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ANISIAN SEDIMENTATION AND TECTONICS OF THE M. PORE – M. CERNERA AREA (DOLOMITES)

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Riassunto. Vengono descritti l'ambiente deposizionale e la tettonica anisica nella regione del M. Pore e M. Cernera (Dolomiti orientali).

La parte anisica (?) della *Formazione di Werfen* (*Membro di Cencenighe, Membro di San Lucano*) è caratterizzata da una sedimentazione clastica e carbonatica, che passa a una facies evaporitica nella parte superiore. Dopo una breve emersione si sono depositati carbonati peritidali (*Formazione del Serla inferiore*). Nel Pelsonico inferiore questa piattaforma si disintegra. La deposizione del sovrastante *Gruppo di Braies* comincia con conglomerati fluviali (*Conglomerati di Piz da Peres intermedi*). La trasgressione successiva inizia con sedimenti di spiaggia (*Formazione di Agordo*) e passa a siltiti bacinali (*Formazione di Dont*, Pelsonico – Illirico inferiore). Nell'Illirico inferiore una regressione causa l'avanzamento di una piattaforma carbonatica («*formazione calcarea*») da ovest ad est verso il bacino. Forti afflussi terrigeni soffocano localmente lo sviluppo della piattaforma nella parte inferiore (?) dell'Illirico superiore (*Conglomerati di Piz da Peres superiori*). Nell'Illirico superiore si estende di nuovo l'ambiente di una piattaforma carbonatica (*Formazione del Serla superiore*). Contemporaneamente si approfondisce un bacino verso est, dove vengono depositate marne e torbiditi (*Formazione dell'Ambata*).

Nella sedimentazione dell'Anisico si riflette il margine orientale di una dorsale («*Dorsale badioto–gardenese*»), che causa l'assottigliamento degli strati verso ovest, nonchè il sollevamento e l'emersione nel Pelsonico inferiore e nell'Illirico superiore. Una soglia parzialmente emersa a sud fornisce conglomerati di tipo *debris flow* dal Pelsonico fino all'Illirico superiore.

La tettonica sinsedimentaria si manifesta con fratture a rigetto subverticale e pieghe–faglie prevalentemente in direzione NNE–SSO. La sedimentazione e la tettonica anisica sono collegabili ad una coppia di sforzi di taglio con trascorrenza sinistra ONO–ESE. E' probabile che esse causassero una faglia trascorrente sinistra con andamento ONO–ESE immediatamente a sud dell'area esaminata, la quale durante la compressione alpidica si trasforma in un sovrascorrimento (Linea del Cernera, Linea di Selva). La coppia di sforzi di taglio anisica potrebbe aver causato faglie distensive e vulcanismo nella regione del M. Pore – M. Cernera ed il sollevamento «*compressivo*» della «*Dorsale badioto–gardenese*» ad ovest.

Kurzfassung. Die vorliegende Arbeit befasst sich mit den Ablagerungsbedingungen der anisischen Sedimente und synsedimentärer Tektonik des Gebiets um den M. Pore und den M. Cernera (östliche Dolomiten).

Der anisische (?) Teil der *Werfener Formation* (*Cencenighe Member, San Lucano Member*) ist durch klastische und karbonatische Sedimentation gekennzeichnet, die im oberen Teil der Abfolge durch Evaporite abgelöst wird. Nach einer kurzen Emersion werden peritidale Plattformkarbonate abgelagert (*Untere Sarlformation*). Im Unterpelson zerbricht diese Karbonatplattform. Die darüber folgende *Formationsgruppe der Pragser Schichten* setzt mit fluviatilen Konglomeratschüttungen ein (*Mittlere Peresschichten*). Eine anschließende Transgression führt zur Ablagerung von strandnahen Sedimenten (*Agordo–*

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Formation) und schließlich von siltigen Beckensedimenten (*Dont–Formation*, Pelson – Unterillyr). Eine Regression ermöglicht im Unterillyr das Übergreifen einer Karbonatplattform («*Kalkige Formation*») über die verflachenden Beckenareale. Verstärkte Zufuhr terrigenen Schutts erstickt das Plattformwachstum lokal (*Obere Peresschichten*, unteres ? Oberillyr). Im Oberillyr schließlich bedeckt eine Karbonatplattform (*Obere Sarlformation*) den größten Teil des Gebiets, während sich im Osten ein Becken eintieft, in dem Mergel und Turbidite zur Ablagerung gelangen (*Ambata–Formation*).

Während des Anis beeinflusst der Ostrand einer strukturellen Hochzone («*Dorsale badioto–gardenese*», im Westen des Gebiets) die Sedimentation. Die Hochzone macht sich durch ein generelles Ausdünnen der Schichten nach Westen und im Unterpelson und Oberillyr durch Hebung und Abtragung bemerkbar. Eine teilweise aufgetauchte Schwelle im Süden kann durch Schüttung grobklastischen Materials im Pelson und Illyr nachgewiesen werden.

Synsedimentäre Tektonik macht sich durch subvertikale, generell NNE streichende, Brüche bemerkbar. Das Sedimentationsmuster und die anisische Tektonik können durch WNW–ESE streichende Scherungskräfte mit Linksseitensinn verursacht worden sein. Sie erzeugten vermutlich eine WNW–ESE streichende Linksseitenverschiebung unmittelbar südlich des Gebiets, die während der alpidischen Orogenese als Aufschiebung reaktiviert wurde (Cernera–Linie, Linie von Selva). Die anisischen Scherungskräfte können für Bruchtektonik und pelsonischen Vulkanismus im Gebiet des M. Pore – M. Cernera und für die «kompressive» Hebung des «Dorsale badioto–gardenese» verantwortlich gemacht werden.

Introduction.

The Dolomites have been a classical ground for the investigation of lateral facies changes since a century. Especially the world famous Middle– to early Upper Triassic facies heteropy of basinal and carbonate platform («reefal») facies have been extensively studied (Mojsisovics, 1879; Leonardi, 1968; Cros, 1974; Bosellini & Rossi, 1974; Assereto et al., 1977; Gaetani et al., 1981).

During the last decade, the Anisian («Alpine Muschelkalk») deposits of this region have attracted the interest of several authors (Bechstädt & Brandner, 1970; Farabegoli et al., 1977; Assereto et al., 1977; Pisa et al., 1979). The Ani-

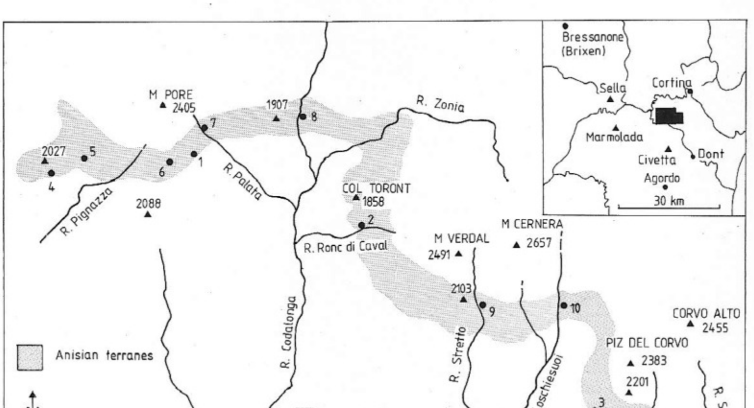


Fig. 1 – Location of the investigated area: 1–11 location of the stratigraphic columns of Fig. 4–6.

sian of the M. Pore – M. Cernera area (Fig. 1, 2), however, only marginally has been involved in the investigations, chiefly of more ancient authors (Nöth, 1929; Van Houten, 1930).

A major handicap of stratigraphic work in the Southern Alps is the almost infinite and confusing number of locally applied stratigraphic terms. A recent

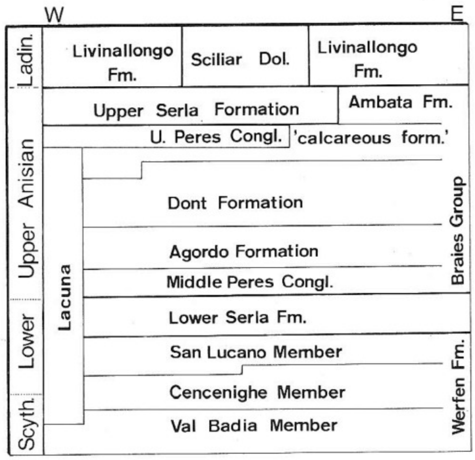


Fig. 2 – Anisian stratigraphy of the M. Pore–M. Cernera area.

attempt to unify the Scythian and Anisian lithostratigraphic nomenclature all over the Southern Alps was made by De Zanche & Farabegoli (1982).

The aims of the present work are to study the Anisian sedimentation pattern and synsedimentary tectonics of the M. Pore – M. Cernera area.

Regional setting.

The Anisian terranes of the Dolomites roughly can be subdivided into two areas. In the Western and Central Dolomites the Anisian deposits generally do not exceed 100 or 150 meters in thickness, while in the Eastern Dolomites (Braies, Valdora, Cadore, Agordino, Zoldano) thickness may reach as much as 500 meters.

In the M. Pore – M. Cernera area thin Anisian deposits in the west (40 m), opposed to thick ones (ca. 500 m) only a few kilometers to the east (Fig. 3), suggest a transitional zone from the structural rise to the sedimentation basin.

The Anisian beds occur in a tectonically simply formed unit («Pore Schubmasse» according to Nöth, 1929, «Cernera–Rocchetta Schubmasse» according to Van Houten, 1930), which comprises Upper Permian (Bellerophon Formation) to Norian (Hauptdolomit) deposits. It is confined to the south by a northward dipping and approximately WNW–ESE running thrust fault with an apparent dislocation of about 1000 meters («Linea di Selva» according to Leonardi, 1968, p. 888). Towards the west, Anisian terranes crop out as far as the Livinallongo area, in the east they are limited by a thrust (Rio Sacuz).

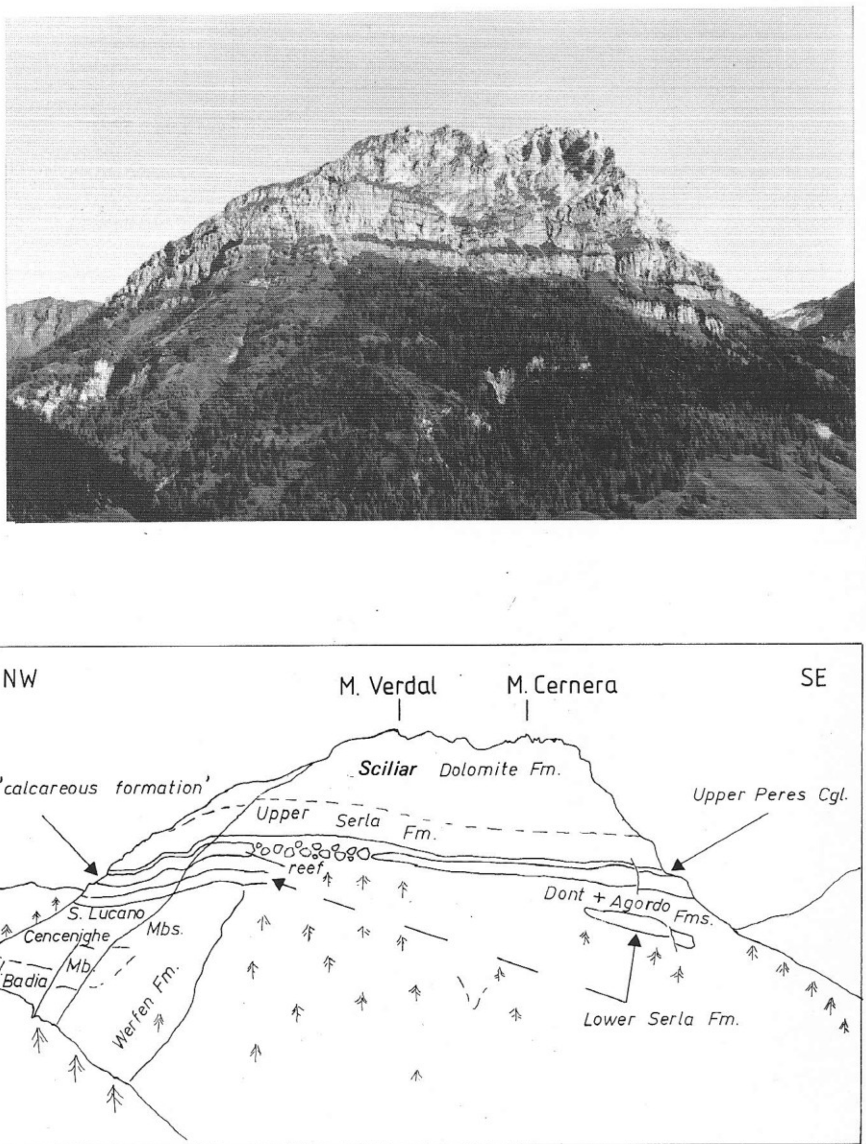


Fig. 3 – The Cernera massif, seen from Colle S. Lucia. Note the stratigraphy and the interruption of the «calcareous formation» at the southwestern slope of M. Verdal.

Stratigraphy

Werfen Formation (Fig. 4).

According to Brandner & Mostler (1982), the Werfen members overlying the Val Badia Member are supposed to be partly of Lower Anisian age. This interpretation is based upon the dating of conodonts of the Val Badia Member in the Western Dolomites, which indicate an uppermost Scythian age (Brandner & Mostler, 1982, p. 25).

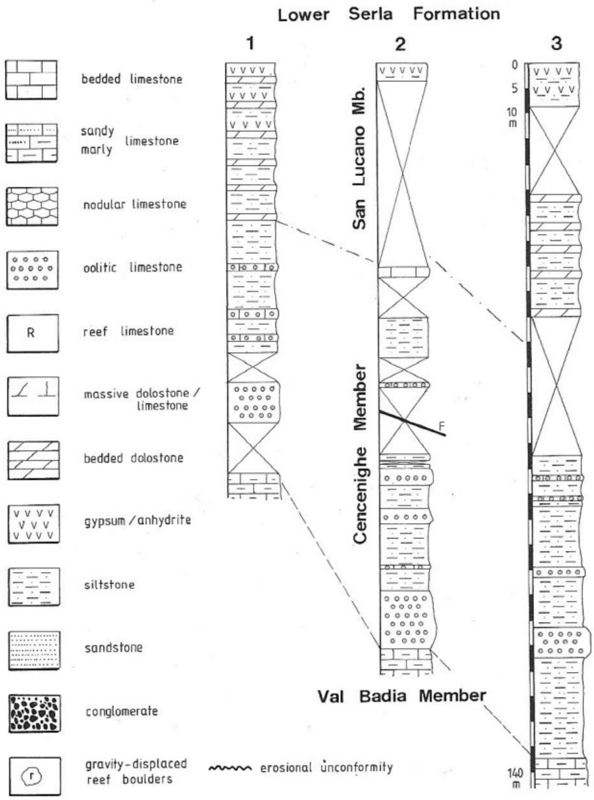


Fig. 4 – Stratigraphic columns of the upper part of the Werfen Formation. For exact location see Fig. 1.

Cencenighe Member (Upper Scythian ? Lower Anisian). The Cencenighe Member (Farabegoli et al., 1977) is composed of varicoloured siltstones and oolitic limestones (Fig. 4, column 1, 2, 3). Fine grained sandstones occur subordinately (Fig. 5, column 8).

The clastic beds, forming the bulk of the sequence, frequently exhibit even lamination at the base grading into ripple lamination at the top. Oscillation ripples as well as burrows commonly occur on the top surfaces. Slumping phenomena have been observed at the base of M. Cernera and Piz del Corvo.

According to Seilacher (1982), most of this clastic sequence seems to represent *storm deposits* (tempestites).

The calcareous deposits are built up by oolitic limestones. They are mostly thick–bedded or massive, generally red and frequently cross–bedded. Among their microfacies, packstone depositional texture prevails. Fossil remains are represented by Foraminifera, shell fragments and Echinoderm debris. The oolites generally have a size of less than one millimeter and display thin hematitic crusts.

These deposits here are interpreted as *oolitic bars.*

The boundary to the underlying Val Badia Member (Bosellini, 1968) is marked by a sharp facies change from grey, micritic, marly limestones to sandstones, siltstones and oolitic limestones. Thickness varies from approximately 40 meters in the west to 90 meters in the east.

The depositional environment of the Cencenighe Member is interpreted as a *tidal flat* with prevailing subtidal conditions (Farabegoli et al., 1977, p. 664).

San Lucano Member (Upper Scythian ? Lower Anisian). The San Lucano Member (Pisa et al., 1979) originally was considered to be a member of the Lower Serla Formation. Here it is included in the Werfen Formation as in Casati et al. (1982), because the Werfen facies continues in this sequence.

The San Lucano Member is distinguished from the Cencenighe Member by the absence of oolitic limestones and the occurrence of dolomites and evaporites (Fig. 4, column 1, 2, 3). Furthermore, the clastic facies frequently exhibits a violet colouration (compare Nöth, 1929, p. 140).

The dolomites are bedded, yellowish, microcrystalline rocks. Farabegoli et al. (1977, p. 667) interpreted them as *supratidal deposits.* In the upper part of the sequence the dolomites become pinky and contain nodules of anhydrite and gypsum. The uppermost meters are composed of finely laminated, micaceous siltstones alternating with thin layers of gypsum or anhydrite (well exposed 500 m SW of Piz del Corvo at 1875 m, and in the M. Pore area 300 m NE of the 2088 m mark at 2050 m).

The boundary to the Cencenighe Member was drawn at the first occurrence of dolomites, the Lower Serla Formation overlies the San Lucano Member with a sharp contact. Thickness varies from 30 meters in the west to 50 meters in the east.

According to Farabegoli et al. (1977, p. 686), the San Lucano Member was deposited on a *tidal flat* with frequent supratidal conditions. The uppermost evaporitic sequence strongly reminds of *shabka* conditions.

Lower Serla Formation (Lower Anisian).

The Lower Serla Formation («Unterer Sarldolomit» of Pia (1937), «Lower Serla Dolomite» of De Zanche & Farabegoli (1982)) corresponds to the «Dolomia di Frassenè» member of the Lower Serla Dolomite of Pisa et al. (1979). The term «Dolomia di Frassenè» becomes obsolete if the San Lucano Member generally will be included in the Werfen Formation.

The Lower Serla Formation is composed of bedded, white to pale grey, microcrystalline dolomites. They generally are composed of algal mats with abundant birdseyes. Intraclasts, pellets and rare oolites occur, as well as authigenic quartz grains. Rare Foraminifera constitute the only fossil remains. Terrigeneous influence is completely absent despite an approximately 30 centimeters thick horizon of marly dolomites at 5 meters above the base of the formation, which is correlatable over the whole area.

At the base of the formation, a thin (few centimeters) conglomerate with pebbles derived from the underlying San Lucano Member, embedded in a silty–dolomitic matrix, has been observed. The Lower Serla Formation is covered by the Braies Group, which overlies these deposits with a sharp and frequently erosional contact.

Thickness varies from 26 meters (M. Pore) to a maximum of 56 meters at the base of Piz del Corvo. Syndepositional faulting in this area is suggested by the abrupt change of thickness of the Lower Serla Formation along fractures. In the immediate vicinity of the faults, the lower part of the dolomites exhibits normal drag flexure, while the upper part is almost undisturbed.

The Lower Serla Formation represents a *peritidal carbonate bank*. According to Richter (1971), authigenic quartz grains may indicate hypersaline conditions.

Braies Group (Upper Anisian) (Fig. 5, 6).

The term «Pragser Schichten» (Prags = Braies) was introduced by Pia (1937). According to their subdivision into several formations and members it has to be considered a group (De Zanche & Farabegoli, 1982; Brandner & Mostler, 1982).

A general feature of this group is the predominance of terrigeneous deposits. Calcareous deposits occur only subordinately.

Peres Conglomerates. The Peres Conglomerates («Peresschichten» of Pia, 1937) constitute two coarse clastic horizons at the base and at the top of the group.

Middle Peres Conglomerates (Lower Pelsonian). The Middle Peres Conglomerates (Fig. 5, column 5 (A); Fig. 6, column 9, 10, 11 (A)) correspond to the «Untere Peresschichten» of Pia (1937), the «Mittlere (and Untere?) Peresschichten» of Bechstädt & Brandner (1970), the «Conglomerati di Voltago» of Pisa et al. (1979) and the «Richthofen Konglomerat» of Nöth (1929) and Van Houten (1930).

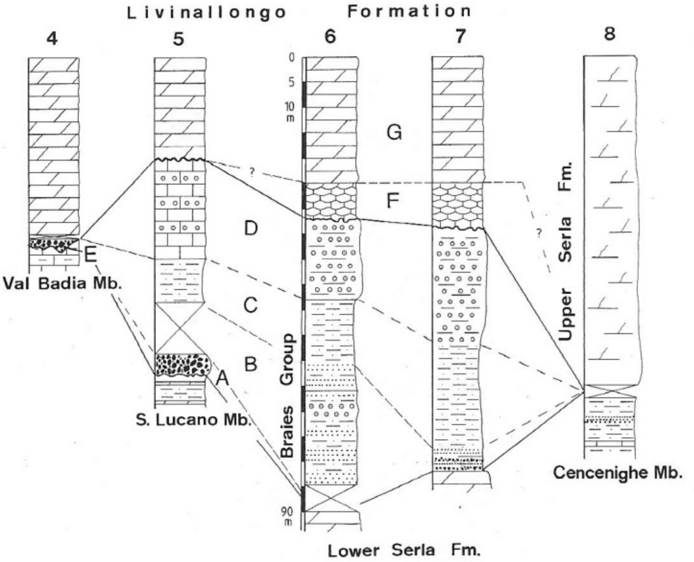


Fig. 5 – Stratigraphic columns of the Braies Group and the Upper Serla Formation of the M. Pore area. For exact location see Fig. 1 and for the legend see Fig. 4. A) Middle Peres Conglomerates; B) Agordo Formation; C) Dont Formation; D) «calcareous formation»; E) Upper Peres Conglomerates; F) Morbiac Limestone of the Upper Serla Formation; G) dolomites of the Upper Serla Formation.

West of Rio Codalonga, the packed conglomerate beds contain low–sorted components of the Cencenighe Member, the San Lucano Member, the Lower Serla Formation, and, subordinately, the Val Badia Member. They form thick beds, separated by thin layers of red or grey, pebbly sandstones or siltstones. The conglomerate pebbles generally are angular to subrounded, indicating only moderate transport. They may reach a size of up to 75 centimeters (components of the Lower Serla Formation) at the base of the unit. Pebbly sandstones form the transitional zone to the overlying Agordo Formation.

East of Rio Codalonga, the conglomerates display an identical fabric to those of the western area. They are distinguished from these, however, by the predominance of clasts derived from the Lower Serla Formation and the San Lucano Member. Pebbles of the Cencenighe Member occur subordinately. These rocks, therefore, have a yellowish colouration, in contrast to their red counterparts in the M. Pore area.

At some places, the conglomerates deeply cut into the subsurface. At the southern slope of M. Pore an erosional channel has been observed (Fig. 7). In this channel, the Middle Peres Conglomerates eroded the entire Lower Serla Formation and approximately 15 meters of the San Lucano Member. Another palaeovalley is exposed at the southern precipice of Piz del Corvo (Rio Giausel), where the conglomerates eroded about 10 meters of the underlying dolomites.

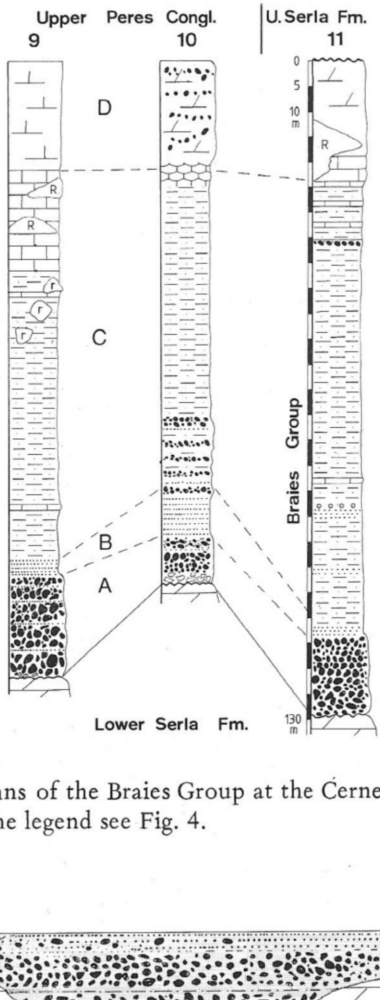


Fig. 6 – Stratigraphic columns of the Braies Group at the Cernera massif. For exact location see Fig. 1 and for the legend see Fig. 4.

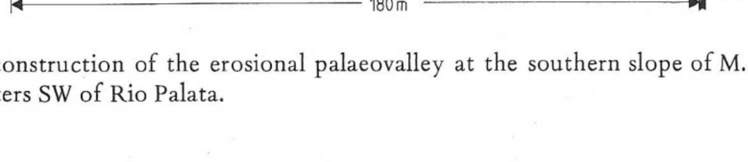


Fig. 7 – Reconstruction of the erosional palaeovalley at the southern slope of M. Pore, 200 meters SW of Rio Palata.

The eastern margin of this 200 meters wide area is constituted by a synsedimentary fault.

In the Rio Giausel – Rio Sacuz area, synsedimentary block faulting yielded few hundred meters wide horst – and graben zones, which are bounded by conglomerates of extremely varying thickness. Additionally, an angular unconformity of about 10° to the underlying Lower Serla Formation can be observed at some places.

The depositional fabric of the Middle Peres Conglomerates (Fig. 8 A) generally is referred to as *fluvial deposits* (Dal Cin, 1967; Rossi, 1973; Farabegoli et al., 1977). In the M. Pore – M. Cernera area, imbrication suggests that they were transported from SSW to NNE (compare Assereto et al., 1977, fig. 3).

Upper Peres Conglomerates (Upper? Illyrian). The Upper Peres Conglomerates (Fig. 5, column 4 (E)) correspond to the «Obere Peresschichten» of Pia

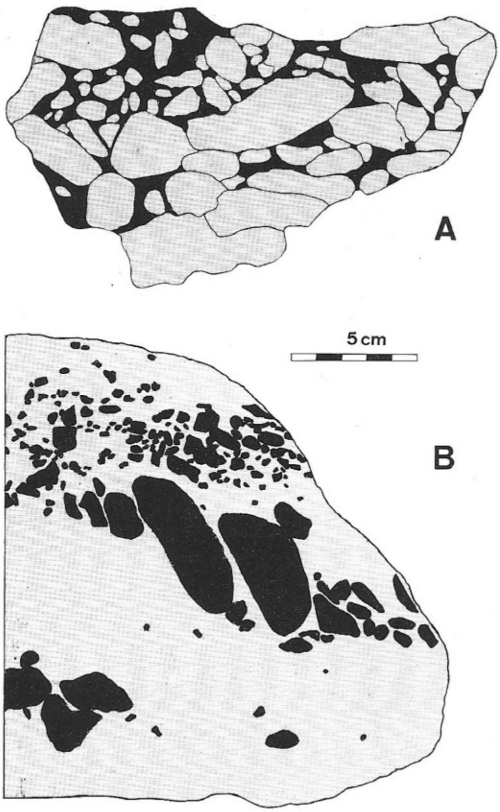


Fig. 8 – Different fabrics of the Peres Conglomerates. A) Fluvial conglomerate (Middle Peres Conglomerates), 200 meters SW of Rio Palata; B) debris flow conglomerate (Upper Peres Conglomerates), Rio Stretto, 2050 m. Current from left to right.

(1937) and Bechstädt & Brandner (1970) and the «Conglomerato di Richthofen» of Pisa et al. (1979).

This unit is composed of paraconglomerates in the lower and fine–grained sandstones and siltstones in the upper part.

The conglomerates crop out as a belt between the southwestern slope of M. Verdal and Rio Stretto (Fig. 3). They are distinguished from the Lower Peres Conglomerates by the high amount of silty–sandy matrix, graded bedding and down–current imbrication of the pebbles (Fig. 8 B). According to Reineck & Singh (1973, p. 126), this fabric indicates rapid deposition. Gravity–flow deposits commonly exhibit similar texture (Nemec et al., 1980). This favours the interpretation that these, about 5 meters thick, conglomerates represent *debris flow* deposits derived from the S to SSW. The frequently rounded pebbles indicate a previous fluvial transport and/or repeated redeposition along a shore.

Intercalated in these mainly red paraconglomerates, which wedge out to the north (Rio Ronc di Caval) and the east (Rio Stretto), is a similar debris flow deposit with components derived from the «calcareous formation». This horizon wedges out to the east within a few tens of meters and probably was transported from W to E.

The overlying sandstones and siltstones are bedded, finely and evenly laminated rocks, which sometimes are burrowed. The prevailing red colour might be a result of the redeposition of Werfen rocks (compare Bechstädt & Brandner, 1970, p. 42) and does not necessarily indicate terrestrial conditions. According to Reineck & Singh (1973, p. 105), these beds can be interpreted as *beach deposits.*

The transition of the «calcareous formation» to the Upper Peres Conglomerates is gradual within some decimeters. Exceptions are those localities where the «calcareous formation» is absent (southwestern slope of M. Verdal, see Fig. 3), or where terrigeneous pollution can be noted in the entire unit (Fig. 6, column 10).

Thickness of the Upper Peres Conglomerates is 15 meters in the M. Verdal area. They wedge out to the north (Rio Zonia) and are limited to the east by a questionable synsedimentary fault (Rio Loschiesuoi).

The depositional environment is characterized by an exposed area in the west and the south, and a shallow marine carbonate area in the east, and may be interpreted as a few kilometers wide channel receiving abundant terrigeneous debris from the south and, subordinately, from the west.

Peres Conglomerates within the Dont Formation. In the M. Verdal – Piz del Corvo area, the Dont Formation frequently contains conglomerate lenses (Fig. 6, column 10, 11). The lenses generally have a lateral extent of some tens of meters and a thickness of several decimeters (Fig. 9). They are composed of well–rounded pebbles derived from the Werfen Formation down to the Val Badia Member embedded in a silty matrix, which frequently contains Echinoderm debris. Channels at the base of the beds have a SSW–NNE direction. These conglomerates can be interpreted as *debris flow* deposits, which were yielded from the south.

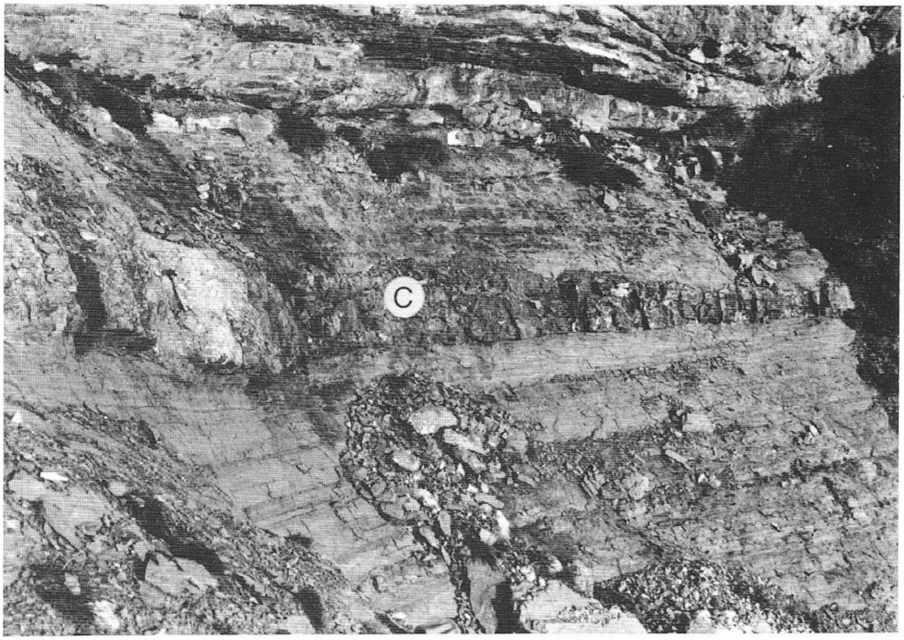


Fig. 9 – Part of a lenticular debris flow paraconglomerate (C), embedded in siltstones of the Dont Formation. Southwestern slope of M. Verdal, 2040 m.

Agordo Formation (Pelsonian). The Agordo Formation (Pisa et al., 1979) (Fig. 5, column 6, 7 (B); Fig. 6, column 9, 10, 11 (B)) is composed of sandstones, siltstones, oncolitic and oolitic limestones, and Dasycladacean and Solenoporacean limestones.

The bulk of the formation is composed of bedded, finely and evenly laminated, sometimes cross–bedded, yellowish–grey sandstones. Siltstones commonly occur in the upper part of the sequence, which differ from the Dont siltstones by the distinct bedding.

Oolitic limestones were observed at the southern base of M. Pore. Grainstone depositional texture prevails. The radial–fibrous oolites are cemented by sparry calcite. Fossil remains are represented by Echinoderm debris and Pelecypod shell fragments.

Oncolitic limestones occur in the Codalonga Valley and are composed of rudstones rich in algal–incrusted lithoclasts.

Solenoporacean limestones occur at the M. Pore. They are rudstones rich in pellets. The Solenoporaceans frequently are totally recrystallized.

Marly Dasycladacean limestones occur east of Rio Giausel, intercalated in decimeter–thick conglomerates with Echinoderm debris.

Thickness is 2 – 25 meters in the M. Pore area, and 0 – 20 meters in the M. Cernera – Piz del Corvo area.

This formation was deposited under *medium to high energetic conditions* (Pisa et al., 1979, p. 73), introducing a new marine sedimentary cycle above the alluvial Middle Peres Conglomerates.

Dont Formation (Pelsonian – Lower Illyrian). The Dont Formation (Pisa et al., 1979) (Fig. 5, column 5, 6, 7 (C); Fig. 6, column 9, 10, 11 (C)) corresponds to the «Pragser Mergelsiltite» of Bechstädt & Brandner (1970). It is characterized by siltstones in the lower and limestones in the upper part.

The calcareous to dolomitic siltstones (sandstones only rarely occur) are yellowish–grey and contain abundant mica, pyrite and coaly plant remains. They are extensively burrowed, and, therefore, the original sedimentary structures, as well as bedding, mostly have been obliterated completely. Ripple marks have been observed at Piz del Corvo.

Fossils are represented by Brachiopods, Gastropods, Pelecypods, Ammonites and Echinoderm debris.

Intercalations are decimeter–thick, dark–grey and commonly graded coquina beds composed of Brachiopod shells and Echinoderm debris in a marly –micritic matrix. These beds frequently exhibit wavy tops and grade upward into siltstones. According to Aigner (1982), they can be interpreted as *proximal tempestites.*

At the southern precipice of Piz del Corvo, a thin lenticular bed of sandy, oolitic limestone has been found. A distinctive feature of this bed is the presence of superficial oolites with nuclei of *volcanic glass and volcanic (?) quartz grains* (Fig. 10).

Due to its stratigraphic position, this bed testifies a *Pelsonian volcanism,* which, however, has not yet been reported from the Dolomites. The occurrence of this volcanism can be explained as a very local phenomenon bounded to synsedimentary tectonics (see chapter "Discussion").

At the southwestern slope of M. Pore, the siltstones display a red colouration. These beds contain coated grains, indicating rather shallow water. Red and grey siltstones interfinger at the southern base of M. Pore. Immediately east of section 7 (Fig. 1), the grey facies exclusively occurs, west of section 6 (Fig. 1) the red facies crops out. In the area between both sections, red and grey siltstones alternate.

The calcareous facies is confined to the Cernera massif and is composed of bedded, sometimes nodular, dark detrital limestones containing silty extraclasts, lumps, pellets, Ammonites, Pelecypod shells and Echinoderm debris. Packstone to rudstone depositional texture prevails.

In the Rio Sacuz area, the limestones commonly are thin–bedded and graded, sometimes exhibiting slumping phenomena. They are characterized by

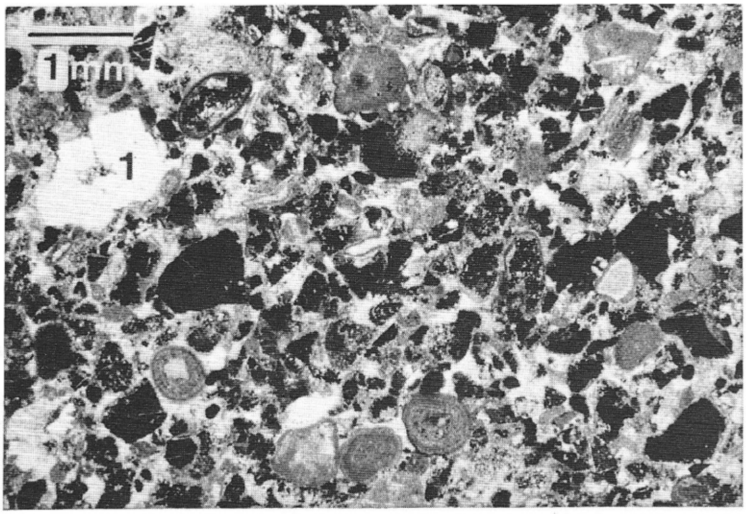


Fig.10 – Oolitic sandstone with volcanic (?) quartz grains (1) and volcanic glass (black). Basal part of the Dont Formation, southwestern precipice of Piz del Corvo, 2120 m. Thin section, Nicols +.

intraclastic packstones with an erosional base grading upward into a wackestone with Radiolaria, thin–shelled Pelecypods and Sponge spicules. To the west, this facies contains meter–sized calcareous blocks derived from the «calcareous formation», bounded by a synsedimentary fault.

The boundary to the «calcareous formation» is sharp in the M. Pore and Piz del Corvo areas and gradual within a few meters at the Cernera massif.

The thickness of the Dont Formation is only 10 meters at the southwestern slope of M. Pore, in contrast to a maximum of 90 meters at the base of Piz del Corvo.

The depositional environment is characterized by *subtidal conditions*. The water depth, however, is difficult to determine due to the absence of depth indicators. At least in the west, it must have been rather shallow.

The Dont patch reefs. The occurrence of patch reefs in the calcareous upper part of the Dont Formation was recognized only recently (Gaetani et al., 1981, p. 32). These patch reefs were found immediately west of Rio Codalonga (100 meters SSE of the 1907 m mark) and in the Col Toront – M. Verdal area (Fig. 6, column 9).

They form mound–shaped bodies with an even base and flancs inclined

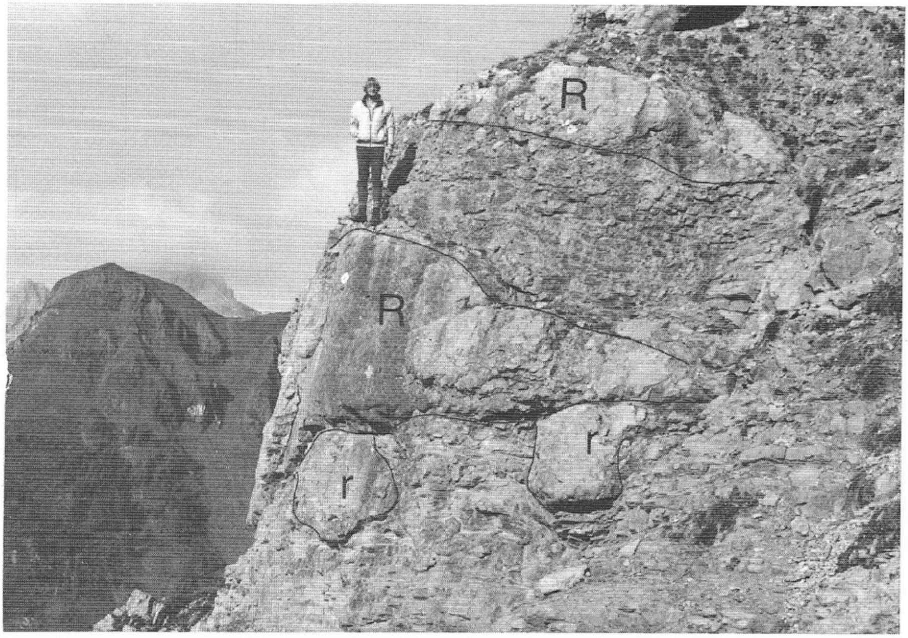


Fig.11 – Patch reefs (R) and gravity–displaced reef boulders (r) in the upper part of the Dont Formation. 200 meters E of Col Toront, 2000 m. Scale 2 meters.

at an angle of about 20°. The lateral extent does not exceed 10 meters, thickness is up to 3 meters (Fig. 11).

The core facies is characterized by Sponge/Algae boundstones (Fig. 12). The reefal organisms are represented by calcareous Sponges (*Peronidella* sp., *Celyphia zoldana* Ott, Pisa & Farabegoli, *Celyphia* sp., *Olangocoelia otti* Bechstädt & Brandner) and Microproblematica, binding a micritic to intrapelmicritic sediment.

The flanc facies is characterized by *Olangocoelia* boundstones (compare Gaetani et al., 1981, fig. 56). *Olangocoelia otti* Bechstädt & Brandner binds an intrapelmicritic to pelmicritic sediment. Up to centimeter–sized primary voids frequently exhibit geopetal fillings.

Fissures, up to several centimeters deep, frequently occur on the reef tops (? shrinkage pores of Gaetani et al., 1981, p. 32), filled with silty sediment.

Laterally, the patch reefs grade into silty detrital limestones containing abundant reef debris.

Commonly, the patch reefs are either superimposed one upon another, separated by detrital, silty limestones, or are situated above gravity–displaced reef boulders (Fig. 11). Slight elevations of the sea bottom, caused by different compaction of sediment, must have favoured the onset of reef growth.

According to Flügel (1982, p. 308), the patch reefs can be interpreted as *reef mounds*. The high energetic facies surrounding them suggests a very shallow depositional area. The reef growth was temporarily inhibited by terrigeneous pollution.

Gravity–displaced reef boulders. In the M. Verdal area, the upper part of the Dont Formation contains exotic calcareous blocks (Fig. 6, column 9), which may reach a size of several meters. Their facies is characterized by *Olangocoelia* boundstones with *Microproblematica* (*Microtubus communis* Flügel, and others), binding intrapelmicritic sediment. The orientation of geopetal fillings in primary voids is mostly different from that of the surrounding sediment (Fig. 13), indicating previous lithification and reflecting the final position of these boulders. Penecontemporaneous deformation of the underlying beds is an ubiquitous phenomenon.

These blocks correspond to the «Rutschblöcke» of the Braies area (Bech-

Fig.12 – Sphinctozoan boundstone from core of lower patch reef of Fig. 11. Thin section, negative print.

städt & Brandner, 1970). They represent rockslide deposits derived from the reefal edge of the contemporarily growing «calcareous formation».

«Calcareous formation» (Lower Illyrian). This informal term comprises shallow water carbonate rocks overlying the Dont Formation («basal member of the Contrin Formation» of Gaetani et al., 1981, p. 32) (D in Fig. 5 & 6).

M. Pore area (Fig. 5, column 5, 6, 7 (D)). The unit starts with red, oolitic limestones (4 meters) containing abundant Echinoderm debris and algal coated shell fragments. The components mostly exhibit hematitic crusts and are cemented by sparry calcite. The upper part (12 meters) is composed of light grey oolitic grainstones to packstones.

A palaeokarst at the top of the unit can be recognized as an irregular relief with height differences up to several decimeters. At a few places, Peres Conglo-

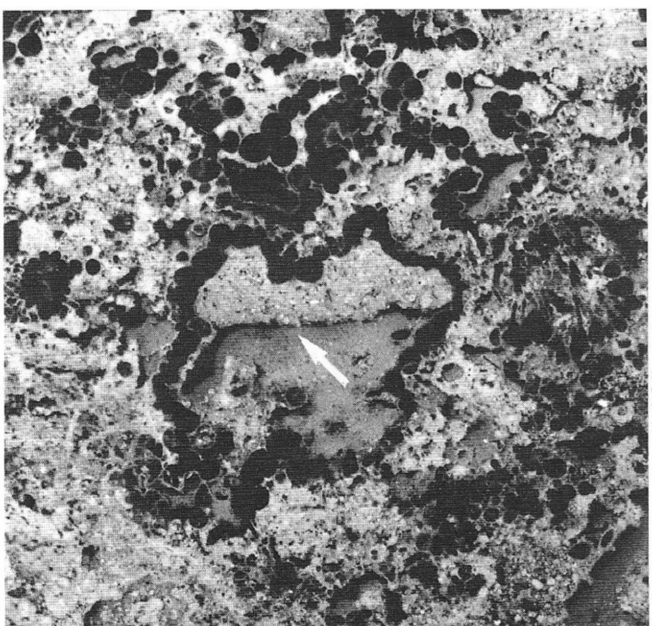


Fig.13 – *Olangocoelia* boundstone with geopetal filling of primary void. Reef boulder of Fig. 11. Up in the field is marked by the arrow. Thin section, negative print.

merates have been found covering the relief. Centimeter–sized solution cavities can be traced several meters down through the underlying limestones, filled with red or green, silty material and capped by blocky calcite in the upper part of the cavities.

Cernera massif (Fig. 6, column 9, 10, 11 (D)). In this area, two horizons can be distinguished. The first one is confined to the Rio Zonia – Col Toront area. This massive (5 meters), dark grey bed is composed of Sponge boundstones overlying intrabioclastic rudstones to grainstones. The reefal organisms (Sphinctozoans, Inozoans, *Microproblematica*) mostly are totally recrystallized. This horizon terminates immediately east of Col Toront, and there it interfingers with siltstones of the Dont Formation. A 5 meters thick sequence of Dont siltites overlies this bed.

Due to its facies and lateral extent of about 500 meters, an interpretation as *biostrome* seems justified.

The second horizon can be traced over the whole area with a thickness of 20 meters. The bulk of this massive, light grey horizon is composed of well–sorted grainstones with abundant lumps, pellets and skeletal grains, which frequently are algal–incrusted (Fig. 14). Sometimes they are rich in reef debris.

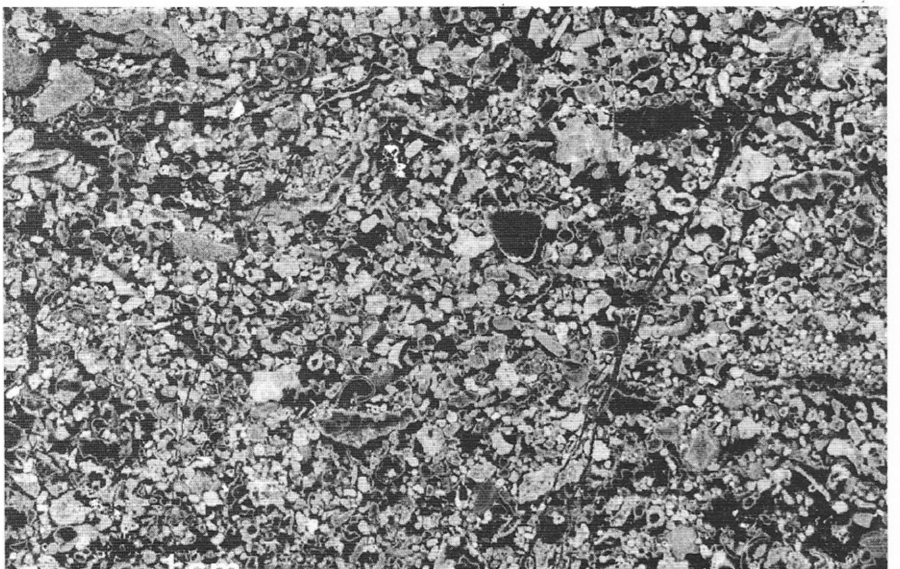


Fig.14 – Grapestone grainstone of the «calcareous formation». 300 meters E of Col Toront, 2020 m. Thin section, negative print.

The reefal facies of this horizon is strongly represented in the area south of Rio Ronc di Caval, exhibiting the same facies that is represented in the exotic boulders of the Dont Formation.

An approximately 10 meters thick, mound–shaped body has been observed at the southwestern precipice of Piz del Corvo. It is intercalated the grapestone facies, totally recrystallized, and might represent a patch reef (Fig. 6, column 11).

At the southwestern base of M. Verdal, this bed is interrupted for about 200 meters (Fig. 3) and replaced laterally by reef boulders embedded in Dont siltites with abundant Werfen pebbles.

In the Rio Giausel – Rio Sacuz area, a palaeokarst horizon constitutes the upper boundary of the «calcareous formation». It can be recognized as a relief with height differences of up to 1.5 meters, which is covered by the basal dolomites of the Ambata Formation.

No sedimentary interruption can be noted in the area between Rio Loschiesuoi and Rio Giausel.

The grapestone facies can be referred to as a *subtidal, restricted carbonate bank* (Flügel, 1978, p. 109), which locally was polluted by terrigeneous supply. Terrigeneous pollution prevented the onset of carbonate production at the southwestern precipice of M. Verdal, and finally obstructed it completely in this zone (Upper Peres Conglomerates).

Ambata Formation (Upper Illyrian). Here, according to the proposal of Pisa et al. (1979, p. 78), the Ambata Formation (Assereto et al., 1977) is considered to constitute the uppermost unit of the Braies Group. As a distinctive feature, it still reflects terrigeneous influence, which has disappeared almost completely in the overlying Livinallongo Formation.

At the base, a 2 meters thick sequence of bituminous, thin–bedded dolomites occurs, which contain abundant pyrite and pyritized or dolomitized Radiolaria. They interfinger with the foreslope talus of the Upper Serla Formation in the Rio Sacuz. This can be noted as a 2 meters thick alternation of calcareous or dolomitic breccias and the above mentioned dolomites.

Then a 25 meters thick sequence of yellowish–grey, evenly laminated and commonly graded silt– to fine grained sandstones, alternating with dark grey marls, follows.

Bedded intercalations are decimeter–thick, graded, detrital limestones with arenitic to ruditic components derived from a shallow water source area. These limestones are locally silicified.

The Ambata Formation represents a *deeper water* deposit with different sources of sediment: a carbonate platform (Upper Serla Formation) yielding proximal *calcareous turbidites* and *debris flows*, and distal *terrigeneous turbidites*, disturbing a marly–clayey background sedimentation.

Upper Serla Formation (Upper Illyrian).

The Upper Serla Formation («Contrin Formation» of Assereto et al., 1977; «Oberer Sarldolomit» of Pia, 1937) starts with the *Morbiac Limestone* (Farabegoli et al., 1977) (Fig. 5, column 6,7 (F)).

At the base, dark stromatolitic limestones occur. Their microfacies is characterized by bindstone depositional texture with laminoid–fenestral fabric. The single laminae are composed of pellets, trapped by recrystallized algal mats. Teepee structures occur in the M. Verdal area. The upper part of the *Morbiac Limestone* is composed of oncolitic and Dasycladacean limestones with *Diplopora annulatissima* Pia.

The *Morbiac Limestone* with a thickness of 6–8 meters is covered by white dolosparites (Fig. 5, column 4, 5, 6, 7, 8 (G)). These dolomites are distinctly bedded (M. Pore area) or coarsely bedded (east of Rio Codalonga). Sedimentary structures have been obliterated by dolomitization. In the M. Pore area, bituminous stromatolitic dolostones constitute the uppermost meter of the formation. They are covered by platy, dark limestones and dolostones (*Plattenkalk Member*) of the Livinallongo Formation.

East of Rio Loschiesuoi, the Upper Serla Formation is composed of calcareous or dolomitic breccias (Fig. 15). It completely wedges out to the east (Rio Sacuz) and forms a *palaeoslope* with an inclination to the basin of about 10°, which is visible in the Rio Sacuz.

The Upper Serla Formation is covered by Sciliar (Schlern) «Dolomite» at the Cernera massif, and by the Livinallongo Formation west of Rio Zonia.

Its facies generally is referred to as a *subtidal carbonate bank* (Gaetani et al., 1981), limited to the east by a basinal area.

Thickness of the Upper Serla Formation is about 40 meters in the M. Pore area, and about 80 meters at the Cernera massif.

Synsedimentary tectonics

Middle Triassic time was a period of considerable tectonic movements in the northwestern Tethys realm (Bechstädt et al., 1978). A mapping of the M. Pore – M. Cernera area at the scale of 1:10000 revealed that the Anisian tectonics of this area can be subdivided into two categories:

– *Synsedimentary faulting.*

Synsedimentary fractures generally have a NNE–SSW direction. They dip vertically or subvertically and exhibit displacements of up to 60 meters. Some faults were reactivated several times during the Anisian (and the Ladinian). Oblique slip along the fault planes sometimes can be recognized. Anisian faulting was confined to two main phases. The first lasted from the deposition of the Lower Serla Formation to the deposition of the Middle Peres Conglomerates. The second was active at the end of deposition of the Dont Formation until the sedimentation of the Upper Serla Formation. The synsedimentary fractures generally were not reactivated during the Alpidic orogeny, except for the fault in the extreme west of the investigated area (Fig. 16), which now forms a reverse fault.

– *Differences of subsidence without noticeable faulting.*

In the Uppermost Scythian and Lower Anisian, differences of subsidence caused gentle flexures of the substratum. A similar development can be noted in the Pelsonian. During these phases of relative tectonic quiescence, mainly

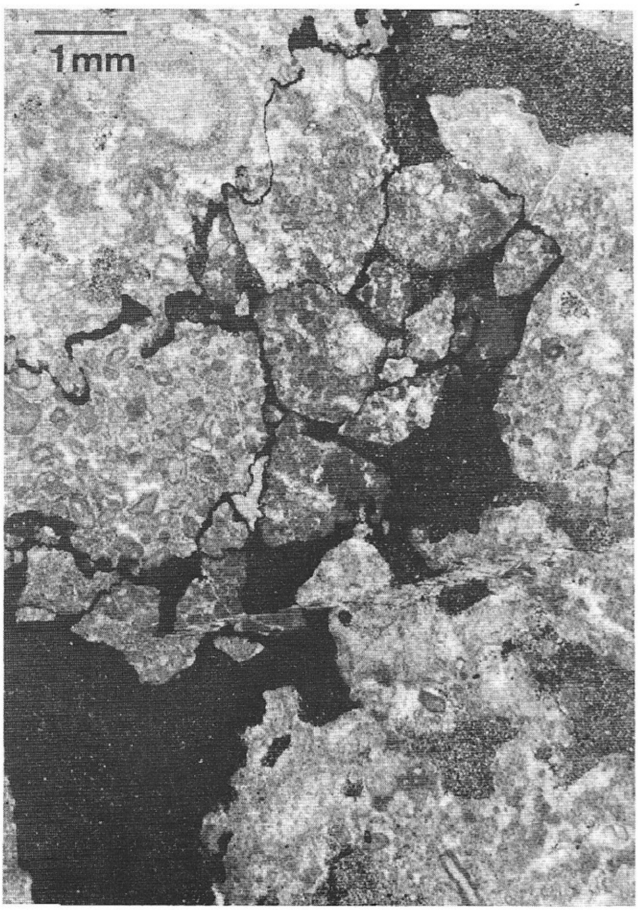


Fig.15 – Breccia from foreslope talus of the Upper Serla Formation with marly, micritic matrix. Note that the components are composed of strongly recrystallized grainstones, and stylolitic contacts. Thin section.

fine clastic sedimentation took place in contrast to the phases with strong movements, which generally were associated with coarse clastic sedimentation (Peres Conglomerates).

Both types of tectonic movements occurred during the deposition of the Upper Serla Formation.

Palaeostructural history

First setting of the «Dorsale badioto–gardenese» in the Upper Scythian ? Lower Anisian.

During the Upper Scythian, the previously terrigeneous sedimentation of the Werfen Formation was replaced by the chiefly calcareous deposits of the Val Badia Member. Differences of thickness of this member occur at a regional scale (80 – 100 meters in the type area;(Bosellini, 1968, p. 8), about 60 meters at the Cernera massif, about 70 meters in the northeastern Dolomites (Casati et al., 1982, fig. 8). The carbonate sedimentation was abruptly replaced by the onset of a minor cycle with prevailing terrigeneous sedimentation.

A facies section across the investigated area (Fig. 16 A) demonstrates that both the Cencenighe Member and the San Lucano Member thin out considerably to the west. For the first time the influence of a ridge zone is testified in the investigated area. This ridge zone has been called «Dorsale badioto – gardenese» by Bosellini (1964, p. 210).

The «Dorsale badioto – gardenese» («Gadertal – Grödner Tal – Schwelle») constitutes an asymmetrical, some tens of kilometers wide anticline with an axial direction running approximately N–S. Its western palaeoflanc is a rather gentle slope, while the eastern margin, repeatedly influencing the sedimentation of the investigated area, is characterized by a steep escarpment (Bosellini, 1968, p. 27). Brandner & Mostler (1982, p. 31) referred to its onset as a tilting movement of the pre–Triassic basement to the west and southwest.

The reduced sedimentation of the upper Werfen Formation in the M. Pore area may be interpreted as a complete wedging out of these deposits further west. However, this possibility remains speculative due to strong Anisian erosion on the «Dorsale», which even involved Permian beds at some places.

It can be presumed, that the San Lucano facies spread from the west to the east. A similar relationship between the Cencenighe and San Lucano Members has been reported from the Agordino (Farabegoli et al., 1977). The regressive tendency (Brandner & Mostler, 1982, p. 25) culminated with a short emersion phase after the deposition of the San Lucano Member.

Carbonate bank sedimentation in the Lower Anisian.

In the Lower Anisian, the sea transgressed the M. Pore – M. Cernera area anew. The terrigeneous influx had vanished and a monotonous tidal flat facies of the Lower Serla Formation was deposited.

The absence of a coarse clastic facies near those zones, where the formation has not been preserved (Rio Codalonga, west of M. Pore) suggests that this carbonate bank developed in the whole area and covered at least marginal parts of the «Dorsale».

During the deposition of the Lower Serla Formation synsedimentary faulting set on and caused different thickness of the sequence in the Piz del Corvo area (Fig. 16 B).

Collapse of the carbonate bank and uplift of the «Dorsale badioto–gardenese».

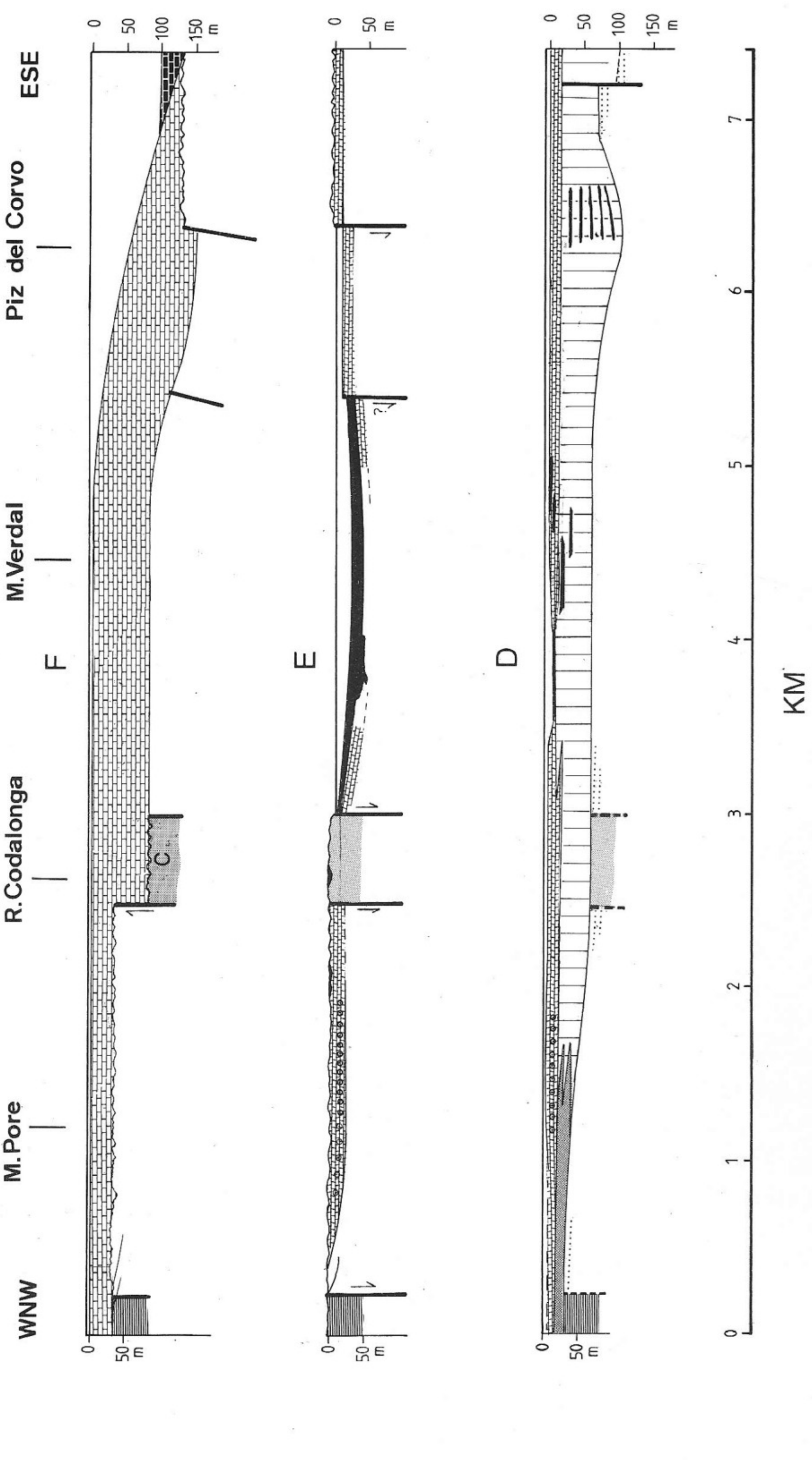
A tectonic phase abruptly stopped the carbonate sedimentation and disintegrated the Lower Serla bank by block faulting, forming horsts and grabens (Fig. 16 C). Contemporarily, an area immediately south emerged and yielded huge masses of coarse clastic material (Fig. 17). Streams deposited these conglomerates in the graben zones and sometimes deeply cut into the substratum (Fig. 7). Additionally, rocks were eroded from the horst of the Codalonga Valley (about 55 meters: San Lucano Member ca. 30 meters, Lower Serla Formation ca. 25 meters) and from the rising «Dorsale», where up to 400 (!) meters were eroded (Bosellini, 1968, p. 31) further west and northwest.

The comparatively reduced thickness of the conglomerates in the area considered here may be explained by a transport of debris from E to W on the «Dorsale» caused by the tilting movement of the whole ridge structure (Brandner & Mostler, 1982, p. 31).

After the deposition of the Middle Peres Conglomerates («Richthofen Conglomerate» on the ridge) the sea once again transgressed and reduced the relief differences by depositing the shallow water sediments of the Agordo Formation. The finding of Werfen pebbles in the vicinity of the structural rises suggests that the Agordo Formation did not cover theses areas.

Basinal sedimentation in the Pelsonian and Lower Illyrian.

A continuing rise of sea level replaced the high energetic near–shore facies of the Agordo Formation by the silty–calcareous low energetic sediments of the Dont Formation. These siltites were probably derived from the north (Farabegoli & Guasti, 1980, p. 920), which is suggested by the commonly greater thickness of these deposits further north (Braies and Valdora area up to 200 meters according to Bechstädt & Brandner, 1970). Sporadically, the deposition of the silty material was disturbed by storms (coquina beds) and gravity–flow deposits from the southern emerged area.


WNW
M.Pore
R.Codalonga
M.Verdal
Piz del Corvo
ESE
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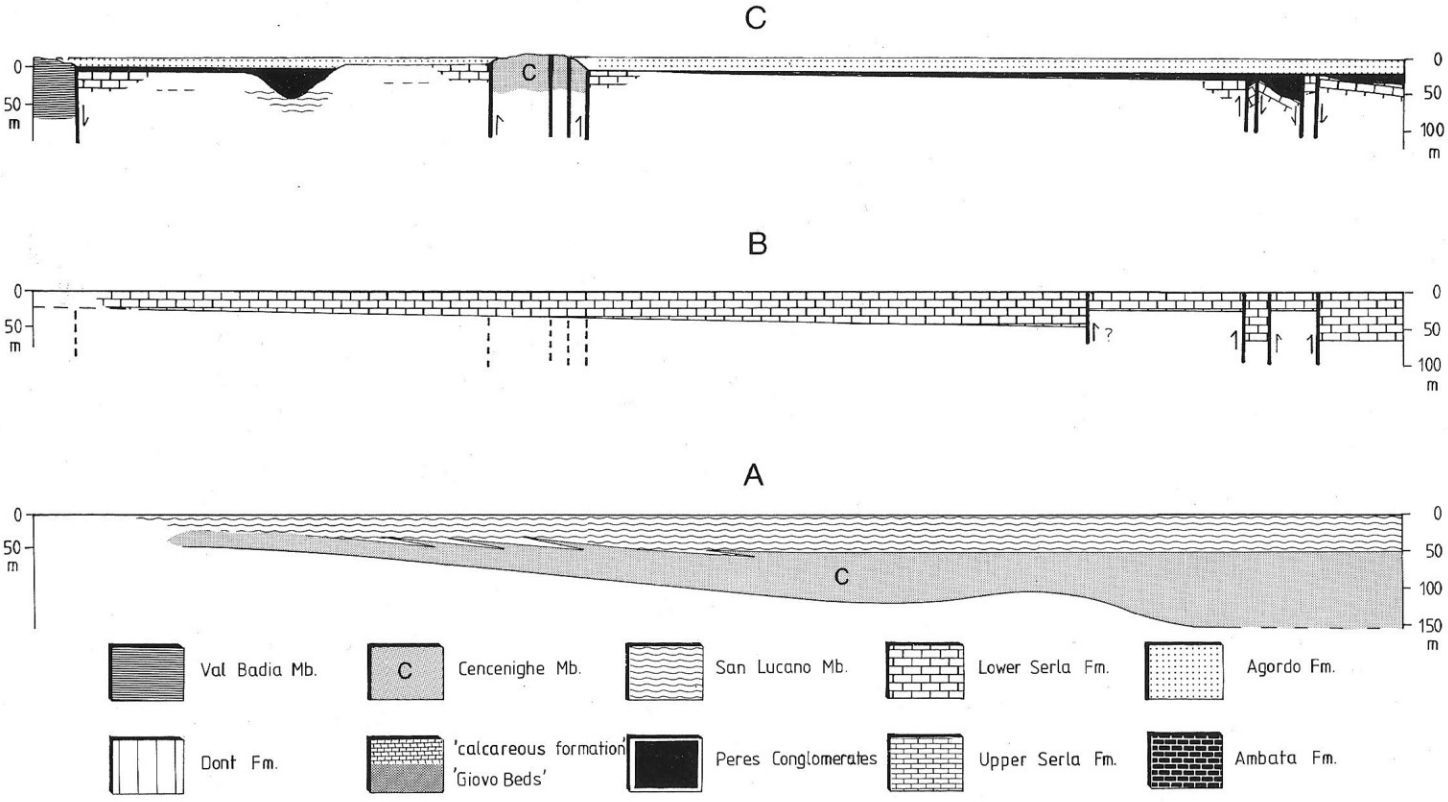

Fig.16 – Facies profiles across the M. Pore–M. Cernera area. A) Upper Scythian ? Lower Anisian; B) Lower Anisian; C) Lower Pelsonian; D) Pelsonian – Lower Illyrian; E) basal? Upper Illyrian; F) Upper Illyrian.

Synsedimentary movements of the subsurfaces caused gentle flexures and, therefore, different thickness of the silty Dont deposits.

The structural rise in the west, the northeastern part of the «Dorsale», seems to have been covered by the sea as well. This interpretation is supported by the lack of a coarse clastic facies in the western part of the investigated area. Additionally, red silty/sandy deposits similar to the red facies of the Dont Formation occur intercalated between a conglomerate at the base («Richthofen Conglomerate» respectively Middle Peres Conglomerate) and on the top (?Upper Peres Conglomerates) in the Livinallongo area (Bosellini, 1968, p. 31) and the M. Forca area (Nöth, 1929, p. 140). Furthermore, Brandner & Mostler (1982, p. 31) recently reported that a silty–sandy–calcareous, chiefly greenish-grey sequence, which they called «Giovo Beds», overlies the Middle Peres Conglomerates on most parts of the «Dorsale».

The red facies of the Dont Formation might represent a red outlier of these «Giovo Beds». Interpretatively, the term «Giovo Beds» was adopted for this sequence in Fig. 16 D.

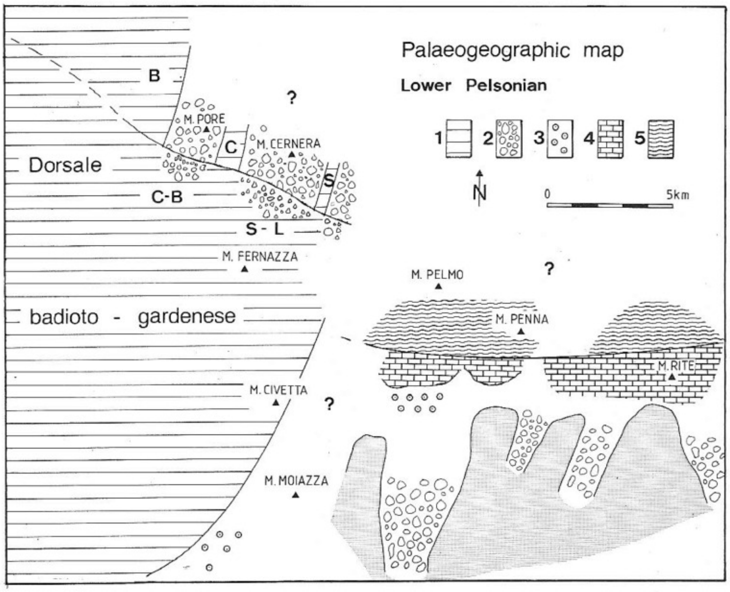

Fig.17 – Palaeogeography of the Southeastern Dolomites during the deposition of the Middle Peres Conglomerates. 1) Emerged area: B) Val Badia Member; C) Cencenighe Member; L) San Lucano Member; S) Lower Serla Formation; 2) conglomerates; 3) oolites; 4) platform carbonates; 5) basinal sediments. Modified from Pisa et al. (1979, fig. 16).

Carbonate bank sedimentation in the Lower Illyrian.

A relative fall of sea level in the Lower Illyrian caused the onset of new areas with shallow water carbonate sedimentation, which were limited to the marginal parts of the «Dorsale». This development probably set on in the M. Pore area with oolitic limestones. The remaining basinal area east of Rio Codalonga continued subsiding, probably along a fault. The faulting activity prevented further progradation of the carbonate bank («calcareous formation»), whose reefy rim (Fig. 18 A), which is either not exposed (between M. Pore and Rio Codalonga) or has been eroded (Rio Codalonga), yielded exotic blocks and calcareous debris.

A biostrome developed at the western basin margin and was halted by terrigeneous pollution. Further basinward, patch reefs settled in this environment.

With the shallowing of the basin, grapestone sedimentation set on in the M. Verdal – Piz del Corvo area (Fig. 18 B). East of Piz del Corvo, the eastern margin of this environment was caused by a short–lived fault, along which breccias slid into the basin. With the cease of faulting, the carbonate bank could prograde, even over this remnant of basinal area.

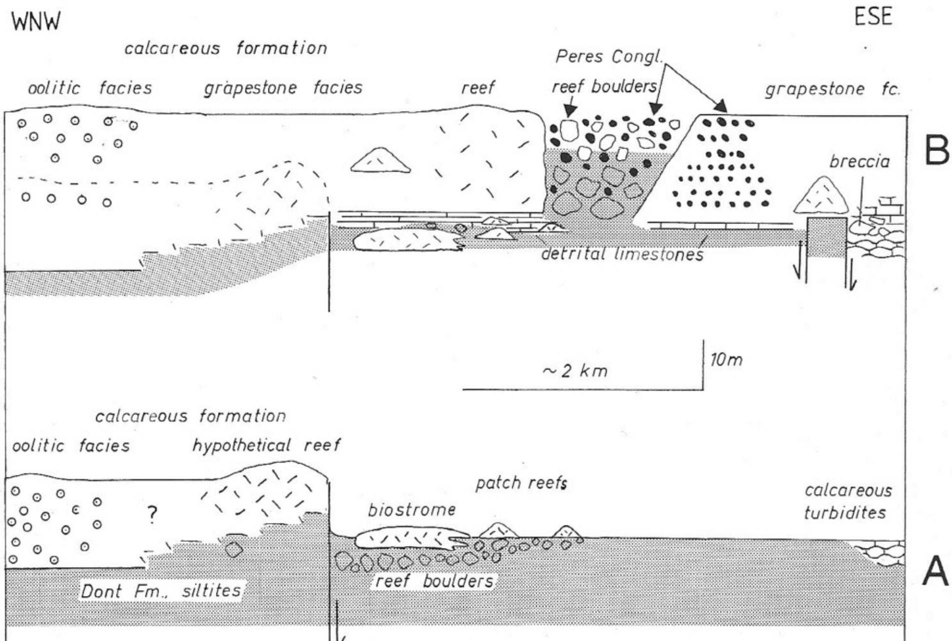

Fig.18 – Relationships between the «calcareous formation» and the Dont Formation.

A new uplift of the southern hinterland caused the reactivation of its drainage system. Consequently, the Rio Stretto area was polluted by terrigeneous debris. Here, clastic pollution is considered to have been responsible for the interruption of the carbonate bank in the M. Verdal area, as there is no evidence of synsedimentary faulting.

Second uplift of the «Dorsale badioto–gardenese» in the Upper? Illyrian.

The tectonic phase, which set on during the deposition of the upper part of the Dont Formation, culminated in the (?) basal part of the Upper Illyrian. According to Brandner & Mostler (1982, p. 32), only the eastern margin of the «Dorsale» was uplifted and eroded anew.

In the surroundings of the investigated area, erosion was probably active in a narrow strip between Pieve di Livinallongo (west of this locality the «Giovo Beds» seem to be preserved) and the Rio Codalonga. The fault west of M. Pore was reactivated (Fig. 16 E), causing normal drag flexure of the beds of the eastern block.

On the western block, the «Giovo Beds» (and questionable outliers of the «calcareous formation») and parts of the Val Badia Member were eroded.

The little amount of erosional debris of these beds in the Upper Peres Conglomerates of the M. Verdal area suggests only slight erosion in the M. Pore area.

Contemporaneously, the southern area (Fig. 19) again yielded coarse clastic material, which suddenly arrested the growth of the «calcareous formation» at the Cernera massif.

Resumption of carbonate bank sedimentation and basinal development in the east.

The cease of terrigeneous pollution was accompanied by a rise of sea level. The carbonate sedimentation commenced anew with the lagoonal Morbiac Limestone. The occurrence of similar deposits west of the investigated area (Masetti & Neri, 1980; Ogilvie–Gordon, 1929, p. 364) suggests a regional distribution of the Morbiac facies.

The western Codalonga fault was reactivated once more, separating a slowly subsiding area in the west from a rapidly subsiding area in the east (Fig. 16 F). A flexure between Rio Loschiesuoi and Piz del Corvo, exhibiting a dislocation of 70 meters, caused the inundation of the eastern area. Starved basinal deposits of the Ambata Formation covered the palaeokarst relief of the Rio Sacuz area.

The carbonate bank of the Upper Serla Formation yielded a brecciated wedge («Überguss–Schichten» = foreslope talus deposits) and calcareous debris into the basinal area.

Finally, at the Anisian – Ladinian boundary, the Upper Serla bank was deeply fractured at a regional scale (Brandner & Mostler, 1982, p.32). At the Cernera massif, the area west of Rio Codalonga collapsed and was covered by the starved basin (Bosellini & Ferri, 1980) *Plattenkalk Member* of the Livinallongo Formation.

The Cernera massif acted as a topographic high where platform sedimentation continued (Sciliar «Dolomite» Formation), and constituted a part of a S–N striking carbonate platform which probably extended from the Pale di San Martino in the south to the Braies area in the north (Blendinger et al., in press).

Discussion

The Anisian sedimentation pattern of the M. Pore – M. Cernera area is characterized by the following features:

- extreme lateral facies changes;
- great thickness of rapidly deposited sediment;
- sediment supply from different sources;
- several unconformities.

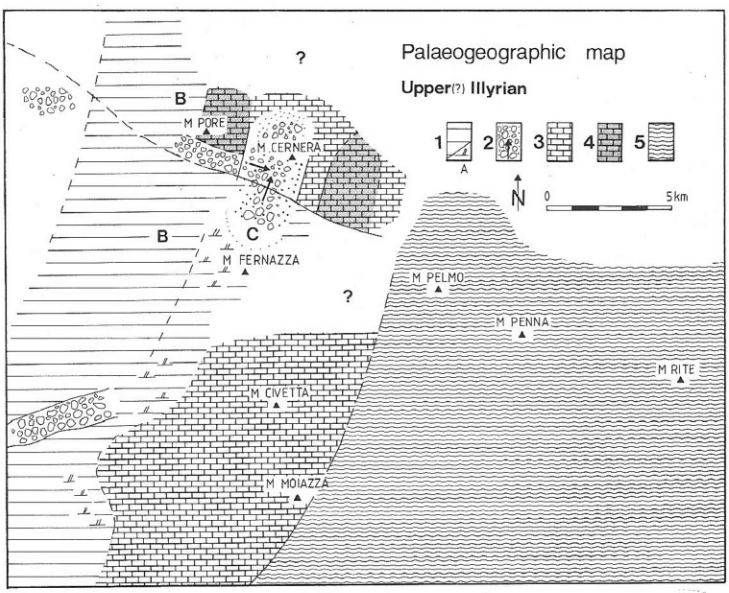

Fig.19 – Palaeogeography of the Southeastern Dolomites during the deposition of the Upper Peres Conglomerates. 1) Emerged area (for the letters see Fig. 17); A) swamps; 2) conglomerates; 3) platform carbonates; 4) emerged carbonate platform; 5) basinal sediments. Modified from Pisa et al. (1979, fig. 19).

The Anisian tectonic of the investigated area displays:

– coexistence of a compressional structure («Dorsale badioto – gardenese», compare Marinelli et al., 1980, p. 6), and a sedimentation basin characterized by extensive syndepositional fracturing;

– vertical to subvertical faults in the sedimentation area;

– horsts forming to grabens.

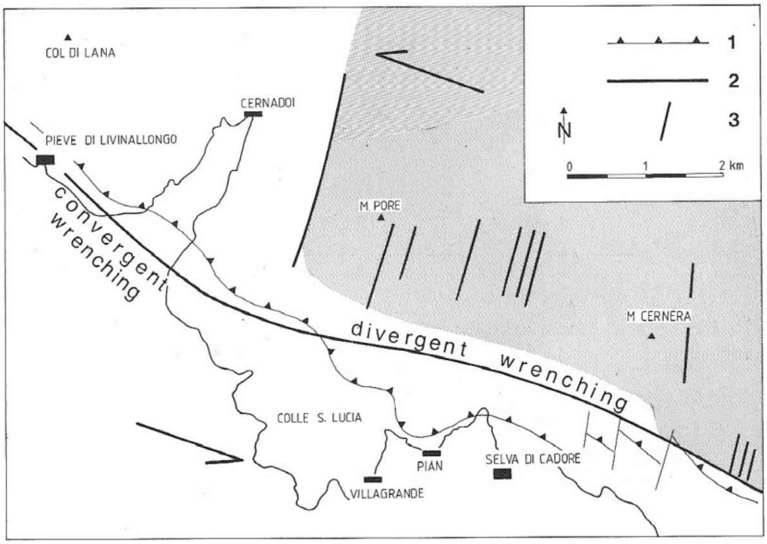

Fig. 20 A – Structural pattern of the investigated area and its closer surroundings. 1) actual (Alpidic) thrust fault (Cernera Line, Linea di Selva); 2) interpretative course of the Anisian wrench fault; 3) Anisian fractures. Structural rises are white.

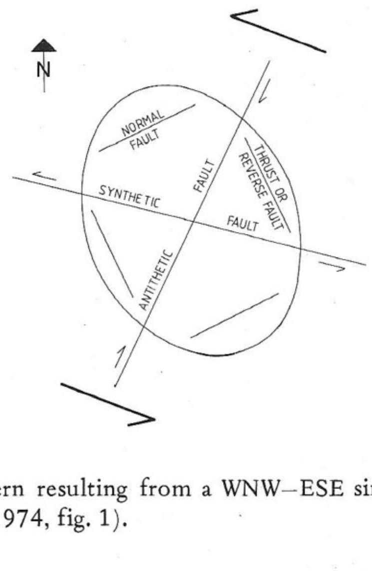

Fig. 20 B – Structural pattern resulting from a WNW–ESE sinistral shear couple. Modified from Harding (1974, fig. 1).

According to Reading (1982, p. 335), the sedimentation pattern of the investigated area can be regarded as being typical for a strike–slip generated depositional area.

The palaeogeographical reconstructions of Fig. 17 and 19 encourage the assumption that a long–lived fault or fault system was located immediately south of the investigated area. This fault confined an emerged land to the north, which yielded coarse clastic debris from the Pelsonian to the Illyrian. Interpretatively, this fault is considered a wrench fault (Fig. 20 A).

Such wrench faults are seldom straight and tend to curve (Harland, 1971). Therefore, they may cause compressional structures whenever the block movement along the fault converges («convergent» or «transpressive» wrenching) and tensional structures in those zones where the block movement diverges («divergent» or «transtensile» wrenching).

Vertical displacements of the blocks along the wrench fault are a ubiquitous phenomenon, frequently constituting the only provable fault motion (Reading, 1982, p. 335).

Considering the slightly idealized course of the fault in the south of the investigated area (south of which the Braies Group is almost completely absent) (Fig. 20 A), by superimposing a WNW–ESE sinistral shear couple (Fig. 20 B), it becomes evident that a sinistral wrench zone with convergent wrenching chiefly west of the M. Pore area and divergent wrenching in the M. Pore – M. Cernera area results.

The synsedimentary, NNE striking fractures can be interpreted as antithetic faults. According to Wilcox et al. (1973, p. 87) "*antithetic fractures inherit some of the tensional component of a wrench deformation and commonly become nearly vertical normal faults with neglegible lateral displacements*".

Normal faults, as well as thrusts or reverse faults, which should be expected from the fault pattern of Fig. 20 B, were not recognized in the investigated area. The former "*... may form along a wrench zone in the initial stage of deformation, but they are easily destroyed as wrench displacement increases...*" (Wilcox et al., 1973, p. 87).

It is not possible, however, to determine the Anisian amount of horizontal displacement along this wrench fault. Post–Anisian reactivations (Ladinian, Alpidic orogeny) are very likely and, therefore, concealed the Anisian displacement.

In this context, uplift and erosion of the «Dorsale badioto – gardenese» might be explained, at least locally, as a consequence of convergent wrenching. Divergent wrenching in the M. Pore – M. Cernera area formed a sedimentation area which was temporarily deeply fractured.

The *Pelsonian volcanism*, whose traces are testified in the lower part of the Dont Formation, can be interpreted as being associated with the *distensive* wrench zone. Volcanic dikes have been reported from these wrench zones (Bishop, 1968; Wilcox et al., 1973).

According to this working hypothesis, two main wrench phases can be noted in the Anisian (Lower Pelsonian, basal? Upper Illyrian). Each tectonic phase is followed by a rise of sea level and may be connected with rifting activities in the Tethys sea.

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