonically disrupted calcareous flysch beds.

I propose that Plattenkopf, on Alp Sardona in Calfeisental (741.8/199.8) should be the type section for the Supraquartzite Flysch (Fig. 1), principally because of its accessibility and relative lack of deformation. The thrusting and tight folding localised within this unit means that few continuous sections exceeding 20 m thickness can ever be found. A 60 m thick composite measured section through the unit at Plattenkopf appears to show an overall coarsening- and thickening-upwards (Fig. 13). There is a much larger proportion of exotic sedimentary material, especially dolomite, in the Supraquartzite pebbly conglomerates than is present in the Sardona Quartzite Formation, although they contain almost the same components. Repeated cycles of thickening and coarsening may characterise the development of the Supraquartzite Flysch, but a lack of continuity of section at other localities prevented testing of this hypothesis.

Bed thickness is proportional to the maximum grain-size of the beds, thickening and

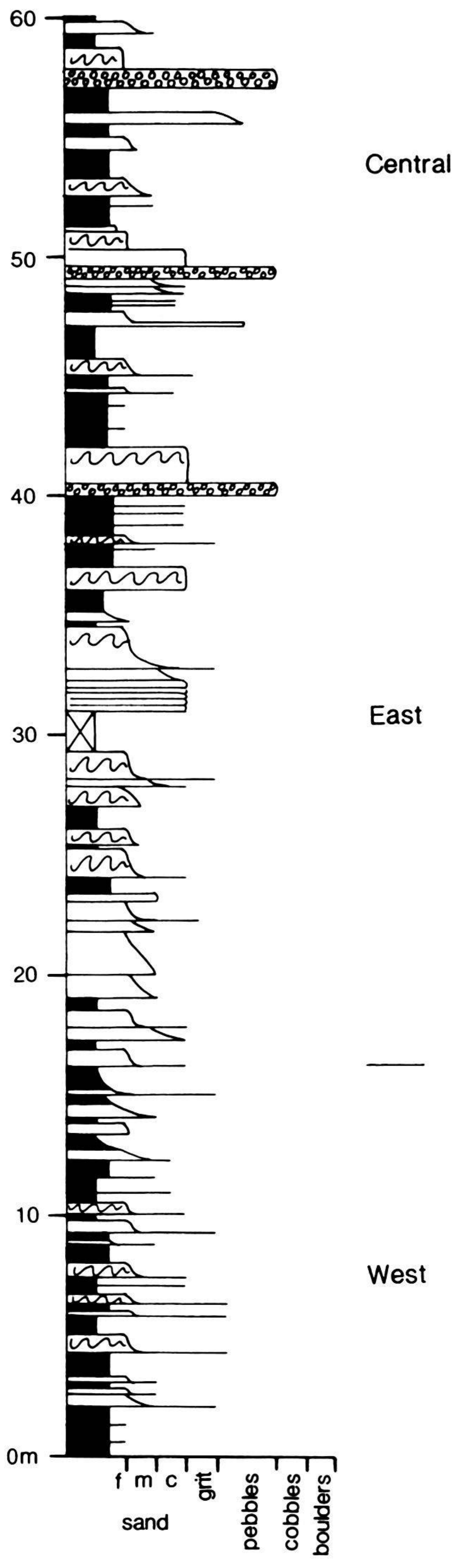

Fig. 13. Composite section through the Supraquartzite Flysch at Plattenkopf, Alp Sardona, in Calfeisental (see Fig. 1 for location & Pl. 3 for detailed sections), which illustrates an upwards-thickening and -coarsening trend.

thinning trends being closely associated with coarsening and fining, respectively (Pl. 3). Many of the sandstones show stepwise grading, fining abruptly above a basal layer of coarse sand, grit, or fine pebbles (facies R & S of Lowe 1982), to undulose- or convolute-laminated fine sandstones lacking obvious grading (facies B_1 of Mutti 1979). The coarse material could have been carried as a traction carpet beneath a sandy or gravelly high-density turbidity current as part of a bipartite flow. Bipartite flows are possible where the upper turbulent part helps to maintain the shear forces and dispersive pressure in the lower gravelly part (Middleton & Southard 1978). Hence, deposition of the gravel leaves behind a sandy HDTC, which may be deposited on top as a normal-graded suspension layer (Lowe 1982) and thereby give rise to the stepwise-graded beds found in the flysch.

Convolute lamination is commonplace in the stepwise-graded sandstones; this feature is more often found in the Bouma C division (lower flow regime, ripple-laminated interval) of calcareous turbidites than in siliciclastic turbidites (Stow 1986). It is formed by plastic deformation during or immediately after deposition, or shortly after burial, in rapidly deposited sediments that become liquefied (Allen 1984). Most of the examples at Plattenkopf appear to have formed either during or immediately after deposition, exhibiting discordances within the deformed layer, or truncations at the upper surface of the bed (Pl. 3). The remains of ripple lamination and cross-bedding can occasionally be seen in convoluted layers.

The stepwise-graded beds that are a feature of the Supraquartzite Flysch at Plattenkopf, characterise the unit elsewhere in Calfeisental and in Weisstannental, but nowhere else is the unit as thick-bedded or coarse-grained as at Plattenkopf. The Supraquartzite Flysch in the Elm district consists of much thinner graded sandstone beds (facies E of Mutti 1979), where background sedimentation was from LDTC; the isolated lenses produced by some of these flows were probably due to a low sediment supply, or reworking by bottom currents.

The thick graded sandstones at Plattenkopf are either amalgamated or separated by alternating, thin-bedded fine sandstones, laminated, light grey-weathering marls and black mudstones (Fig. 14; facies D_3 of Mutti 1979). The marked colour change at the boundary between light grey, calcareous mudstone and black, non-calcareous mudstone probably coincides with the change from suspended sediment fallout from a dilute turbidity current to hemipelagic fallout. Only the upper 2–5 cm of the marl are intensely burrowed, indicating that rapid deposition of the marl precluded animal activity. Repeated generations of burrowers in this layer may indicate that there were long time intervals between turbidity currents. A lack of bioturbation in the black mudstone suggests that it was deposited in an anoxic environment. *Chondrites* traces are the most common in the marls, but only its ramifying tunnel systems are preserved: the shafts with which it was connected to the oxygenated sediment surface appear to have been removed due to erosion by a subsequent turbidity current, and therefore the marl interval was not completely preserved.

Spreiten traces identified as *Teichichnus* were found together with *Chondrites* traces in a boulder adjacent to the West Plattenkopf section. These ichnogenera form part of the *Cruziana* ichnofacies of Seilacher (1967), which he associated with the sublittoral zone (i.e., shelf zone, at water depths of 0–200 m). Such 'shallow-water' traces have been found preserved in deep-sea fan sequences together with orthodox 'deep-water' traces like *Helminthoida* (Crimes 1976; Crimes et al. 1981). A few *Helminthoida* traces were ob-

Fig. 14. Detail of the thin, heterolithic interbeds in the Supraquartzite Flysch at West Plattenkopf, Alp Sardona, Calfeisental (741.4/199.8): normal-graded, dark-weathering sandstones alternate with lighter-coloured marls and black, non-calcareous mudstones.

served in positive relief on the base of thin sandstones at Plattenkopf; their lack of preservation on the base of thicker sandstones probably signifies that grazing traces were generally removed by the erosive action of larger-volume turbidites.

Paleocurrents from the Supraquartzite Flysch seem to vary according to bed thickness (Fig. 9d), making their interpretation difficult. The restricted geographical spread of the measurements also means that any conclusions may not be generally applicable over a larger area. However, data from thick beds exclusively plot in the southern half of the rose diagrams, which may mean that the main clastic input was from the northern margin of the basin, but followed a radial distribution network. The modal paleocurrent direction of the thin beds may indicate derivation from the southern basin margin. However, there appears to be a great deal of scatter in the data from these thinner beds, so they cannot be confidently linked to a particular source area on either basin margin.

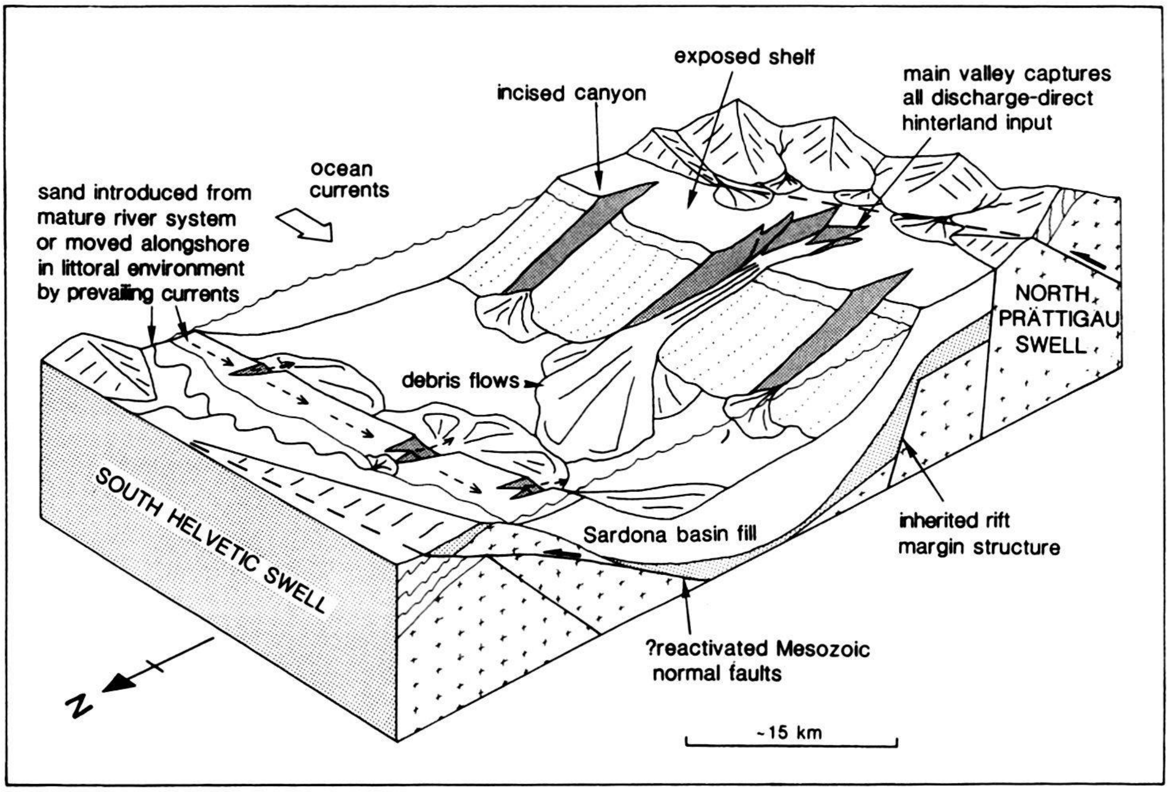

Fig. 15. Early Eocene reconstruction for the Sardona Quartzite Formation illustrating the effects of a Paleocene relative sea-level fall, which led to the development of a prograding fan system on the northern basin margin (South Helvetic Swell) and a series of debris flows being shed from the southern basin margin (North Prättigau Swell); see text for discussion.

4. Depositional environments

Over the period from the Maastrichtian (latest Cretaceous) to Bartonian (mid Eocene), ~650 m of sediment accumulated in the Sardona flysch basin (Fig. 2). The thinness of the accumulated sediment compared with other Alpine flysch basins suggests that it may have been deposited in a structurally restricted starved basin, where accommodation space and sediment availability limited sediment accumulation rates. Although sedimentation rates in the Sardona basin increased during the transition from pelagic to siliciclastic sedimentation, to 30–60 m/myr (Lihou 1995a), these rates are an order of magnitude less than the general range for flysch deposits (Schwab 1986).

The minimum (N-S) width of the Sardona flysch basin was 13–18 km, based upon considerations of the current structure (Lihou 1996). Rüefli (1959) guessed that the original basin width may have been 20–25 km, based upon the facies distribution. He also maintained that there were long-lived **South Helvetic** and **North Prättigau Swells** defining the northern and southern margins of the Sardona depocentre, respectively (Fig. 15). This situation is analogous to that of the Niesen flysch basin in western Switzerland, which was flanked by a marginal Ultrahelvetic swell to the north and the Tarine massif to the south (Homewood et al. 1984; Ackermann 1986). The Sardona and Niesen flysch basins may have been paleogeographically equivalent, based upon the similarity of their heavy mineral signatures and conglomerate clast compositions (Lihou 1995a, Lihou & Mange-Rajetzky in press).

The Sardona Flysch possesses all the typical characteristics of Cretaceous-Tertiary Swiss flysch listed by Homewood & Lateltin (1988): it is composed of first cycle detritus, contains clasts derived from Variscan basement and its Mesozoic sedimentary cover (Tab. 1), is composed of gravity flows, minor contourites and hemipelagic sediments, and sedimentation was controlled by tectonic factors more than by eustatic or climatic variations. Lihou & Allen (in press) consider that the transition from carbonate sedimentation to siliciclastic flysch sedimentation in the Sardona basin during the Maastrichtian, marked the beginning of the Alpine foreland basin. In previous publications, flysch sedimentation was considered to be syn-collisional, but not controlled by continental collision, and flexural subsidence was concluded to have begun only during North Helvetic Flysch deposition (Pfiffner 1986; Lateltin 1988; Homewood & Lateltin 1988).

In contrast to most Alpine flysch basins, the main source of siliciclastic material for the Sardona Quartzite Formation was from the north to northeast, as documented earlier. Erosion of the neighbouring Helvetic shelf, whose Mesozoic deposits consist mainly of limestone and shale (Trümpy 1980), is unlikely to have contributed much quartz sand to the Sardona basin, contrary to the suggestion of Homewood & Caron (1982). Added to this, the Sardona basin was probably isolated from the Helvetic margin by the South Helvetic Swell, which itself was a more likely source for the quartzite. The sandy Sardona Quartzite is supermature, containing more than 90% quartz, even taking into account its subsequent silicification (Lihou & Mange-Rajetzky in press). Consequently, it may have been derived from a mature river system providing a large volume of sand. However, a river system developed on a narrow structural high such as the South Helvetic Swell would have a short transport distance from source to basin and therefore supply relatively immature detritus. If the sand was instead transported along the South Helvetic Swell from a laterally distal source and introduced into the basin at discrete points, this would explain both its maturity and the preserved paleocurrents. A possible distal source for the Sardona Quartzite would be the Bohemian massif in southeastern Germany, now approximately 200–250 km away, which was exhumed by reverse faulting along its southwestern border during the Paleocene (Ziegler 1990). This material could have been transported into the Sardona basin by prevailing westerly directed longshore drift in the littoral zone and/or a fluvial drainage basin flowing to the west (Fig. 15).

The same pattern of changing facies associations within the Sardona Quartzite Formation can be seen throughout the basin (Fig. 7). There are three elements to this pattern that require explanation: firstly, the lower, sandy Sardona Quartzite has a sharp contact with the underlying Ölquartzite; secondly, deposition of the supermature sandy Sardona Quartzite is accompanied by the influx of extremely immature debris flow boulder conglomerates into the basin to form the Crystalline Conglomerate; and thirdly, the sequence is terminated by the deposition of the pebbly Sardona Quartzite. The temporal pattern of changing facies associations within the Sardona Quartzite Formation could be explained by a single mechanism of a large-magnitude relative sea-level fall (RSL) during deposition of the Ölquartzite (Fig. 15).

If sea-level in the Sardona basin fell rapidly at the beginning of the Paleocene, exposing former shelf areas, rivers may have discharged directly into the basin at the top of its marginal slopes (Fig. 15). Hyperpycnal flow from sediment-charged rivers in flood could then have resulted in relatively uniform LDTC being fed into the basin (Normark & Piper 1991), to be deposited as thin sandstones in the Ölquartzite unit. Infrequent sandy

HDTC and debris flows, which produced the coarse sandstones and conglomerates of the Ölquartzite unit, would have been conveyed into the deep basin via submarine canyons that were incised into the northern basin slope owing to the fall in RSL (Fig. 15). Slope incision and winnowing of shelf sand during the RSL fall would have meant that reworked sand could be incorporated into the sandy HDTC that formed the sandy Sardona Quartzite unit, as they passed through the submarine canyons en route to the basin. These turbidites were probably deposited as lobes at the end of their feeder channel systems, where there was an abrupt decrease in the gradient, or the flows became unconfined (Normark & Piper 1991). The distal deposits of these turbidites probably formed the thicker sandstones in the Ölquartzite unit. Deposition of the Ölquartzite unit was probably confined to areas of relatively low sediment accumulation on the lower fan or basin floor, or in topographic lows between sandy lobes on the middle fan. A bimodal distribution of paleocurrent directions, and sedimentary characteristics, suggest that they were prone to reworking by westerly directed contourites acting in the narrow basin.

The sudden appearance of the sandy Sardona Quartzite may be due to rapid progradation or lateral migration of fan lobes across the basin floor, smothering the older Ölquartzite deposits. Cyclic thickening and thinning of the sandy Sardona Quartzite beds may be related to lobe progradation and retreat. Temporary lobe switching by flow stripping of thick sandy turbidites at channel bends, as documented on the Pleistocene Navy Fan (Piper & Normark 1983), could account for repetitions of Ölquartzite-like heterolithic intervals within the sandy Sardona Quartzite.

The pebbly Sardona Quartzite probably developed in a more proximal, distributary channel setting. The presence of some mechanically unstable clasts in these pebbly sandstones and conglomerates favours limited or no marine storage en route to the basin, so they were not predominantly derived from reworked shelf material, as were the Ölquartzite and sandy Sardona Quartzite units. The canyons and distributary channels feeding the sandy Sardona Quartzite lobes could have acted as conduits for pebbly material reaching the edge of the shelf during large flood events in the fluvial system. The pebbly conglomerates may represent the channellised, residual coarse fraction of very large-volume turbidites that were flow stripped and their upper portions deposited as sandy Sardona Quartzite lobes, whilst the massive sandstones and pebbly sandstones may represent HDTC and cohesionless debris flows, respectively, that remained confined within the distributary system.

The Ölquartzite, and sandy-pebbly Sardona Quartzite facies associations are probably genetically related. According to the deep-sea fan models of Mutti (1979), the Sardona Quartzite Formation may represent a 'highly efficient' system composed of separate distributary and distributional systems: the pebbly Sardona Quartzite may represent the remains of the distributary system, the sandy Sardona Quartzite may represent the prograding lobe and lobe-fringe deposits, and the Ölquartzite may represent the interlobe and basin plain deposits. In terms of this 'efficient fan' model, the temporal pattern of facies associations exhibited by the Sardona Quartzite Formation suggests basinward progradation of the fan system during the Paleocene/early Eocene lowstand.

The Pleistocene Navy Fan off the California Borderland could be a reasonable analogue for the Sardona Quartzite Formation, in terms of its size (radius of ~15 km), its constricted basin setting and the water depth in which it formed (1,800 m). Piper & Normark (1983) documented its development, linking the growth of sandy lobes to a late

Pleistocene eustatic lowstand, when the Tijuana River fed directly into the Coronado Canyon at the edge of the continental shelf.

Deposition of the immature debris flow Crystalline Conglomerate within the Sardona Quartzite Formation can be explained by direct hinterland input from the fluvial into the deep-marine system, via submarine canyons cut at the edge of the shelf. It may have been sourced from the southern basin margin. Hence, there would have been two sources contributing material to the flysch basin at this time, on the northern and southern basin margins, namely the South Helvetic and North Prättigau Swells, respectively (Fig. 15). The Sardona basin is the only example of an Alpine flysch basin being supplied by coeval, dual sources. Its paleocurrent pattern is also markedly different to that of most Alpine flysch basins, in that the northern margin acted as the principal source area. By contrast, in most other flysch basins, the souhern margin was the principal source area (Wildi 1985; Caron et al. 1989).

The switch back to carbonate flysch sedimentation in the Sardona basin at the beginning of the middle Eocene can be explained by drowning of the shelf areas, creating a greater area for carbonate production within the photic zone. The condensed unit of black shales locally preserved between the top of the Sardona Quartzite Formation and the base of the overlying Supraquartzite Flysch, may result from sediment starvation during the main transgressive phase. The fact that a shallow-marine fauna was incorporated into the Supraquartzite Flysch also suggests that the shelf area had expanded enough to allow recolonisation by shallow-water organisms. Shelfal water depths may have increased over time, evidenced by the decreasing proportion of reworked, shallow-water benthic foraminifera preserved in the flysch (Wegmann 1961).

Abundant *Helminthoida* and *Chondrites* trace fossils in the Supraquartzite Flysch suggest that sedimentation still took place in the bathyal-abyssal realm. The calcareous turbidites may have been deposited below the calcite compensation depth (CCD), since background hemipelagic mudstones preserved between the turbidites are non-calcareous. Rapidly deposited, allochthonous carbonate within the turbidites could still be preserved below the CCD.

The pebbly conglomerates that are found in the Supraquartzite Flysch may have resulted from submarine reworking of lag deposits remaining from the Paleocene lowstand, or cannibalisation of older conglomerates. Tectonic activity in the source areas, namely the South Helvetic and North Prättigau Swells, locally exposing Mesozoic cover rocks to submarine erosion, may be the reason for the predominance of sedimentary clasts in the Supraquartzite Flysch. Paleocurrents from thin Supraquartzite Flysch turbidites suggest that they originated from both the northern and southern basin margins, but thicker flows came from the northern margin only. The 180° spread in the unimodal paleocurrent indicators from these thicker beds (Fig. 9d) may have arisen because they were deposited from a radial distribution network.

5. Conclusions

The Sardona unit preserves a continuous record of sedimentation on the Ultrahelvetic margin of Europe from the Late Cretaceous (Cenomanian) to mid Eocene (Bartonian). The oldest preserved sediments are the Cenomanian-Campanian Globotruncana Limestone and Marl, interpreted to be pelagic and hemipelagic deposits from a starved, deep-marine environment. These give way gradually to Maastrichtian age, thin-bedded, calcareous turbidites of the Infraquartzite Flysch, which were probably deposited in bathyal-abyssal water depths within a structurally restricted, starved basin enclosed between the South Helvetic and North Prättigau Swells. Sedimentation rates within the Sardona flysch basin were only 30–60 m/Myr.

At some time in the early Paleocene, a change from calcareous to siliceous turbidite sedimentation took place that was probably associated with a large-magnitude RSL fall which exposed shelf areas to erosion. This RSL fall allowed direct discharge from the source area into the deep-marine basin via submarine canyons cut at the edge of narrow shelves bordering the basin. Paleocene-lower Eocene lowstand deposits are represented by the Sardona Quartzite Formation, which is subdivided into four facies associations. The Ölquartzite and sandy-pebbly Sardona Quartzite units are interpreted as a 'highly efficient' fan system that prograded out into the basin from the northern basin margin; the thin, heterolithic interbeds of the Ölquartzite represent low-volume, dilute turbidites deposited in areas of low sedimentation on the basin plain, or between ephemeral sandy lobes; lobe deposits are preserved as the large-volume turbidites of the sandy Sardona Quartzite, which were composed of mature, reworked shelf material and distally derived sand perhaps originating from the Bohemian massif; the pebbly Sardona Quartzite comprises the more proximal, pebbly sandstones and conglomerates from the distributary system. The fourth facies association of the Sardona Quartzite Formation is the cobble- to boulder-grade Crystalline Conglomerate, which is composed of intraformational and exotic, sedimentary and crystalline clasts and is interpreted as a series of cohesive debris flows derived from the southern basin margin. Hence, there was a dual supply to the Sardona basin from narrow structural highs (South Helvetic and North Prättigau Swells), which acted as independent sediment sources and restricted input from external sources.

The youngest preserved sediment in the Sardona unit, the calcareous Supraquartzite Flysch of mid Eocene age, superficially resembles the Infraquartzite Flysch, but is much coarser. The switch back to calcareous turbidite sedimentation, probably at the beginning of the Lutetian, can be explained by a drowning of the shelf area, creating a greater area for carbonate production within the photic zone. This hypothesis is supported by the fact that a condensed unit of black shales, perhaps representing sediment starvation during the transgression, is locally preserved between the Sardona Quartzite Formation and the Supraquartzite Flysch. The fact that a shallow-marine fauna is incorporated into the Supraquartzite Flysch also suggests that the shelf area had expanded enough to allow recolonisation by shallow-water organisms. Some deepening of the Sardona basin during the mid Eocene is postulated, such that the Supraquartzite Flysch was deposited below the local CCD. Deposition within the Sardona basin was probably terminated in the Bartonian when the area was overthrust from the south and incorporated into the orogenic thrust wedge.

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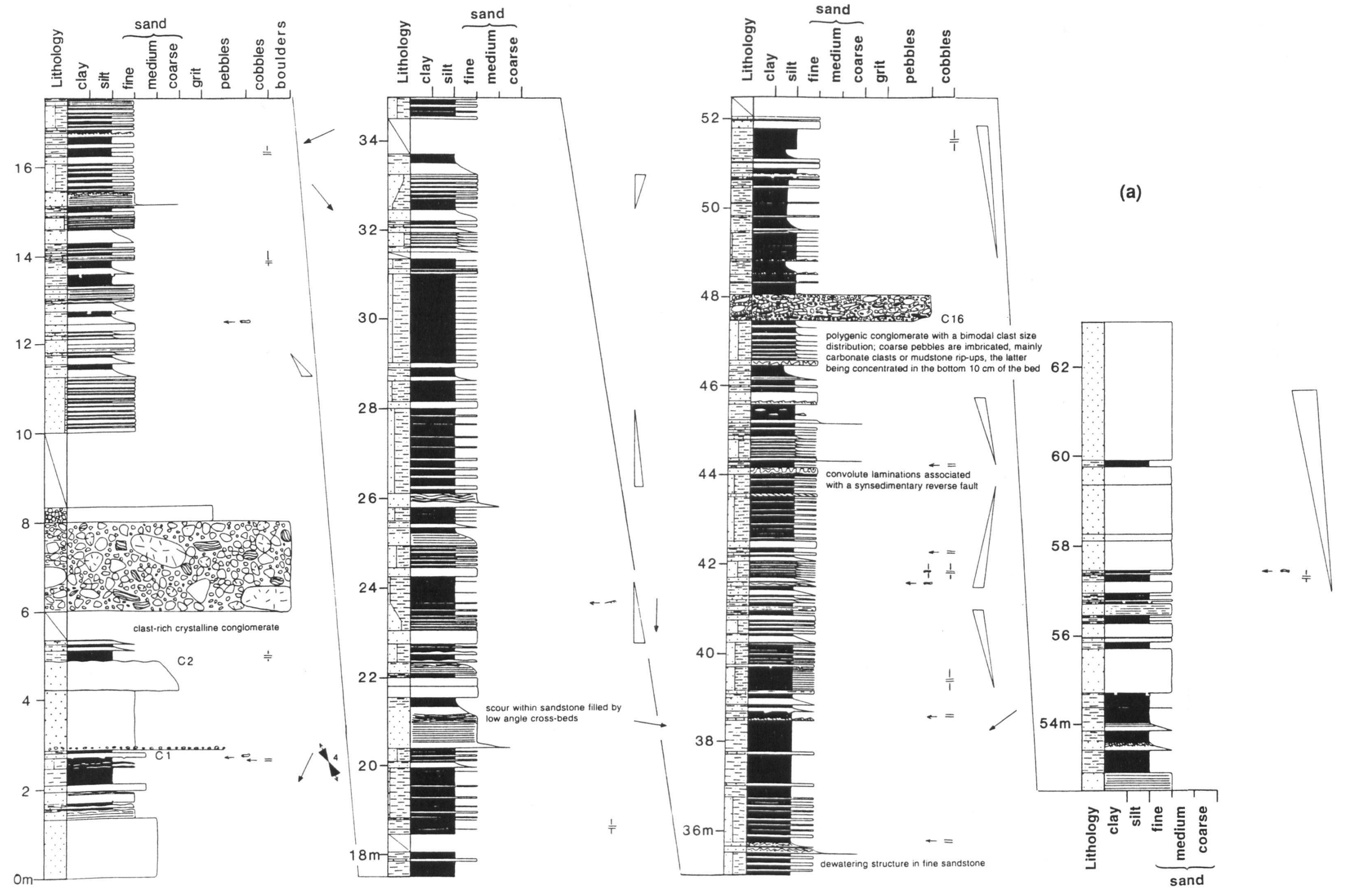

Plate 1. Measured section through the Sardona Quartzite Formation (lower part) at its type section at Jakobiweid (742.1/193.1; 1740 m) on Alp Sardona in Calfeisental.

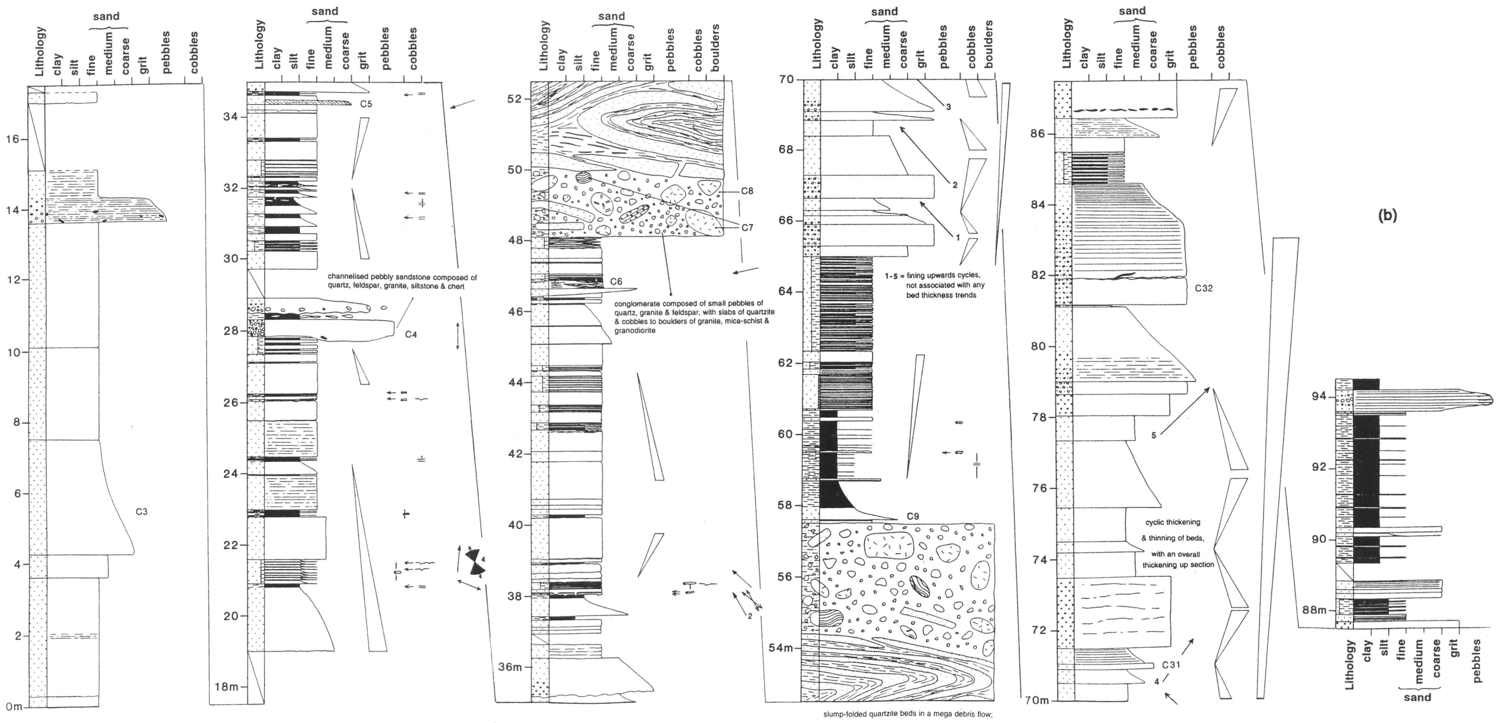

Plate 2. Measured section through the Sardona Quartzite Formation (upper part) at its type section at Troseggtobel (74).1/198.0; 1800 m) on Alp Sardona in Calfeisental.

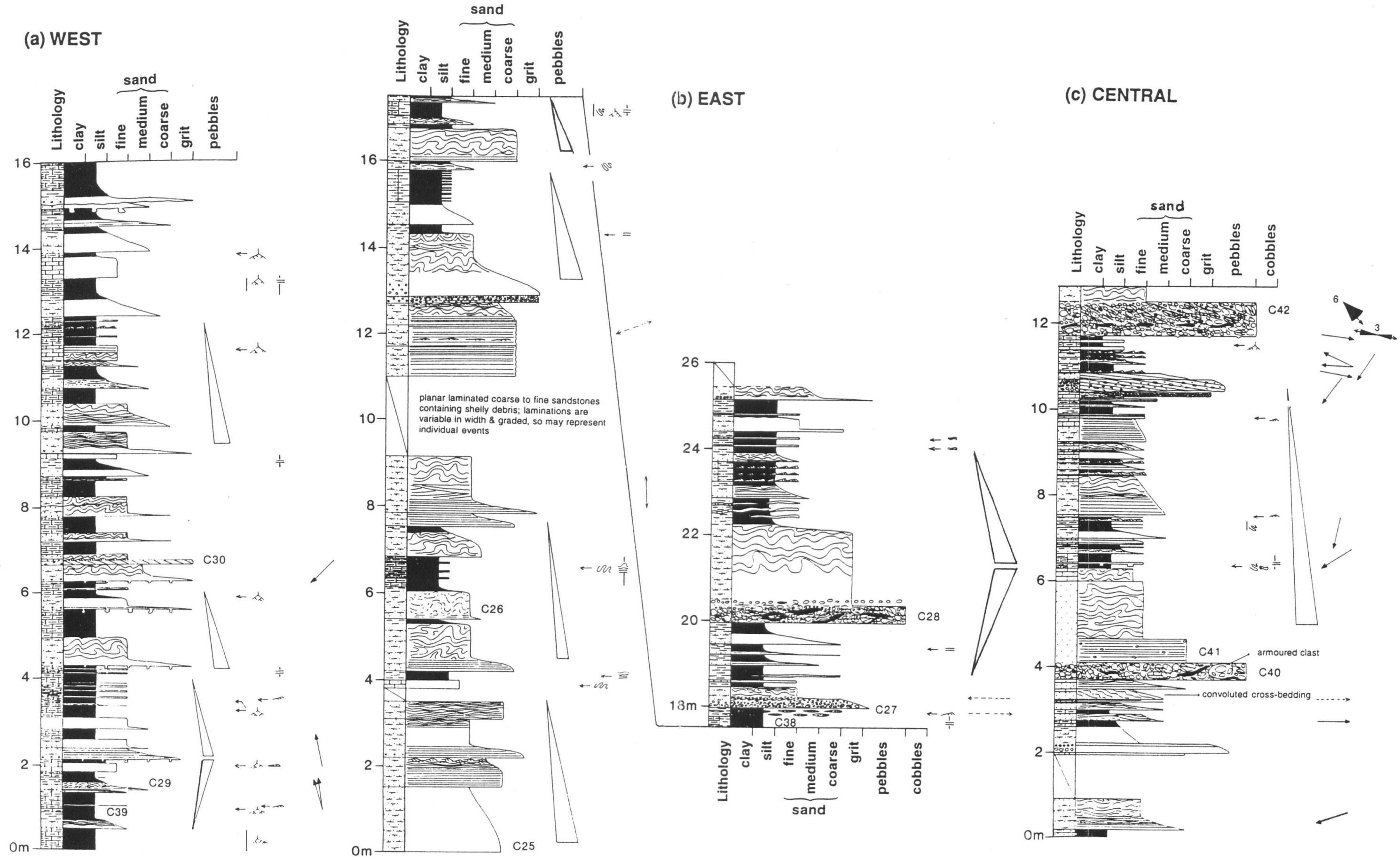

Plate 3. Measured sections through the Supraquartzite Flysch Formation at its type section at Plattenkopf, Alp Sardona, Calfeisental: (a) western section (741.4/199.8; 2260 m), (b) eastern section (741.9/199.8; 2200 m) and (c) central section (741.8/199.8; 2240 m).