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COMPARISON OF SLOPE AND BASINAL SEDIMENTS OF A MARGINAL CRATONIC BASIN (PEDREGOSA BASIN, NEW MEXICO) AND A MARGINAL GEOSYNCLINAL BASIN (SOUTHERN BORDER OF PIEMONTAIS GEOSYNCLINE, BERNINA NAPPE, SWITZERLAND)

by

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A THESIS SUBMITTED
IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF

Doctor of Philosophy

Thesis Director's signature:

James Lee Wilson

Houston, Texas

July, 1973

ABSTRACT

COMPARISON OF SLOPE AND BASINAL SEDIMENTS OF A MARGINAL CRATONIC BASIN (PEDREGOSA BASIN, NEW MEXICO) AND A MARGINAL GEOSYNCLINAL BASIN (SOUTHERN BORDER OF PIEMONTAIS GEOSYNCLINE, BERNINA NAPPE, SWITZERLAND)

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The transition from slope to basinal facies is well exposed in the Pennsylvanian of New Mexico's Pedregosa Basin and in the Liassic of Switzerland's Piemontais Geosyncline, and this research focused on comparing these slope and basin sediments deposited in two tectonically different environments.

The shape of the Pedregosa Basin is controlled by basement faulting; its trend is oblique to the Ouachita-Marathon Geosyncline. During periods of stabilized high sea-level stands, carbonate sedimentation prevailed in shelf areas and bioherms built up at the shelf edge. Bioclastic foresets and huge slumps of brecciated material from these bioherms extended from the shelf edge into the basin where deep-water carbonates were deposited. During periods of low sea-level stand, shallow water channels transported clastics over the shelf and eroded 400 foot deep channels on the slope. The basin was filled mainly with shales, and channels on the slope were filled with

sand. At the base of channels on the slope, there are probably fan- or cone-shaped sand bodies.

Normal faults in the Liassic slope sediments of the Piemontais Geosyncline reflect the structural tension that persisted during early geosynclinal stages. Thick beds of structureless breccias occur on the downthrown side of these faults. In the downslope direction, the breccias grade into turbidite facies; farther downslope, thick wedges of turbidite thin at their distal edges. No evidence was found of channels or canyons that would serve as point sources for sediment dispersion down the slope. The turbidites grade upward into radiolarian chert; this subsequent decrease of detrital influx is attributed to the general deepening of the geosyncline.

The main difference between the two basins is, that the marginal cratonic basin deepened during carbonate sedimentation and became filled during clastic sedimentation, which is in contrast to the marginal geosynclinal basin which never filled and had only one set of sedimentation type breccias - turbidites. The sediments of the latter basin grade upwards into radiolarian chert. The sediments of the marginal cratonic basin are covered by continental deposits.

CONTENTS

INTRODUCTION

The thesis area in Switzerland is an outgrowth of my Master's thesis area, which was completed in 1969 and consisted of mapping, description and interpretation of the Piz Alv area. The much larger and more complicated Piz Mezzaun area (north of the Piz Alv area) was studied for the Ph.D. work. Half of the time was devoted to an understanding of the structure, the other half to sedimentological observations. Short studies were conducted in the Valle del Monte area (Italy). The Piz Alv area was restudied. The fieldwork in this high alpine area was done in the summers of 1969 and 1970.

Fieldwork in the Big Hatchets was done in the summer of 1971. The logistics in this desolate desert area were difficult compared to the well populated and cool Swiss area in the high Alps. My field assistant (D. E. Mellor) and myself stayed in a camp which was frequently moved around in the Big Hatchet area. The daily problems consisted of flat tires, lack of water and rattlesnakes. The interesting ibex population of the Engadin Valley was replaced by the even more interesting mountain lions.

Two areas with well exposed shelf edge and slope sediments were found in the Big Hatchets. These two areas were studied in detail and the beds were walked out over distances of 1 to 2 miles. The measured section of these two areas could be correlated with a shelf and basin section (Exxon well). The area in Switzerland was so tectonically complicated that beds could not be walked out; lateral control over distances of 1 to 2 miles is therefore lacking. Finding complete and representative sections was another problem. The control sections were measured in detail. Only mapping directed towards a structural understanding was done. The structural interpretations will be published separately from this thesis.

The area in Switzerland was mapped by the late R. Staub. His map could be used as a base. However, errors and the small scale did not allow use of this map for any detailed studies necessary for this thesis. The general geology of the Big Hatchet area was worked out by the late R. Zeller. His map and publications could be easily used.

The comparison of the basinal and slope sediments of these two megatectonic environments should bear some results towards an understanding of the development of different shelf, or continental edges, as well as a more precise prediction of different sedimentary settings, their facies types and distribution and their reservoir and source rocks.

ACKNOWLEDGEMENTS

Rice University made it possible to execute this internationally-oriented thesis. A fellowship was given to me by the Department of Geology of this university. The Bureau of Mines and Mineral Resources of New Mexico granted the money necessary for the fieldwork in the Big Hatchets and the laboratory work in Houston, and provided a Jeep to use during the fieldwork. The Geological Institute of the Swiss Federal Institute of Technology provided technical staff and instrumentation to prepare the thin sections of the Swiss field area.

Don BAKER authorized

I am most thankful to Dr. J. L. Wilson who served as my thesis director. He spent much time with me in the field and many hours behind the microscope. His ideas of geology were the most important factor in my geological education. I am also thankful to Dr. B. C. Burchfiel (thesis advisor) who introduced me to the megatectonic aspects of my thesis, to Dr. R. Truempy (thesis advisor) who was a great help in my Swiss field area and Dr. J. R. Warme (thesis advisor) for his help.

Frank Kottlowski (assistant director of the Bureau of Mines and Mineral Resources of New Mexico) made it possible to realize the fieldwork in the Big Hatchets. Garner Wilde (Exxon Co., Houston, Tex.) identified the fusulinids of the Big Hatchet area. Dr. W. King (University of New Mexico, Las Cruces) identified some fusulinids during the fieldwork. S. Thompson (Esso Production Research Co., Houston), who worked in the Big Hatchet when Exxon (then Humble Oil and Refining Co.) drilled their well in this area, provided much information and many good ideas. Ch. Campbell (Esso Production Research Co., Houston) helped interpret the sandstones in the Big Hatchets. J. Tovar (Petroleos Mexicanos) showed us the late Paleozoic sediments in the area of Ascension (Chihuahua) on a very interesting field trip. C. Daetwyler, J. Lowell, Ch. Mansfield and M. Ziegler (all at Esso Production Research Co., Houston) helped a lot with their interesting and critical discussions and comments.

I also thank Dr. K. J. Hsü and Dr. A. Gansser of the Swiss Federal Institute of Technology. Field assistant in New Mexico was D. E. Mellor, an undergraduate from Rice University. He was a good help in the field and back in the laboratory.

Cuttings, logs and core chips of the Exxon well west of the Big Hatchets were provided by Exxon Co., Midland, Tex. and the Bureau of Mines and Mineral Resources in New Mexico.

Special thanks for helping in the field areas are given to: Mr. and Mrs. Everhart (east side of the Big Hatchets), Mr. and Mrs. Freeman (west side of the Big Hatchets) and Mrs. B. Rageth (La Punt-Chamues-ch, Engadin Valley).

The writer is extremely thankful to his I.I.E. host family in Houston: Mr. and Mrs. Turner.

This thesis is dedicated to my parents.

STATEMENT OF PROBLEM

Basinal and slope sediments are studied in two different megatectonic environments. These environments are:

(a) marginal cratonic basin. Permo-Pennsylvanian sediments of the Pedregosa Basin in SW New Mexico.

(b) marginal geosynclinal basin. Southern border of Piemontais Geosyncline; Bernina nappe, Jurassic sediments; Engadin Valley, Switzerland.

The study of each basin has the following goals:

(a) facies types, shapes of sedimentary bodies, and their distribution through time.

(b) interpretation of sedimentary processes.

(c) development of the basin; i.e. formation, geometry and filling of the basin.

Reasons for comparing and contrasting these two basins include:

(a) an attempt to understand the development of continental edges and the history of the related basins in different megatectonic environments.

(b) cause and differences of subsidence in different megatectonic environments.

(c) to be able to characterize each type of basin sufficiently to permit recognition with limited data.

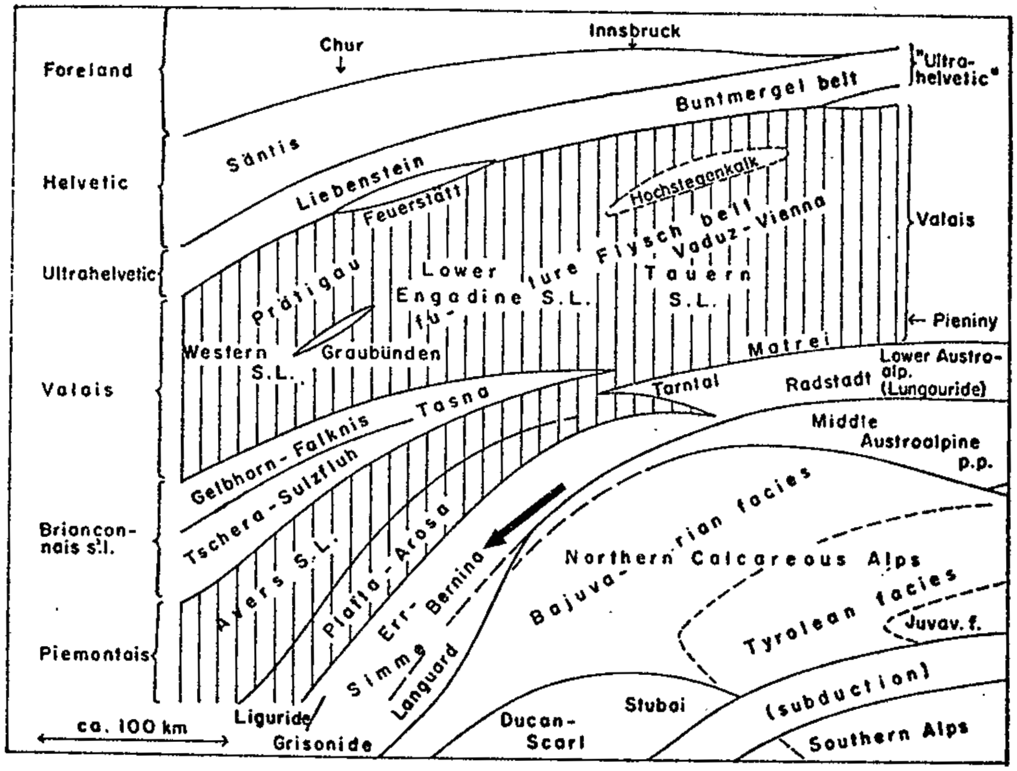
(d) to be able to predict distribution of special facies types, e.g. possible reservoir and source rocks for petroleum.

EXAMPLE OF A MARGINAL GEOSYNCLINAL BASIN

(Jurassic Sediments of the Bernina Nappe, Engadin Valley, S. E. Switzerland)

(a) INTRODUCTION

The Jurassic sediments of the Bernina nappe were chosen to illustrate an example of a marginal geosynclinal basin. The Bernina nappe occurs east of the Engadin Valley in southeastern Switzerland and lies tectonically on the Platta-Arosa nappe which contains eugeosynclinal sediments (radiolarian chert, deep water clastics, etc.) and ophiolites. An oceanic crust for this nappe seems to be very probable and is assumed for this study. The Platta-Arosa nappe is the southernmost part of the Piemontais Geosyncline (Figure 1). The Bernina nappe with its thick crystalline core has a continental crust and belongs to the continental block which lies south of the Piemontais Geosyncline. The Bernina nappe belongs to the Lower East Alpine realm and is the northern edge of the continental block represented by the East Alpine nappes and the South Alps. In summary, the Bernina nappe is the transition between the continental block



From: R.Truempy,1971.

Figure 1: Very hypothetical palinspastic paleogeo= graphic map of the Alps in eastern Switzerland and western Austria during late Jurassic and early Cretaceous time.

The vertical signature marks the eugeosynclines, which had partly an oceanic crust. The non-ruled area are continental blocks.

The arrow indicates the thesis area,which is at the northern edge of the southern continent.

to the south and Piemontais Geosyncline (with oceanic crust) to the north.

(b) GEOLOGICAL SETTING

The Bernina nappe occurs on the east side of the Engadin Valley (Plate 1). The sedimentary cover is preserved in the Piz Alv, the Valle del Monte area and the Piz Sassalbo areas. The Piz Alv area was studied for my diploma thesis (M. A. Schüpbach, 1969) and was restudied for this work. The Valle del Monte area has been only studied for the Ph.D. thesis. The Piz Sassalbo area, the most southern one, has not been studied. The largest area studied in connection with this work is the Piz Mezzaun area which lies to the north of the other areas (Plate 1). The sediments here occur in a half window near Zuoz. The window exposes eight tectonic slices, disconnected from the main crystalline core of the Bernina nappe and transported northward together with the advance of the Languard nappe. The latter is the next tectonically higher nappe above the Bernina nappe.

The crystalline core of the Bernina nappe consists of various Hercynian metamorphic and igneous rocks. The

Triassic sediments are mostly dolomites, besides some few cornicules and limestones. Several tuff horizons occur throughout the section. The depositional environment consisted of inter to supratidal and shallow marine conditions.

The Triassic facies types are the same in the whole realm of the Bernina nappe and are usually similar to the facies of the nappes which tectonically over- or underlie the nappe. A wide spread shallow sea on a continental crust evidently persisted through Triassic time in this area.

The facies types of the Jurassic sediments vary tremendously within and between the different study areas of the Bernina nappe. There are proximal ("near shore") sediments in the southern areas (Piz Alv, Valle del Monte) and distal sediments in the northern area (Piz Mezzaun). This facies distribution reflects the general Jurassic paleogeography of a continental block to the south and the Piemontais Geosyncline with oceanic crust to the north (Figure 1). A thin formation of Cretaceous sediments covers the Jurassic sediments.

(c) THE JURASSIC SEDIMENTS OF THE PIZ ALV AREA

Four different formations could be distinguished in the

Piz Alv area. Most of the formations extend also in other areas.

(1) Untere Kieselige Kalke Formation (Lower Siliceous Limestones)

This formation name is informal. It was introduced by the writer in 1970. Still today it is too early to introduce formal names for all formations, because of insufficient study of the sediments of the Lower East Alpine nappes.

It is possible that the "untere kieselige Kalke" are equivalent to the Agnelli formation on the Julierpass (W. Finger, 1972).

The "untere kieselige Kalke" have a fixed stratigraphic position, which is always directly above Rhaetian limestone and below the tectonic sediments (breccias and turbidites). The 7 meters of "untere kieselige Kalke" consists of very thin bedded (1-2 cm) dark greyish limestone. Shaly intervals are rare and thin. Secondary quartz occurs only in small concentrations, except in the upper part of the formation, where some thick nodular bands of it occur. Parallel lamination is the only sedimentary structure present. The age is Sinemurian based on

Eparietites sp., an ammonite characterizing the Sinemurian. The specimen was collected by R. Heusser (Swiss Fed. Inst. of Tech.). No sedimentary hiatus occurs between the Rhaetian limestones (Koessener Formation) and the "untere kieselige Kalke" and the latter probably includes the Hettangian.

The depositional environment of the "untere kieselige Kalke" was open marine, indicated by the few crinoids and the ammonite. Probably deposition occurred below the wave base but is not truly deeper marine.

(2) Alv Formation

The formalized name Alv Formation is introduced here. The type locality is the mountain Piz Alv ("Romantsch language: White Mountain). This formation is known in older literature as Alv Breccia.

The same formation occurs also in the Val del Monte area and in the highest tectonic slice of the Piz Mezzaun area (tectonic slice of Il Corn). In all these areas the Alv Formation is the same. The breccias of the Alv Formation have a red and yellow matrix. The clastic components are derived mostly from the Hauptdolomite Formation (Triassic) and vary between sand-sized and megablocks several hundreds of meters in diameter. All components are angular. Some clasts have joints which are not present in the surrounding matrix. A few clasts are derived from the Alv Formation itself.

(a) Microscopy of the Breccias of the Alv Formation

The matrix of the breccia is a dolomitic micrite to micro-sparite; some of the dolomite crystals are euhedral. Angularity of the components indicates that they were hard during transportation and deposition and early cementation had occurred. Some Hauptdolomite components have stylolitic rims. Fine grained highly oxidized laminae can be seen in the thinsections. They represent minor bedding planes. The matrix of the breccia has an irregular distributed yellow and red color from oxidized iron in the matrix. Staining of some slabs with Potassium ferricyanide showed that ferrous iron is also present in the matrix. It is assumed that the iron became oxidized during the erosion at the top of the horsts from which the breccias came. Some of the iron then became reduced during cementation in the phreatic zone.

Parts of the breccia have an interesting network of cracks. These parts are irregularly shaped and are as wide as cms across. The cracks themselves are usually less than a millimeter to 1 cm wide and can be several cm long.

Three generations of cracks could be distinguished (Figure 2).

The first generation of cracks are filled with the same material as the surrounding matrix (i.e. dolomitic micrite to microsparite). The difference between the two is the color. However, the color of the cracks is not different from the range of colors occuring in the matrix. Staining of the cracks showed that they contain also reduced iron as well as ferric iron. The chemical history of the iron was therefore the same as in the surrounding matrix.

The second generation of cracks cut through the first generation. These cracks are filled with large sparry dolomite crystals which appear brownish in the thin section. Calcite crystals are rare and are always in the center of the crack. The dolomite crystals contain only ferrous iron.

The third generation of cracks contains clear calcite and dolomite crystals. Sometimes there is a thin laminae of dirt in the center of the crack.

Figure 2a: First generation of cracks in the breccia of the Alv Formation. This generation of cracks has been leached in the vadose zone, then filled up with mud and cemented in the phreatic zone.
Arrow points to the dark-red crack.
Scale is in inches and centimeters.

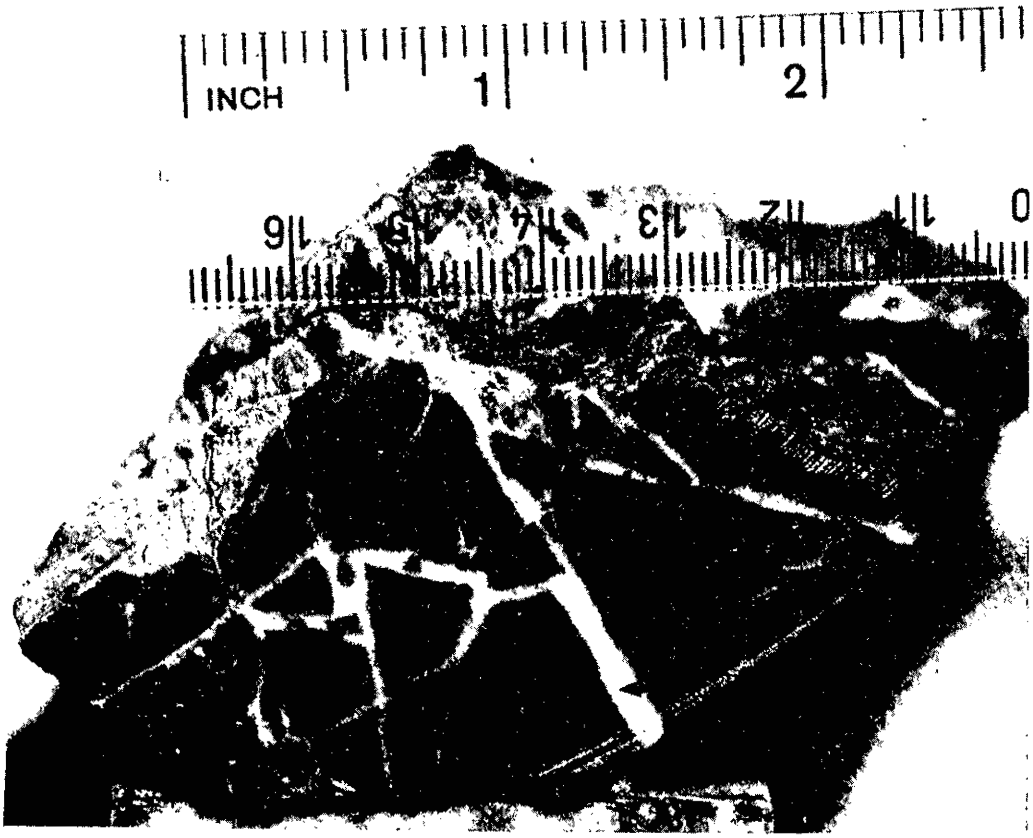


Figure 2b: Second generation of cracks in the breccia
of the Alv Formation. This generation of cracks has
been leached and cemented with sparry dolomite in
the vadose zone.
Arrow indicates one of the cracks.
Scales are in inches anc centimeters.

Offsetting and breaking loose of particles occur along all three generations of cracks. Sometimes only the first or second and third or only the second generation is present.

The lithification of the breccia occured in the phreatic zone because reduced iron is in the matrix. Reduced iron is typical for the phreatic environment; submarine environments have usually oxidized iron, because cementation may more likely happen on the water-sediment interface. Leaching in the vadose environment of the already hardened and lithified breccia resulted in the first generation of cracks. The thus-formed cracks became filled with the same kind of mud as formed the matrix of the breccia. The mud of the cracks became lithified again under phreatic conditions incorporating reduced iron. The second generation of cracks represents again leaching in the vadose environment. But this time the cracks became filled with sparry dolomite in the vadose zone, lacking reduced iron. The third generation of cracks became opened the same way the others did, but they filled with sparry dolomite and calcite under phreatic conditions incorporating reduced iron.

According to the history of its cementation, the breccia of the Alv Formation became lithified in the phreatic zone, but became leached several times in the vadose zone. Vertical tectonic movements are responsible for the change of zones.

Marine fossils, like belemnites, brachiopods, and textularidaes are rare and occur only in certain parts of the breccia. Calcitic matrix in these parts of the breccia occurs always together with the marine fossils.

(b) Sedimentary Structures in the Breccia of the Alv Formation

Very poor bedding in the breccia of the Alv Formation is indicated by thin irregular fine grained layers. These layers are not continuous laterally. Some of them stop along steep lower boundaries, suggesting that the surface of the breccia was very irregular. Others just thin out laterally as if they might have been eroded away by the overlying breccia. Their length varies between several meters and tens of meters. These layers are graded; here and there they contain a large component (several cm in diameter) which deformed the underlying layers. Convolution and loading is rare and is poorly developed.

Small scale foreset bedding and slumping can be observed in some layers.

A detailed study was conducted in a place where the layers stop along a steep base (see Figure 3):
The layers have a thickness of 1-3 cm; their components range in size between 1 and 6 mm. Some of the layers are graded, but only their upper part. Sometimes the components decrease in size laterally from 1 cm to 0.5 mm over a distance of 5 cm. The decrease in the size of the components is usually followed by a decrease in thickness of the layer. Due to compaction the layers are bent upwards along their steep base.

Small channels (Figure 4) could be seen in the breccia of the Alv Formation.

(c) Summary and Environmental Interpretation of the Breccias of the Alv Formation

1) Relation to other facies types: (see paragraph 5, p. 37). Interbedded with the Fain Breccia.
2) Geometry and paleographic relationships: (see paragraph 8, p. 40).

Figure 3: Infilling of the irregular surface or cavity of the breccia (Alv Formation) by thin bedded sediments, whose upper parts are sometimes graded. Component size increases also laterally from right to left.
Scale in centimeters.

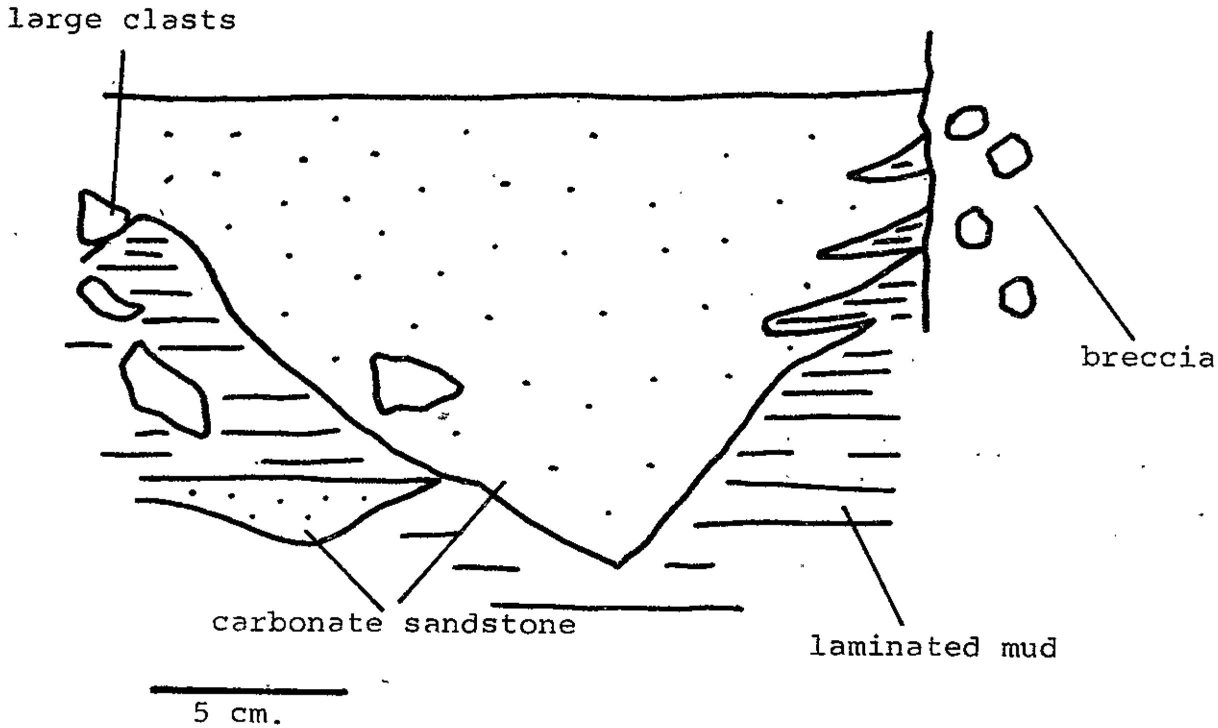


Figure 4: Small channel in the breccia of the Alv Formation.

Rim and cone like buildups along the side of the horsts.

In deep erosional holes on top of the horsts.

3) Sedimentary structures:

No large scale bedding observable.

Thin beds drape over a hummocky surface; they are not laterally continuous. The thin beds consist of graded layers.

Small scale channels.

4) Clasts:

Hauptdolomite clasts are dominant, no crystalline clasts.

Size: sand-sized particles to sedimentary blocks which are several hundreds of meters long.

Angular.

No sorting.

5) Petrography:

Matrix dominantly dolomitic.

Multiple phreatic cementation and vadose leaching.

Minor parts of the breccia have calcitic matrix which contains often marine fossils and no vadose leaching.

6) Interpretation:

The Alv Formation breccia was deposited near sea level and even above sea level because of the dominant characteristics of vadose leaching and the absence of marine fossils. The absence of large scale sedimentary structures exludes a transportation in terms of debris flows or large channeled currents. Individual sliding or rolling of each block or sheet-like deposits over the whole cone seems to be the major mode of transportation. Small currents which probably were subaerial, left behind the fine layers on the hummocky surfaces of the breccia and eroded here and there small channels. Brecciation in situ of the Hauptdolomite might have happened, but was not important.

Flashfloods etc. collected the erosional debris and transported them towards the horst edge. Certain special tracks were preferred by these floods leading to the cone-like buildups at the base of the horsts. However debris was shed all along the horsts leaving behind the typical rim of the Alv Formation breccia along the base of the horsts.

The lower parts of the breccia stayed continuously below sea level. The vertical movements of land or sea level fluctuations influenced only the upper part of the breccia.

The lower parts of the breccia are the ones with the calcite matrix and the marine fossils.

(3) <u>Fain Formation</u>

The Fain Formation is a breccia, which is distinctly different from the breccia of the Alv Formation. The type locality is in the Val da Fain ("Romantsch" language for Valley of the Hay).

The matrix of this breccia is a bedded limestone, consisting of alternating grey to black and yellowish-brown micritic limestone (Figure 5). The components of the breccia derived from the Hauptdolomite. Their size varies between 0.1 mm and several meters. This size variation does not include the sedimentary blocks which can be as large as several hundreds of meters. Extreme boudinage and stretching due to tectonism is dominant in this breccia. The black layers, the dolomite components and the crinoidal fragments are more brittle than the yellowish layers. The latter flow around the first which form the boudins. The boudins of the dark layers are oriented with the longest axis horizontal and in an E-W direction, and the shortest axis being vertical.

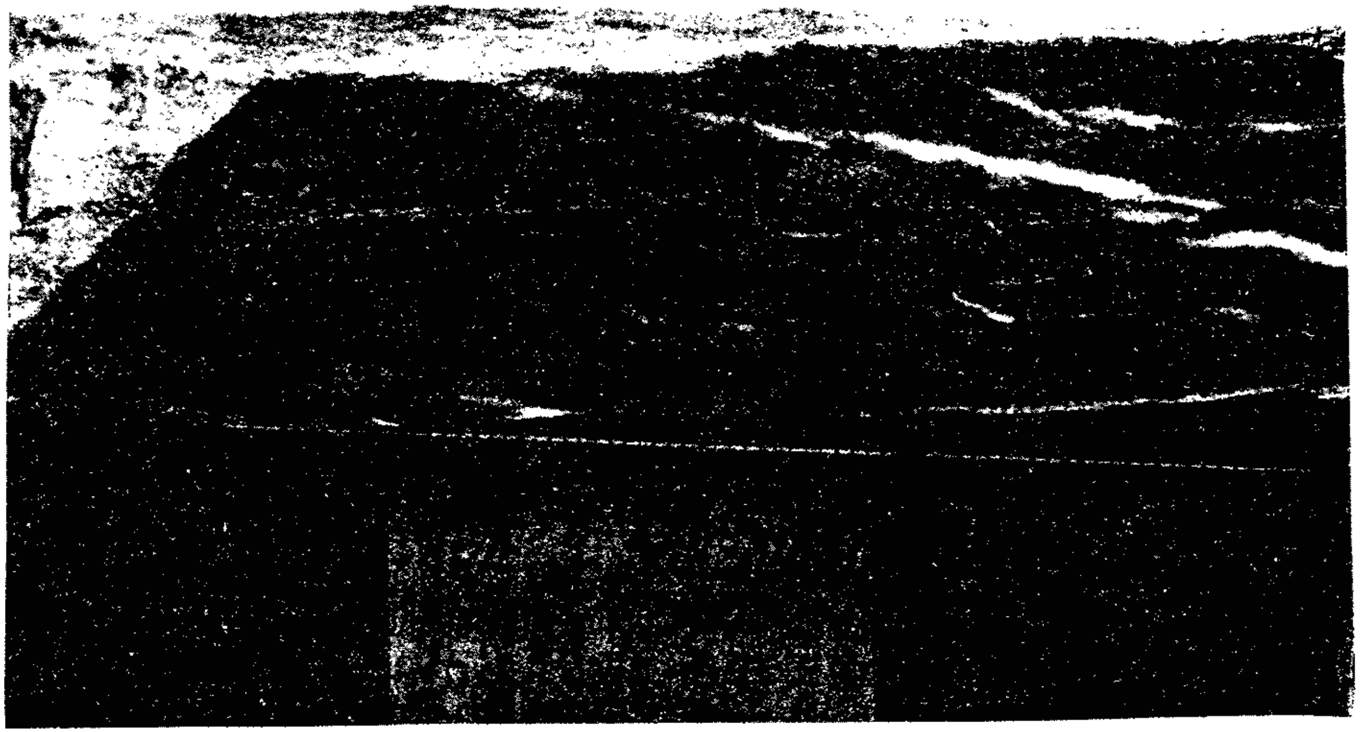


Figure 5: Breccia of the Fain Formation with belemnites,bivalves, crinoids and dolomite components.
Scale in inches and centimeters.

The formation of the boudins is contemporaneous with the formation of the Piz Alv syncline, whose axis has a N-S direction. The fauna consists of crinoids, belemnites, and brachiopods. The components of this breccia only cover 1% of the total outcropping area and are widely scattered. However the components may occur in faint layers, where their density increases to 5 and 10%. In places they disappear laterally. An example of the component size in function of the distance to their disappearance is given below:

15 m lateral distance: size of components: 0.4 m

13 m lateral distance: size of components: 0.4 m

5 m lateral distance: size of components: 0.25 m

0.0 m lateral distance: size of components: 0.0 (no comp.)

Thin section petrography also shows the flowings of the dark limestons around the more brittle dolomite and crinoidal fragments. The density of the smaller fragments is around 20%. Grading is not present. The only dominant structure is parallel lamination. Secondary silification occurs in bands and sometimes around components (see Figure 6). The quartz crystals are tectonically deformed, which means that their formation is older than the

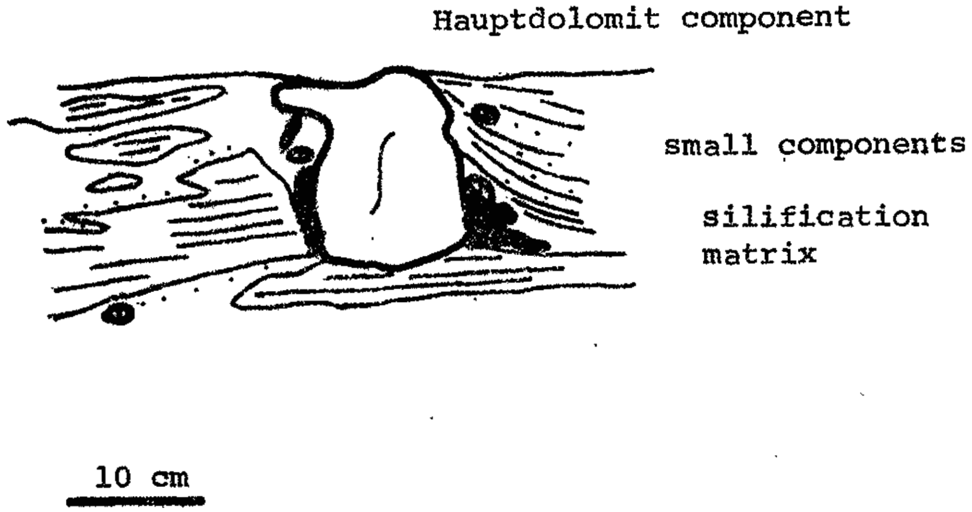


Figure 6: Typical distribution of fine and coarse clasts in the breccia of the Fain Formation.

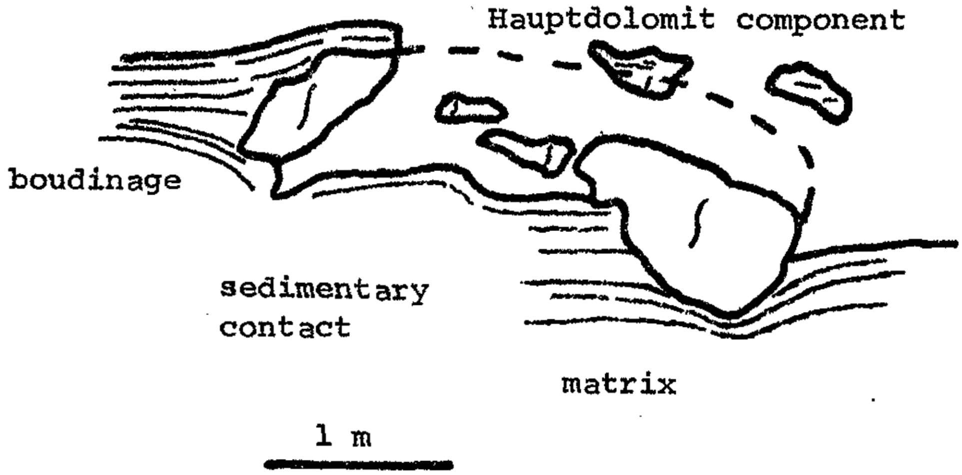


Figure 7: Large component in the breccia of the Fain Formation. The contact between the component and the matrix can be tectonic (boudinge) or sedimentary. The latter contact shows that the component sank inti the mud.

tectonic events. Detrital quartz is rare, but some sections contain up to 15% of medium to coarse silt. In places turbidite beds are within the breccia. They are coarse grained (A types, see page 53) and their grains are more abundant (20-30% density) than in the breccia. The grading is poor.

Figure 7 is an example of a large component. The bedding flows around the component due to the tectonic stretching. But in one place the bedding is cut off in a right angle. This feature is interpreted as a sedimentary structure made by the component sinking into the still soft mud. The fact that the components sank into the mud individually and the very low frequency of the components suggest an individual transportation of each component rather than a mass flow mechanism.

(a) Graphic Representation and Interpretation of the Breccia of the Fain Formation

The ten largest components were measured in each breccia layer. Sometimes less than ten components were measured because of the outcrop situation. The largest diameter was plotted against the average diameter of the ten measured components on a semilog paper (Figure 8). All the points lined up on a curve. The meaning of this curve can be explained the following way: A fine-grained breccia, for example, has a maximum diameter of 20 cm and an average diameter of 10 cm. Supposedly one would coarsen the breccia, then the coordinates of the maximum diameter and the average diameter would tend to follow the curve. But, the coarsening of the breccia is equivalent to the shedding of a coarser breccia at a different time than that of the finer one. This means that the size distribution of the components is constant for different breccias through time. A time constant distribution in the breccia can only form if the production of the clasts and the mode of transportation between source and depositional site remains constant through time.

This curve allows us to distinguish between the type of breccia which has a relatively simple genesis and breccias whose components had different and complex histories.

The breccias of the Fain Formation are the marine equivalents of the subaerial breccia of the Alv Formation, because they contain abundant marine fossils (crinoids,

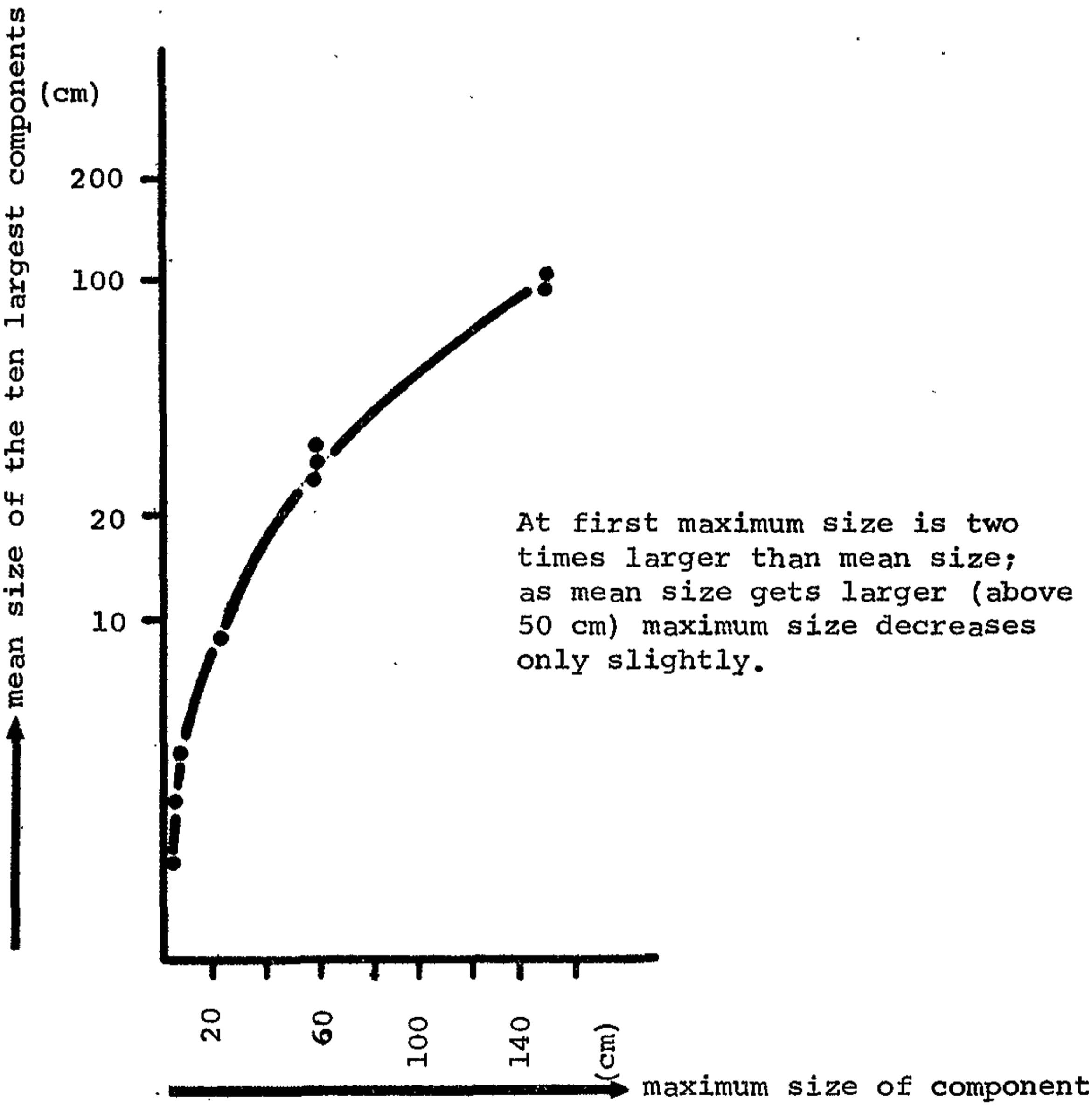


Figure 8:

Plot of maximum size of component versus mean size of the ten largest components in breccias of the Fain Formation.

brachs, and belemnites) and their position is always further basinward than the breccias of the Alv Formation (see p. 40). Boulders and pebbles slid individually into the depositional realm of this breccia. The lack of fine grained material might be explained by turbidity currents which were generated in the realm of the Fain Formation and removed all the finer material.

(4) Mezzaun Formation

The name Mezzaun Formation is formal and is introduced here for the first time. The Mezzaun Formation contains all Jurassic rock types which were previously described as "Streifenkalk," "Kieselstreifenkalk," "Kieselkalk," "Fleckenmergel," etc. of the sediments of the Bernina nappe east of the Engadin Valley. The type locality of this formation is the mountain Piz Mezzaun. The formation also occurs in the Piz Alv region. The formation consists mostly of turbidites and some few slump breccias.

(a) Mezzaun Formation in the Piz Alv Region

The turbidites of the Mezzaun Formation in this area occur north of the Val da Fain (see Plate A) where they lie on an erosional surface of a horst. On the south side of the Val da Fain there are thin beds of the Mezzaun Formation, interbedded with the Fain Formation. The turbidites of the Mezzaun Formation in the Piz Alv area are well-bedded (bedding thickness varies between few mm and 30 cm) and are dark gray. Thin breccia beds are very rare. The most common sedimentary structure is parallel lamination. Crossbedding is rare. Grading of the whole bed is not easy to define because the grain size varies mostly between fine and very fine sand. Most of the beds are classified as "b" layers of the Bouma sequence. Some layers with grain size up to coarse sand are present but not at all common.

The components of these turbidites are either Hauptdolomite fragments or contemporaneous fossil fragments (mostly crinoids). Fine to coarse quartz sand is rare; it can make up as much as about 5%. The grains are angular to subrounded.

Secondary chertification occurs as nodules and irregular elongated bands.

Sometimes wavy surfaces separate the beds or are within a single bed along the parallel lamination. These wavy surfaces are small bumps (some cm high and about 30 cm across). Their shape is spherical and they are not elongated in one direction. They are due to poor convolution. Well developed convolutions are rare. Slumpfolds and other slope indicating structures are difficult to see because of the strong Alpine deformation but in some cases still can be observed. Figure 9 shows an interesting slump feature. Two different layers of one turbidite bed can be seen on the picture. Layer A is the coarser lower part (grading) of the bed and lithified earlier than layer B which is the upper finer part. During slumping the bed was under tension which pulled apart the lower lithified layer A. Layer B, which was still soft, floated on top of the blocks of layer A and also slid in between the blocks. When the slump came to a stop, the single blocks of layer A crashed into each other and caught the upper soft part in between them. This resulted in compressional features (folds) of those parts of the upper layer which were in between the solid blocks. The upper layer contains chert nodules, implying that the chertification occurred after the slumping and after the lithification of the lower layer. The next bed above fills in the troughs where layer B is down warped between two blocks. This slump structure points out the dip of the slope which in this case is northward.

Figure 9: Slumping of a bed whose lower part (A) is lithified and whose upper finer part (B) is still soft.
For further explanation see text.
Scale: long edge of note book is 12 cm.
Arrow points in the direction of paleodip.

(5) Stratigraphic Relationships of the Facies Types in the Piz Alv Area

The breccia of the Alv Formation is concentrated on the south side of the Piz Alv where it reaches a thickness of about 200-300 m (Plate B). One interval of Fain Formation is interbedded with the breccia of the Alv Formation. This interval has a thickness of about 50 m. It seems to thicken eastwards. On the Piz Alv north side there is only one little unit of Alv Formation. The thickness of this unit is about 10 m; it thins eastwards. Most of the section on the north side of this mountain consists of the breccia of the Fain Formation. Few intervals of the Mezzaun Formation also occur.

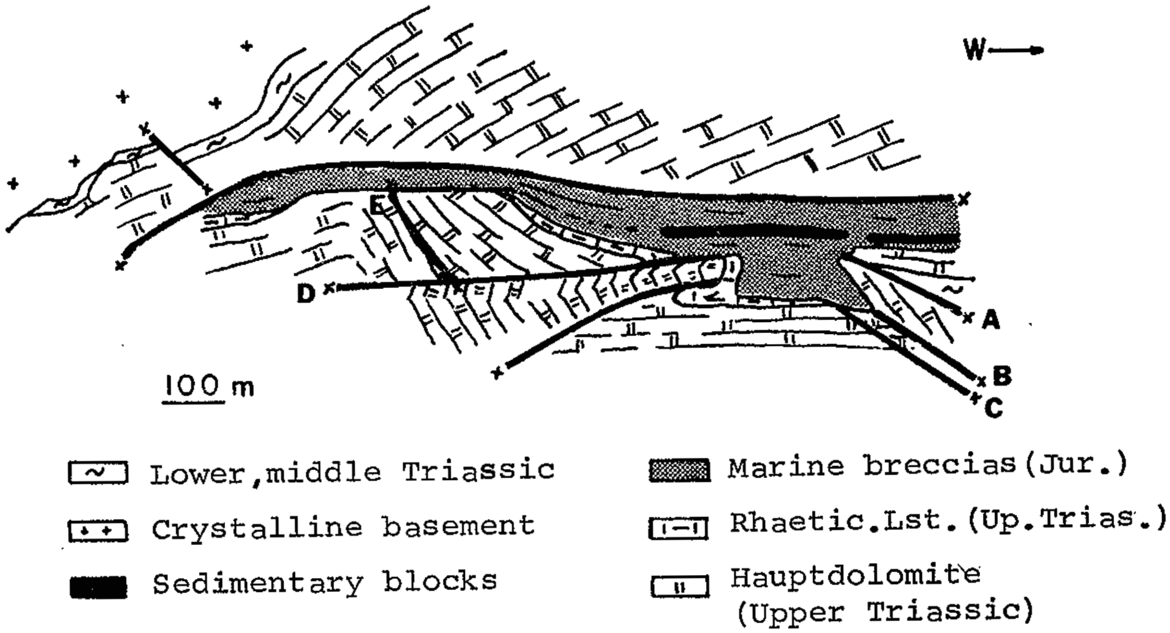
The Jurassic section on the south side of the Val da Fain consists mainly of the turbidites of the Mezzaun Formation. The Mezzaun Formation covers an old horst surface at the western end of the Val da Fain. This surface was not flat and small pinnacles of Hauptdolomite shed some breccias of the Fain Formation type into the turbidites of the Mezzaun Formation. Some erosional "holes" on these pinnacles are filled with breccia of the Alv Formation. Eastwards, the Mezzaun Formation rests on the "untere kieselige Kalke."

The latter occurs everywhere above the Rhaetian limestone as a normal succession and is older than all the other Jurassic formations. The Alv, Fain, and Mezzaun Formations are time equivalent and represent different facies types. They can overlie any stratigraphic horizon from the "untere kieselige Kalke" down to the crystalline basement.

(6) Sedimentary Blocks

Sedimentary blocks occur in the Alv and Fain Formations. The largest blocks are about 400 m long and 30 m thick. They derived from the Triassic Hauptdolomite Formation. They have been interpreted by R. Staub (1945) as being "Schuppen"; i.e. tectonically emplaced bodies of Hauptdolomite. But their relation to the Jurassic sediments is strictly sedimentary: smaller blocks are usually in the same stratigraphic horizons as the large ones. Compaction of the sediments below the huge blocks can be observed. If these blocks are "Schuppen" they must be bounded by faults. These faults should also occur in the surrounding rocks, but they do not exist there.

The large sedimentary blocks are interpreted to be edges of horsts which broke off and slid down slope into the basin. Figure 10 shows a horst whose edge broke off and



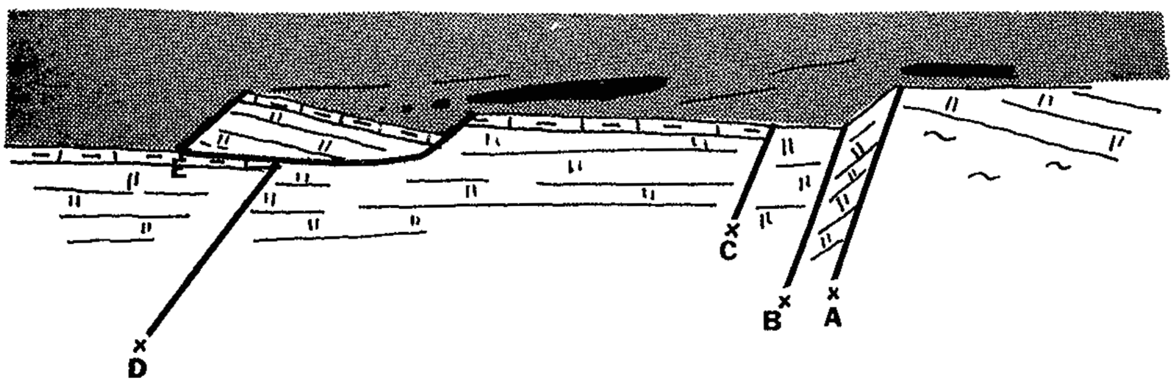


Figure 10: Tectonism during sedimentation.
Upper figure is a parallel projection of the Piz Alv north side. Lower figure is a reconstructed cross section of the same area for Lower Jurassic time.
Faults and thrusts not reprented in the lower figure are alpine.

slid along a curved fault towards the basin. In this case, however, the block is still connected to the horst.

(7) Synsedimentary Tectonism

The very coarse sedimentation, abrupt facies changes, and the sedimentary blocks within the Jurassic formations indicate syndepositional tectonic activity. It is a very difficult undertaking to find traces of this Jurassic tectonism in the highly deformed Alpine area. However, Jurassic structures may be separated from the later Alpine deformation by studying the Jurassic erosion of the Triassic sediments, knowledge that parts of the Alv Formation were deposited in a subaerial environment and the thickness distribution of the Jurassic sediments. Figure 10 is an example of the hypothetical reconstruction of the north side of Piz Alv. The reconstruction shows that normal faulting was the characteristic tectonic style in this area during Jurassic time.

(8) Paleogeography of the Piz Alv Area

The preceding chapters are the base of the reconstructed paleogeography of this region (Figure 11).

The Jurassic section can be divided into two time

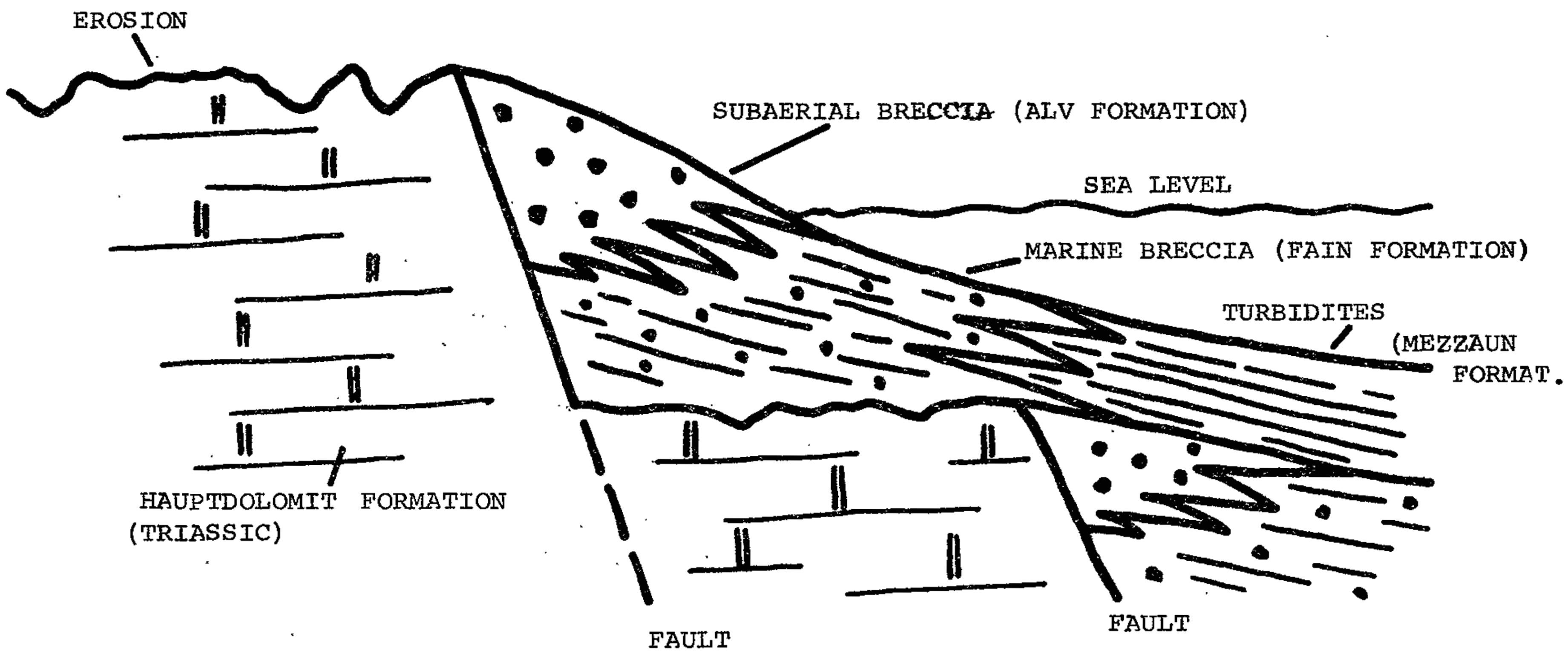


Figure 11: Schematic facies distribution of the Jurassic sediments in Piz Alv region.

Alv Formation, Fain Formation and Mezzaun Formation are Jurassic.

Hauptdolomit Formation is Triassic.

Faults were active during Jurassic time.

stratigraphic subdivisions (Figures 17-19):

A lower unit with a horst to the west, an Alv Formation cone to the south, and marine deposition to the east with Fain Formation. The western part of this horst is rimmed with Hauptdolomite. The latter has deep erosional "holes" which are filled with the breccia type of the Alv Formation. No crystalline basement components could reach the basin to the east because of this rim of Hauptdolomite at the horst edge.

The northern part of the upper unit consists of turbidites of the Mezzaun Formation. They cover the old horst surface of the lower unit in the west and the breccias of the Fain Formation in the east. The turbidites change into the breccias of the Fain Formation southward on the north side of Piz Alv. Further southward (Piz Alv south side) the breccias of the Fain Formation are interbedded with the breccias of the Alv Formation. Thus, the facies changes southward into more proximal types. Therefore, an E-W trending horst is anticipated in the area south of the Piz Alv. Unfortunately there is only poor control in this area.

(d) THE JURASSIC SEDIMENTS OF THE PIZ MEZZAUN AREA

(1) Geological Setting

Eight tectonic slices represent the Bernina nappe in the half window of the Piz Mezzaun (Plate C). They are disconnected from the main body of the Bernina nappe and were transported northward with the advancing Languard nappe. The Languard nappe is the tectonically next higher nappe and covers the eight tectonic slices to the south and east. The Engadin fault is the boundary to the west. To the north, the tectonic slices are upthrusted onto the Languard nappe along a steep thrust which is younger than the thrusts separating the elements and the thrust at the base of the Languard nappe (Schüpbach, 1973).

The highest elements are usually transported the farthest in the direction of thrusting and the lowest elements the shortest in an area with imbricated thrusts. This rule is not valid in the Piz Mezzaun area because some of the eight elements are folded and overturned. These folds are dragfolds along a young upthrust which thrusts the Lower Berninan slices onto the Languard nappe. Some of the lower elements were therefore higher in the original stack of imbricated thrusts and elements.

This folding of the thrusts, however, can not be proven in the field because erosion has removed the fold closure.

Structural cross sections based on contour maps of each thrust show only some warping and do not portray the whole folding.

Reconstruction of the paleogeography and the tectonic interpretation of folded thrusts is therefore based on sedimentological correlation.

For this correlation it is assumed that all elements belong to one sedimentary basin. This assumption seemed reasonable by studying the facies types. The eight tectonic slices are (from the tectonic top to the base):

1) Tectonic slice of Il Corn
2) Upper tectonic slice of the Val Lavirum
3) Lower tectonic slice of the Val Lavirum
4) Upper tectonic slice of the Piz Mezzaun
5) Tectonic slice of Medras
6) Tectonic slice of Stevel
7) Lower tectonic slice of the Piz Mezzaun
8) Tectonic slice of Chamues-ch

Plate C shows the stratigraphy and distribution of these elements.

(2) Stratigraphy

Crystalline basement and Triassic facies types, when preserved, are the same as in the Piz Alv area.

The highest tectonic slice, the Il Corn slice, has Alv Formation at its base; above is Fain Formation, which in turn is overlain by Mezzaun Formation. The lowest tectonic slice, the slice of Chamues-ch, has "Steinsberger Lias" at its base. The "Steinsberger Lias" occurs at the base of the Jurassic sections which are paleogeographically further north (Bugliaun series). The "Steinsberger Lias" is overlain by the Mezzaun Formation in the tectonic slice of Chamues-ch. The tectonic slices between the highest (Il Corn) and the lowest (Chamues-ch) slice contain only the Mezzaun Formation. Slump breccias in the Mezzaun Formation occur in two elements. A thin irregular layer of a breccia similar to the Alv Formation lies between the Hauptdolomite and the Mezzaun Formation in the tectonic slice of Medras, suggesting subaerial exposure. The area of the tectonic slice of Medras must have been a horst during early Lias.

(a) Age of the Mezzaun Formation

The Mezzaun Formation is overlain by radiolarian chert. The radiolarian chert is very wide spread in the East Alpine realm and the South Alps. Its age is Malm. The base of the Mezzaun Formation in the Stevel slice contains few meters of a dark limestone which had several specimens of _Arietites_ sp. This ammonite is characteristic of the Sinemurian. The dark limestone might be equivalent to the "untere kieselige Kalke" in the Piz Alv area. The Mezzaun Formation is therefore slightly younger than the Sinemurian and older than the Malm. The Mezzaun Formation in the tectonic slice of Medras contained belemnites (_Passalotheitis_ sp.; determined by C. R. Stevens) whose age is middle Liassic or younger.

(3) _Sedimentology of the Mezzaun Formation_

The Mezzaun Formation consists of turbidites. Bouma sequences, transported bioclastics, grading and the concept of deep water sedimentation prove the existence of turbidites. The source of the turbidites was in the Piz Alv area, where they grade up dip into the breccias of the Fain and Alv Formations.

These turbidites have mostly Bouma's "a" and "b" intervals. Intervals "c" and "d" are very rare. This can be explained by the grain size distribution produced in the source area.

The lower part of a typical "a" interval (Bouma) is a few cm thick, well graded, and contains rock fragments of the Hauptdolomite Formation and crinoidal debris. The grain size varies between 2 cm and about 2 mm. The next higher part of Bouma's "a" interval is much thicker (several cm - 1 m). It contains mostly crinoidal debris. The grain size at the base us usually 2 mm and at the top between 0.2 and 0.1 mm. Grading is poor because of the limited grain size distribution.

Bouma's "b" interval has a grain size variation between 0.1 and 0.2 mm. The clasts are crinoidal debris, and no grading is observable. Parallel lamination is very common. Bouma's "b" interval is directly overlain by pelagic shales and/or pelagic marls with Chondrites. Bouma's "c" and "d" intervals do not exist because the grain size necessary for these intervals is not present. The grain size distribution produced in the source was very limited. Erosion of the Hauptdolomite Formation on the horsts did not produce many grains much smaller than the original dolomite crystals. The remaining grains in the turbidites are mostly crinoidal fragments consisting of single crystals which were not broken down further. The production of very fine sand, silt and clay sized material was therefore limited. The coarse grain size distribution affects also the coarse lower parts of the turbidites so that there is no matrix between the coarse crinoidal debris and rock fragments. The grain contacts are obliterated by solution (stylolitic rim) or by calcite overgrowth around crinoidal fragments. Real cementation is occasionally observed and consists of micrite and microsparite. Secondary silification is common and occurs in bands and elongate nodules. The thickness of these chertified zones varies between some mm and 30 cm; the length varies between 10 cm and several m. The chertified zones contain quartz crystals, which have a very irregular shape and include parts of the surrounding calcitic matrix, suggesting that they were formed after the deposition of the sediment. The size of these quartz crystals is inversely graded. Most clastic grains are crinoidal fragments which are single crystals. The smaller a crystal is the easier it is to dissolve and replace it by quartz. Silification advances, therefore, faster in the finer grained upper parts of a bed, resulting in the inverse grading of the quartz crystals. The amount of sponge spicules varies; sometimes they are a rather important constituent.

Few Bouma "a" intervals have quartzitic clastics. In one example clast size and composition changes from a coarse carbonate sand into quartzitic fine sand. The quartz grains are angular to subangular. The matrix in between them is still calcitic.

(4) Correlation of the Mezzaun Formation Between Different Tectonic Slices

The problem of the paleogeographic reconstruction in this area was stated above (p. 43): the eight tectonic slices stacked upon each other can not be put back into a logical sequence from the structure alone. Only facies correlation can be used to put the tectonic slices back into their original position so that they form a logical paleogeography. But before a facies correlation is possible, time correlation must be established.

This chapter and the following subchapters deal with the facies and time correlation between tectonic slices.

The Mezzaun Formation is the key for this correlation. The other Jurassic formations do not occur in all the slices and the Triassic facies types are too monotonous and do not change from one slice into another.

The facies correlation for reconstructing the tectonic slices has to be based on a sedimentological model. This model has sedimentological charactersitics which vary as a function of the depositional environment and therefore of the paleogeography. This variation will be used for correlation. As example one of the characteristic variables could be the bedding thickness of the turbidites as a function of the paleogeography.

(a) Model of the Sedimentological Environment in the Mezzaun Formation

Introduction

The sedimentological model developed in this chapter will be used later for the correlation of the tectonic slices. The model is partly hypothetical and partly based on field observation.

An empiric description of the turbidites in terms of their variation as a function of the paleogeography (depositional environment) will be given first. Then the turbidites will be classified into intervals (i.e. sedimentological subdivision of each turbidite bed) in order to approach the problem in a more statistical manner. The last part of this combines the empiric description in terms of paleogeographical variables and the classification of the turbidites. From this combination results a statistical pattern of the intervals (classification of turbidites) in terms of paleogeographical variables.

An Empiric Description of the Turbidites in Terms of their Variation as a Function of the Paleogeography

The turbidites of the Mezzaun Formation grade laterally into the breccias of the Fain and Alv Formations. The Piz Alv area which contains the coarse breccias of the Fain and Alv Formations and subaerially eroded horsts is the edge of the basin. The tectonic slices of the Piz Mezzaun area contain the basinal sediments in form of the turbidites of the Mezzaun Formation. The turbidites of the northern Piz Alv area must therefore represent the slope sediments, because they are the link between the breccias of the Piz Alv area and the basinal turbidites of the Piz Mezzaun area.

Field observation revealed that the turbidites on the paleoslope (i.e. northern part of the Piz Alv area) are thin bedded, fine grained and lack pelagic sediments (marls, shales). The turbidites in the eight tectonic slices of the Piz Mezzaun area are thick bedded, coarse grained and contain some pelagic sediments, although some of the slices contain turbidites whose bed thickness and grain size is intermediate, but their content of pelagic sediments far exceeds the amount of the other two turbidite types. Based on these observations the following model is developed:

The turbidites largely bypassed the upper slope and deposited only thin beds with fine grained material. The turbidity flow lost most of its velocity and energy when it reached the basin where the slope was less steep. Loss of energy resulted in deposition of the coarse fraction. The resulting deposits are coarse and thick bedded. The distal part of the basin received only fine grained and few turbidite deposits, because only a small amount of the sediment was carried that far. Pelagic sediments form thick (up to 5 m) layers between the turbidite beds. This distal part of the basin is represented by those finegrained turbiditos containing an excess of pelagic sediments which now are present in some of the tectonic slices in the Mezzaun area.

In summary, a single turbidite bed is thin and fine grained on the slope, thick and coarse grained at the base of the slope and in the near-slope basinal areas and then thins and becomes finer-grained farther out in the basin.

Classification of the Turbidites

A classification of the turbidites has to be made in order to quantify the model obtained from the field observations. The deep water sediments in the Mezzaun Formation were classified in the following way:

The capital letters A-D, M will be used later in the text and in the figures. The letters were capitalized to distinguish them from Bouma's designation of intervals (see Figure 13).

A: Is the massive lower interval of a turbidite, which is usually graded and has no other sedimentary structures. It is equivalent to Bouma's "a" interval. The grain size varies between several cm at the bas and 0.2 mm at the top.

B: Is the same interval as Bouma's "b" interval.

Parallel lamination is dominant. The grain size varies between 0.2 and 0.1 mm. The grading is therefore poor.

C: This interval consists of several turbidites. It is introduced here to describe alternation of turbidites and pelagic sediments which are too thin to measure in the field. The thickness of a single turbidite or layer of pelagic sediments does not exceed 0.5 cm.

D: Pelagic shales.

M: Pelagic marls with Chondrites and other trace fossils.

Bouma's "c" and "d" intervals are not present because the grain size necessary to form these intervals is not present.

Relative Abundance and Thickness of the Intervals (A-D,M) on the Slope and in the Basin

The relative abundance of the intervals is a function of the paleogeography, i.e., they vary in their abundance in different environments like slope, near slope basinal areas and distal parts of the basin. The relative abundance of these intervals can therefore be used as a characterisitic for the correlation of the turbidites, i.e., to reconstruct the tectonic slices into a logical sequence. The turbidites bypassed the upper slope. Only fine-grained intervals were deposited in the northern part of the Piz Alv area. Most of the intervals are of B and C type. Pelagic sediments are rare because of the bypassing of many turbidites. The A intervals occur mostly in the Piz Mezzaun area or in terms of paleogeography at the base of the slope. They disappear in the more distal regions of the basin. The B and C intervals continue farther out in the basin. This means that the B and C intervals are relatively abundant on the slope and the B intervals in the distal parts of the basin. The C intervals disappear in the distal parts of the basin, because the pelagic sediments between the turbidites (B) become thicker than 0.5 cm; the turbidites will then be classified as bioherms. The relative amount of the intervals M and D, which are pelagic sediments, increases the more distal the section is.

The relative abundance of each interval is therefore known in different environments: slope - near slope part of the basin - distal part of the basin.

A relative abundance means that no exact number is known, but the abundance can be expressed in terms of a percentage. Therefore a diagram can be drawn which has on one axis the environments such that the most proximal environment is to the left and the most distal to the right of the horizontal axis (Figure 13). The vertical axis of this diagram contains the relative abundance (percentage) of the intervals. This way the change of the relative abundance of the intervals is expressed continuously from a proximal to a distal environment.

The distribution in thickness of each interval is also described above: rather thin beds on the slope and thick beds in the near slope environment. The intervals thin gradually from this environment to the distal part of the basin. This thinning of each interval towards the distal part of the basin can also be used in a statistical manner:

The more distal section should have the thinner beds than the more proximal one. The thickness of each interval type can therefore be plotted on a histogram (Figure 15a), whose vertical axis contains the frequency and the horizontal axis the thickness classes (thin intervals to the left, thick intervals to the right). Given the case that there are two turbidite sections with equivalent age. One of the two histograms of these two sections will have the wider variation of thicknesses (i.e. contains thicker intervals), that indicates a more proximal environment to the other section which has thinner intervals. Nature of pelagic sediments and the grain size distribution are other variables which can be used as environment indicator. However they will be used only to confirm the results obtained.

The model of the sedimentological environment in the Mezzaun Formation provides two continuous variables (nature and thickness of intervals) indicating the environment. These variables will be used in the next chapter for correlation.

(b) Methods of Correlation

As mentioned above the correlation of the Mezzaun Formation between the tectonic slices must be a time and facies correlation.

In this highly deformed alpine area it was not possible to measure representative sections in each tectonic slice. Sections were measured in the northern part of the Piz Alv area, in the tectonic slice of Medras, the lower tectonic slice of the Piz Mezzaun and the tectonic slice of Stevel. Some sections cross the whole formation, others represent just part of it.

Bed by bed correlation is impossible in these turbidites. Correlation of the entire formation between the tectonic slices was also impossible, because the complete sections are not always present, and there was no way to tell how much was missing. Therefore, the sections were divided into "units."

The distinction between these units is based on the presence or absence of certain intervals and their thickness. The sections were plotted in such a way that the interval thickness is on the horizontal axis and the interval type (A-D, M) on the vertical axis (Plate D). Each interval is represented by a line, whose length marks the thickness and whose position (distance from the horizontal line) matches up with a certain type. The division into units in these sections can be made visually. For example, the section of the lower tectonic slice of the Piz Mezzaun was divided into two units. The lower unit is thin bedded and contains marls (M) and shales (D) which are missing in the upper thicker bedded unit. The section in the northern Piz Alv area (S-chuedella) was divided into two units. The section of the tectonic slice of Medras was divided into four units. The tectonic slice of Stevel could not be divided into units because of its uniformity.

Two independent operations were applied for the correlation:

(a) Relative distribution of the intervals (A-D, M) on the slope and in the basin.

(b) Thickness variation of the A and B intervals in the more distal portion of a turbidite.

The correlation and reconstructed paleogeography was confirmed by two additional observations:

(c) The maximum grain size at the base of the A intervals.

(d) The pelagic sediments.

Relative Abundance of the Intervals (A-D, M) as a Function of the Environment

The relative abundance of the intervals as a function of the environment was presented in the chapter "Model of the Sedimentological Environment in the Mezzaun Formation" and is shown in Figure 13.

The relative abundances (percentages) of each interval was calculated for each unit. The sections were then arranged so that their distribution of intervals in the correlated units resemble the theoretical diagram of Figure 13.

It turns out that the arrangement which matches best the theoretical diagram is a correlation presented in Figure 14. All units 1 can be correlated and all units 2 can be correlated (time correlation). The facies correlation resulted in the following position (from proximal to distal, or south to north) of the tectonic slices: northern Piz Alv area - tectonic slice of Medras - lower tectonic slice of Piz Mezzaun - tectonic slice of Stevel (Figure 12). The whole section of the tectonic slice of Stevel was compared with the different units of the other sections; the former is too uniform to divide into units. It is still possible to determine the environment of the section of the tectonic slice of Stevel because the section is uniform and the facies correlation is done in terms of relative abundance of intervals.

Thickness Variation of the A and B Intervals

The thickness variation of the intervals can indicate the depositional environment between the base of the slope and the distal part of the basin (see chapter "Model of sedimentological environment in the Mezzaun Formation"). Different units are compared with histograms representing the thickness variation of a certain interval. Thicker intervals (wider variation) occur in the more proximal environment (Figure 15a).

Correlation was made in the sections of the lower tectonic slice of the Piz Mezzaun and the tectonic slice of Medras. The correlation did not include the section of the tectonic slice of Stevel, because this section contains only one unit and several are needed. The section of the northern part of the Piz Alv area was also not included, because it contains slope sediments, which are inappropriate for this correlation.

The units of the different tectonic slices must also be correlated in time before they can be correlated in times of facies. Time correlation was made on the basis of trend relations between units.

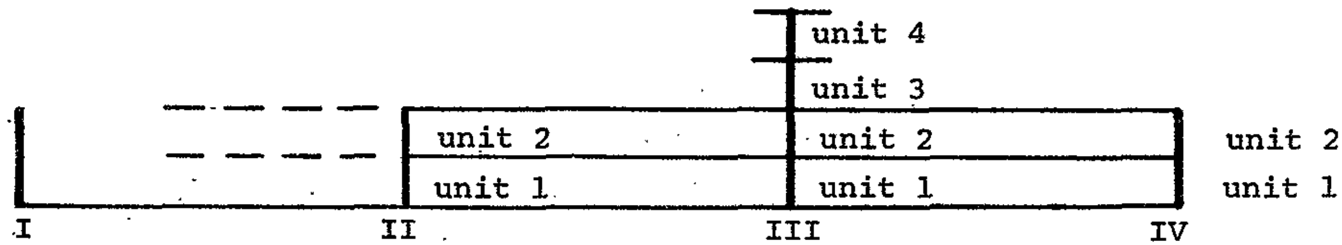
Each section is subdivided into units and the relative proximality (or distality) was determined in each unit by the thickness distribution of the A intervals. A certain section (as example the section of the tectonic slice of Medras) showed a trend(Figure15b): unit 1 (Medras section) is less proximal than unit 2; unit 2 is more proximal than unit 3, which is also more proximal than unit 4. The trend in the section is from a distal environment to a proximal and then back to a series of distal environments. The same analysis can be made in the section of the lower tectonic slice of the Piz Mezzaun (Figure 15b). This section contains only two units. The trend in this section is from a more distal to a more proximal environment (from unit 1 to 2). This trend can only be correlated with the lower two units (unit 1 and 2) of the section of the tectonic slice of Medras, because these have the same trend (from a more distal to a more proximal environment upwards in the section). The upper part of this section (from unit 2 on upwards) has the wrong trend, from proximal to distal.

The correlation can be rechecked by using the B intervals.

The above analysis was only the time correlation. This correlation does not give any information on the paleogeographic position of the sections (tectonic slices). Facies information will give this information. This correlation was made the same way as the time correlation, but this time the time correlated units of the different sections are compared with the help of the histograms (Figure 16). Unit 1 of the lower tectonic slice of the Piz Mezzaun has the same age as unit 1 of the tectonic slice of Medras. This analysis was first made with A intervals and then checked with B intervals. Both analyses show that unit 1 of the lower tectonic slice of Medras is more proximal than unit 1 of the lower tectonic slice of the Piz Mezzaun. The same analyses are then repeated for units 2 of both tectonic slices (both units have the same age according to the time correlation). The analysis with units 2 resulted in the same paleogeographic order, pointing out the correlation endured multiple independent checks.

Maximum Grain Size at the Base of the A Intervals

The maximum grain size at the base of the A intervals did not give a direct clue for correlation but agrees generally with the correlation pattern stated in paragraphs (a) and (b) on pages 59 and 60. The maximum grain size at



I - IV are measured sections in different tectonic slices.

Figure 12: Time and facies correlation of the units (Mezzaun Formation) in the measured sections.
The sections to the left are in the distal, the ones to the right in the proximal environment.
The section of the tectonic slice of Stevel (I) is not subdivided into units, because of its uniformity. Only the facies and not the time correlation could be done.

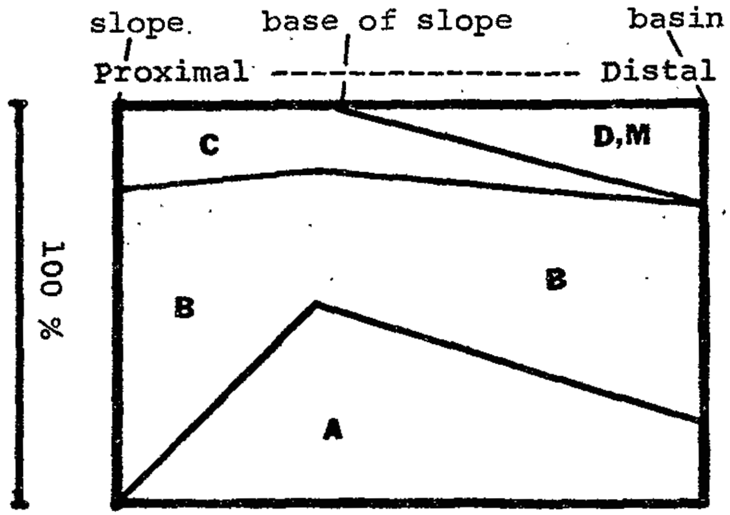
I: Tectonic slice of Stevel

II: Lower tectonic slice of the Piz Mezzaun.

III: Tectonic slice of Medras.

IV: Northern part of Piz Alv area (S-chuedella).

The sections (and units) are not to scale.



On the vertical axis is A-D distribution of a certain unit.

On the horizontal axis is the distance from a proximal to a distal environment.

Theoretical distribution of the abundance of the intervals A - D from a proximal to a distal environment.

B Same like Bouma's "b" interval, parallel lamination.

A Massive interval, graded, same inter= like Bouma's "a".

C Interlayered B and D (M) intervals which are thinner than .5cm.

D M Pelagic sediments, shales and marls with Chondrites.

Figure 13: Classification of the turbidites into intervals (A,B,C,D and M). The diagram in the upper part of the Figure shows the relative abundance of the intervals from a slope to a basinal environment (proximal-distal).

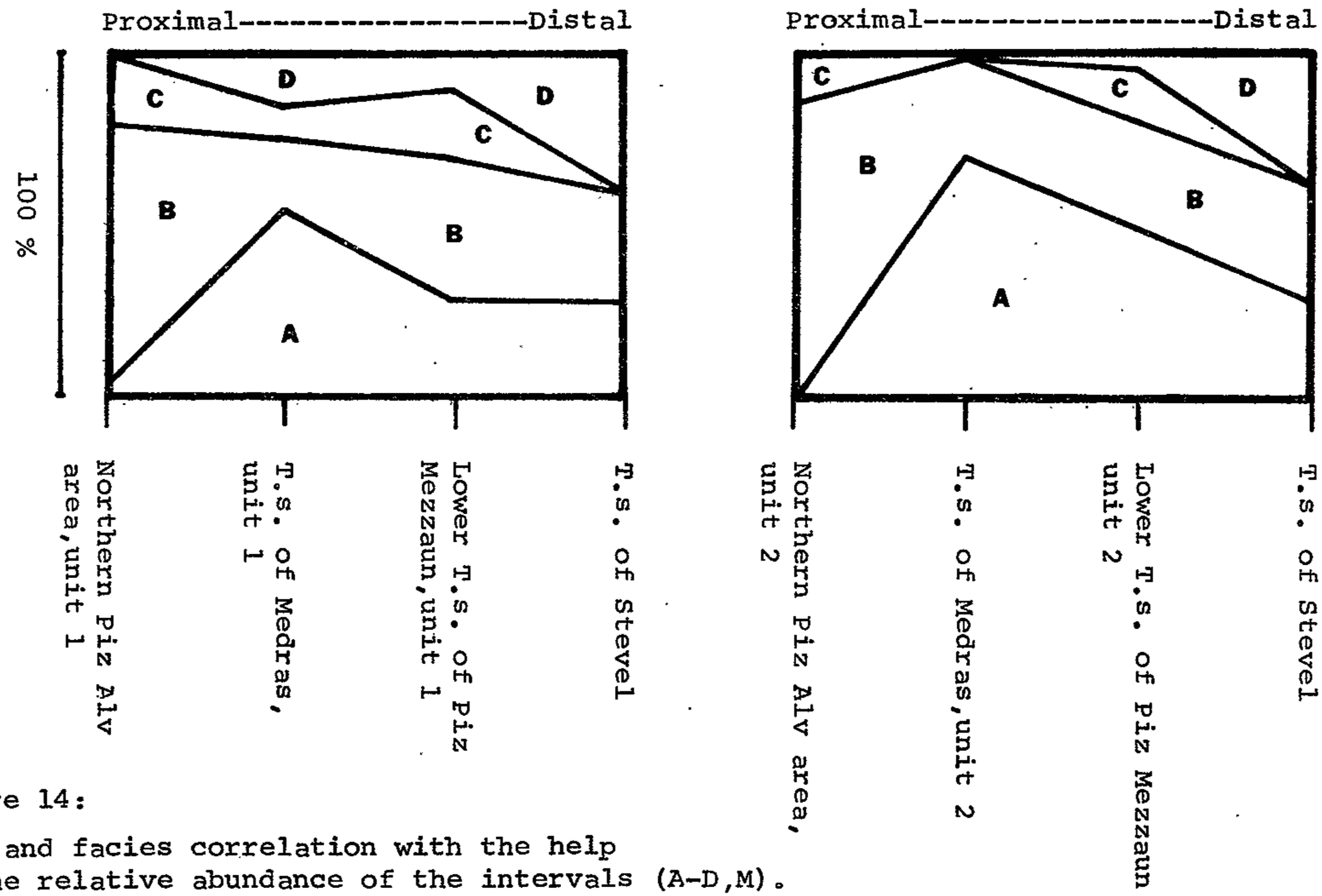


Figure 14:

Time and facies correlation with the help of the relative abundance of the intervals (A-D,M). These two diagrams resemble the theoretical diagram (Figure 13) most. The diagram to the left correlates uints 1 and the one to the right units 2.

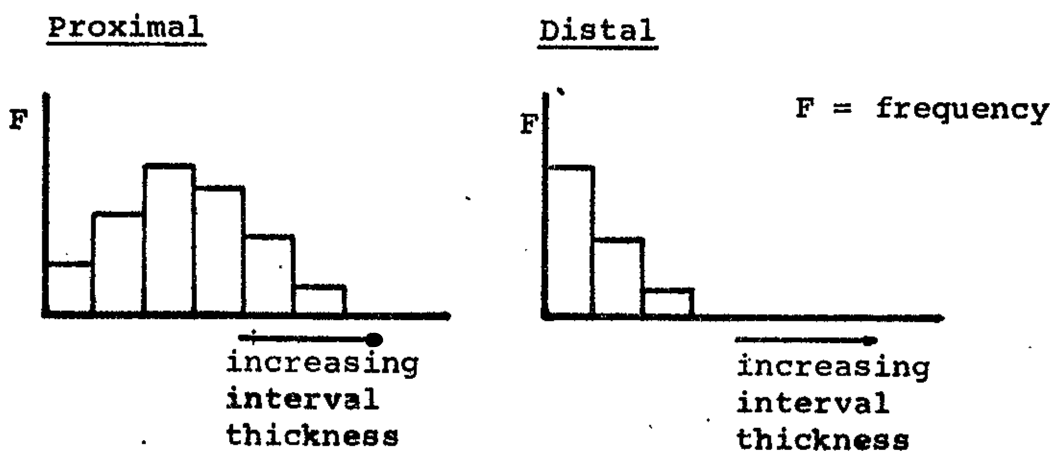


Figure 15a: Thickness distribution of intervals (A or B) indicates proximality or distality. This analysis is only valid in distal portions of the turbidites; i.e. in the basin and not on the slope.

A wide variation of thickness (i.e. thicker beds) is indicative for a more proximal environment. A narrow variation (i.e. thinner beds, they all occur on the left side of the diagram) is indicative for a more distal environment.

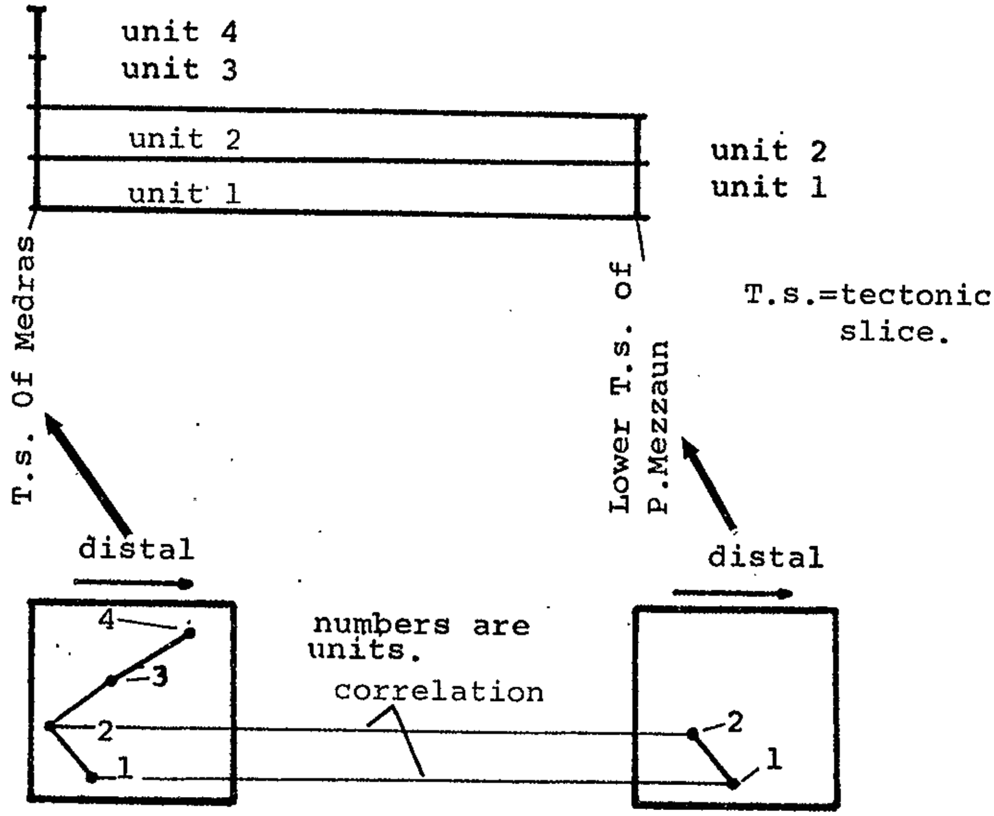


Figure 15 b: Time correlation of units in different sections.

The diagram in the lower left of this Figure represents the section of the tectonic slice of Medras. This section has 4 units. A trend towards proximality can be obeserved from unit 1 to 2; from unit 2 to 4 the trend is towards distality. The diagram in the lower right is the section of the lower tectonic slice of Piz Mezzaun. The section has 2 units and the trend from unit 1 to unit 2 is towards proximality. The correlation is based on the similarity of trends. Proximal trends are correlated with proximal ones, and distal trends with distal ones. The resulting correlation is in the upper part of this Figure.

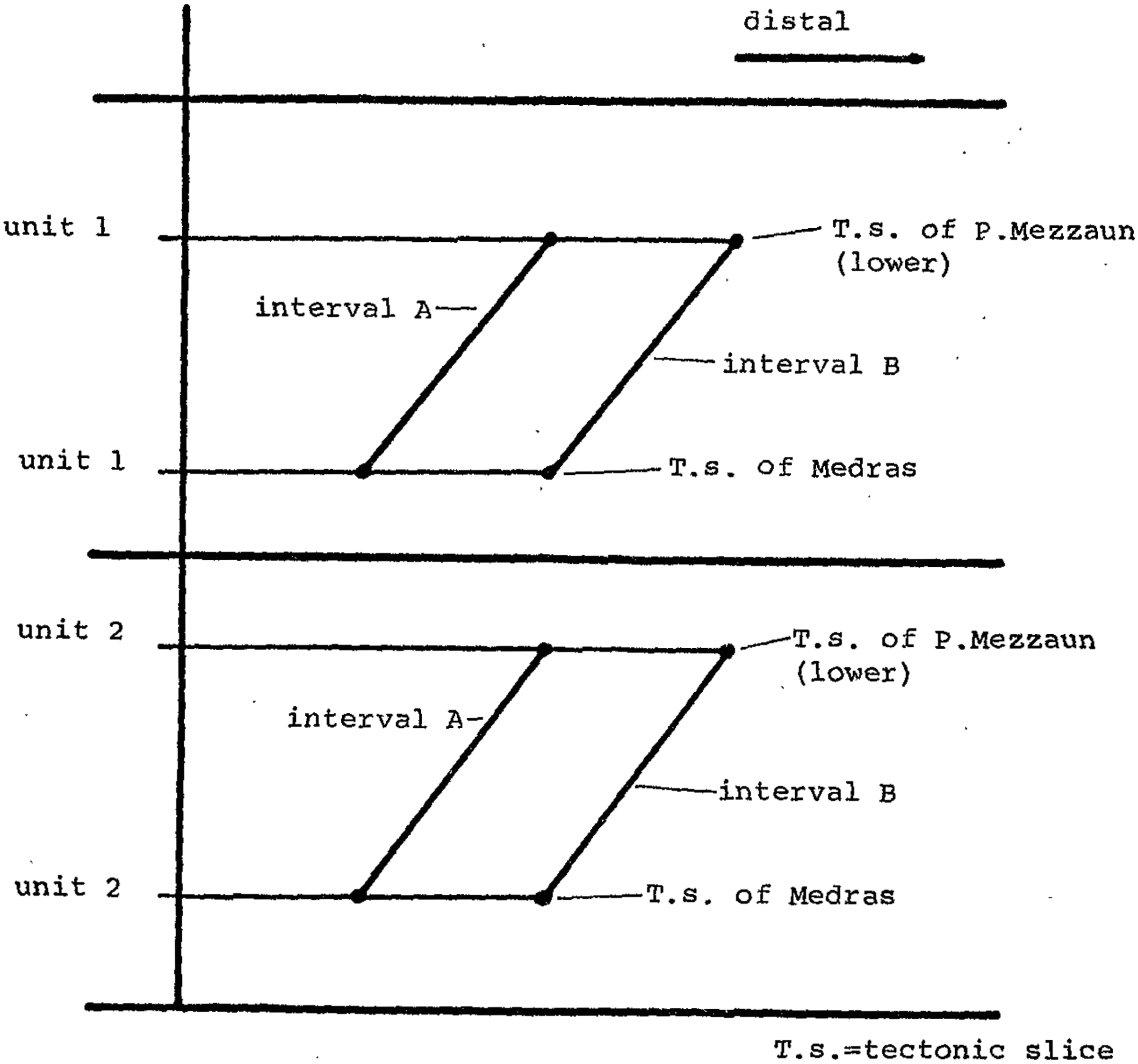


Figure 16: Facies correlation of the different units. Unit 1 of the tectonic slice of Medras correlates with unit 1 of the lower tectonic slice of the Piz Mezzaun; units 2 of both slices correlate the same way (see Figure 15 a,b). Intervals A of both units are compared (this Figure) in terms of their distality with the help of the histograms. Unit 1 of the lower tectonic slice of the Piz Mezzaun is more distal. This analysis is checked by using B intervals. A correlation with units 2 (A,B) gave the same results (lower part of this Figure).

the base of the A intervals varies between 0.86 and 3.2 mm in the section of the tectonic slice of Medras and is about 0.63 mm in the lower tectonic slice of the Piz Mezzaun.

Pelagic Sediments

A vertical as well as a horizontal facies change from a more proximal into a distal environment exists in the Mezzaun Formation in the Piz Mezzaun area. The Mezzaun Formation in the tectonic slice of Medras changes from the first unit to the second unit from a relative distal to a proximal environment. But from the second unit on upwards the change is towards a more distal environment. The fourth unit contains large amounts of pelagic marls (M) (Figure 12,17). This unit grades upwards into shales and radiolarian chert. A deep environment with no detrital influx usually reflects a high concentration of MnO. The shales contain about 0.22% of MnO. The radiolarian chert can contain concretions of manganese with very high percentages.

The lateral gradation of the lower part (units 1 and 2) of the Mezzaun Formation in the tectonic slice of Medras into the Mezzaun Formation of the lower tectonic slice of the Piz Mezzaun and into the tectonic slice of Stevel (i.e. a gradation into more distal environments) shows first an increase of pelagic marls (M) with Chondrites and then further distally a simultaneous increase of pelagic shales and a decrease of pelagic marls. The former contain as much as 4.68% MnO. Radiolarian chert would be theoretically the most distal sediment type, if the necessary water depth was present and the environment tectonically preserved.

(5) Fans, Channels and Submarine Canyons

The presence of submarine fans and channels would severely interfere with the correlation methods used here. Submarine fans have channels which have a much more proximal environment than the surrounding fan surface.

Channelling is the most characteristic attribute of fans. Nowhere could one observe channelling in the Mezzaun Formation, not even in the northern Piz Alv area where a lateral cross section of about 7 km exists. Fans occur often at the end of submarine canyons, the submarine canyon acting as a point source, in front of which the fan builds up. But no canyon is seen in the Piz Alv area and therefore no point source is envisaged. However, the breccias of the Alv Formation collected in cone-like buildups along the horsts (Figure 17) suggesting a subaerial point source. The cones extended a short distance into the basin and occurred in a narrow strip parallel to the edge of the horst. A little breccia has been shed all along the horsts. A well localized influx of detritus, which could act as a point source, seems to be absent.

Therefore a wedge of turbidites occurred all along the continental edge, with no lateral buildups or fans.

(6) Conclusions

Both correlations resulted in the same time and facies pattern, which was confirmed by the grain size distribution and pelagic sediments. The resulting paleogeography is shown in Figure 12. Units 3 and 4 of the section in the tectonic slice of Medras are not represented in the other sections, except perhaps in the section of the tectonic slice of Stevel. The missing part on top of the other sections is interpreted as being cut away by thrusts.

The resulting paleogeography influences the structural geology in such a way that folded thrust mentioned on page 43 is supported.

(e) PALINSPASTIC PALEOGEOGRAPHY OF THE PIZ MEZZAUN, PIZ ALV AND VALLE DEL MONTE AREAS

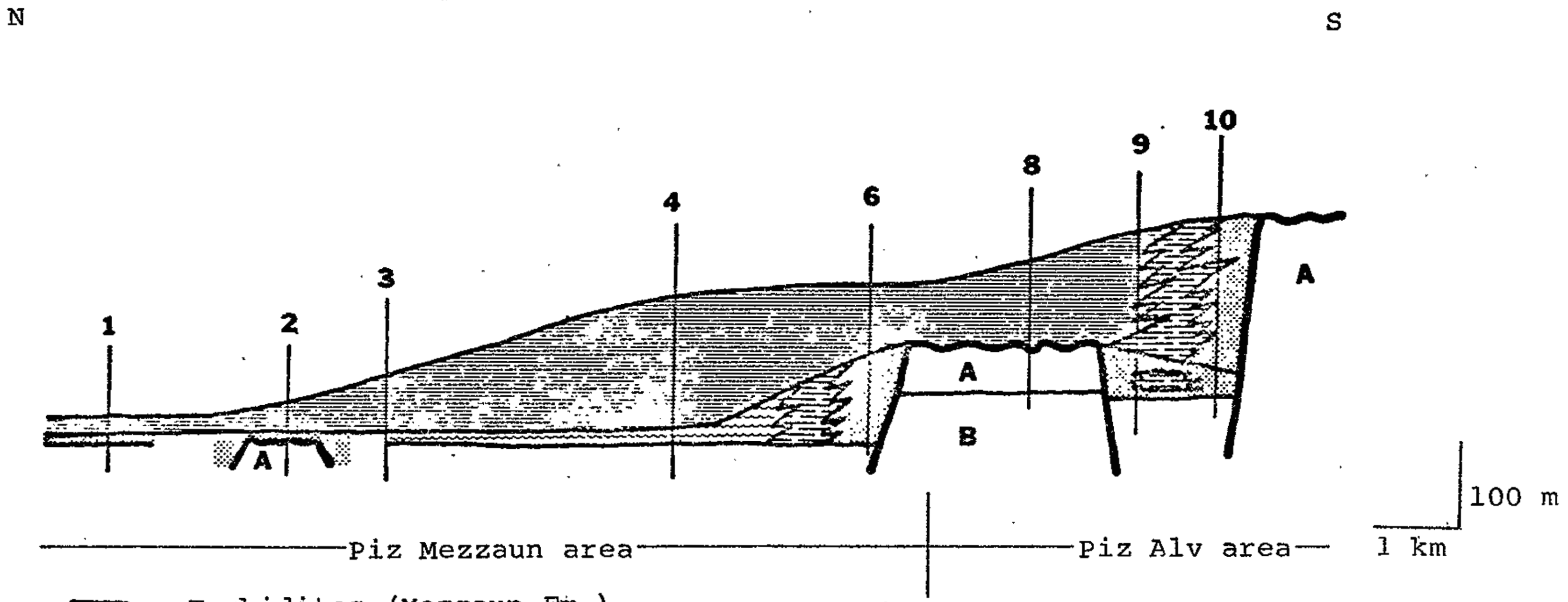
It is possible to reconstruct the paleogeography of the Piz Mezzaun area with the help of the correlations discussed previously (Figures 17, 18, 19).

The Piz Alv area and the Piz Mezzaun area are separated by the crystalline rocks of the tectonically higher Languard nappe over a distance of about 10 km. (Plate A). The Piz Mezzaun area consists of eight tectonic slices stacked upon each other (Plate C). These slices represent the sediments of the Bernina nappe, which originally were between the Piz Alv and Piz Mezzaun areas. The advancing Languard nappe picked up the sediments of the Bernina nappe in the form of the tectonic slices and stacked them all upon each other in the Piz Mezzaun area (Schüpbach, 1973). One can link them together in the reconstructed paleogeographic maps and cross sections of the Piz Alv and Piz Mezzaun areas without leaving out any extended paleogeographic areas. However, the paleogeographic distances between the sections in different tectonic slices are very difficult to judge.

Figure 18 is a palinspastic paleogeographic map of the lower subdivision in the Piz Alv area and the thin sediments below the turbidite in the Piz Mezzaun area. The map does not include the Hettangian and Sinemurian siliceous limestones (untere kieselige Kalke), because the deposition of these sediments predates the tectonic activities (horst and graben formation) and the diversification of the facies types. This early tectonic activity (Figure 18) includes horst structures in the "distal" and "proximal" parts of the basin. The sediment types along the horsts are the same as in the upper unit (Alv and Fain Formations) but the basinal facies is different. Thin dark limestones occur with dark irregularly distributed breccias containing components which are predominantly from the uppermost part of the Triassic dolomites. The breccias are interpreted as short distance slide and slump features. The sedimentation in the basin was slight and unimportant. Turbidites are missing because the slopes were not long enough to generate them. There is geological control for the orientation of the middle or main horst. The orientation of the northern and southern horsts (Valle del Monte) is less certain; they are drawn more or less parallel to the main horst. Crystalline basement is eroded in the southwestern part of the main horst (Figure 18). The direction of the main shedding of crystalline components must have been towards southwest, because no such components are found north, south, or east of this horst.

The upper subdivision of the Piz Alv area has the same age as the turbidites of the Mezzaun Formation in the Piz Mezzaun area, because they could be correlated with the turbidites of the northern Piz Alv area, which in turn can be correlated with upper subdivision of the Piz Alv area (Figures 17, 19). Only one horst was formed during the time of the upper subdivision, but can not be seen because it is south of the Piz Alv area (p. 42). Its orientation is inferred from the distribution of facies. The clastic influx from the horst was enormous. The slope of the horst graded northward over the older main horst (Figure 17) and connected with the northward slope of the older horst. These slopes were long and steep enough for the generation and transportation of the turbidites into the newly formed basin to the north.

Figure 17 : Palinspastic cross section of the Piz Mezzaun and Piz Alv areas.



Turbidites (Mezzaun Fm.)

Marine breccia (Fain Fm.)

Subaerial breccia (Alv Fm.)

Dark limestones and breccias

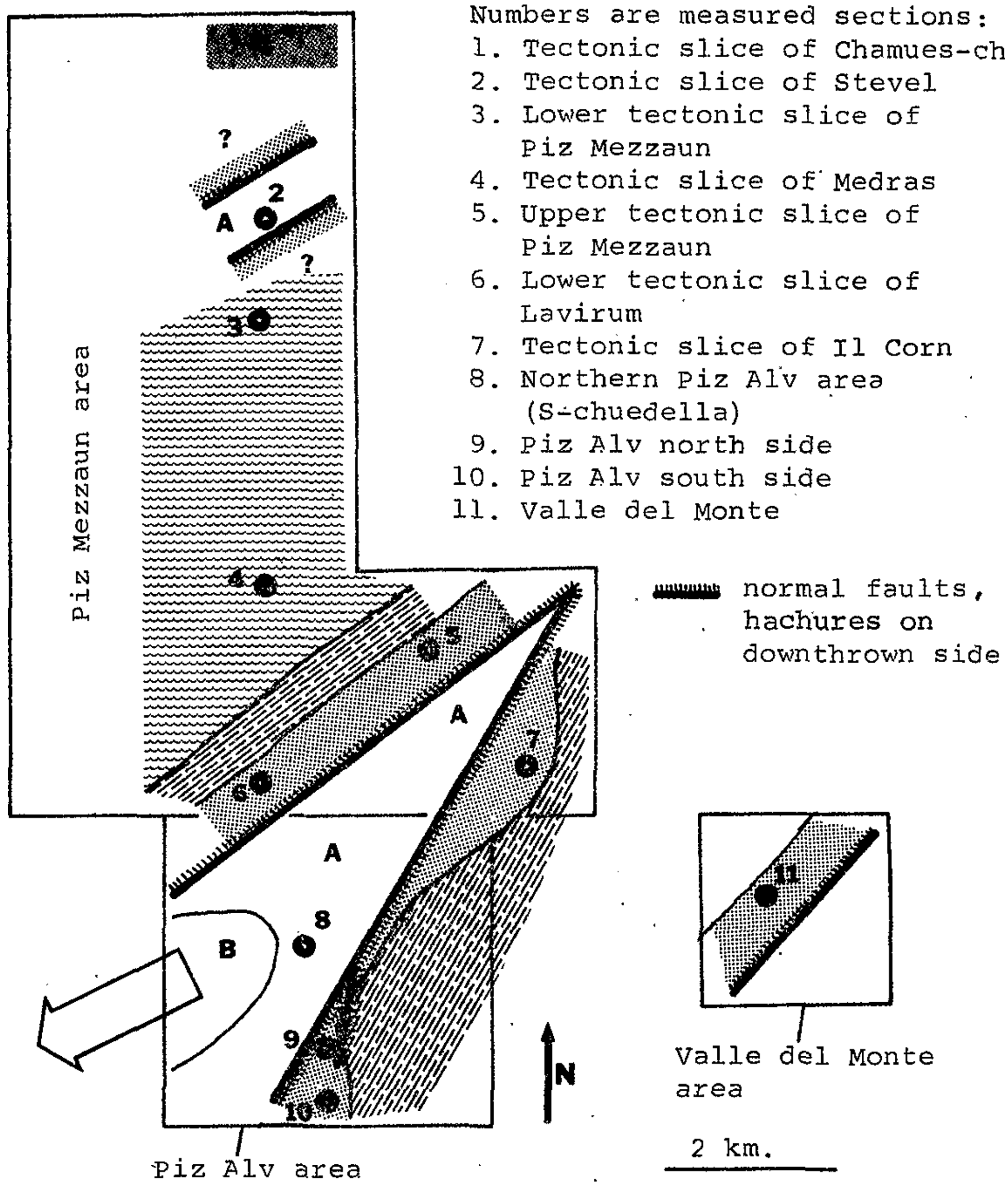
"Steinsberg Lias"

Siliceous limestone (Lower Lias)
Triassic dolomites

Horsts; A: Erosion of Triassic dolomites
B: Erosion of crystalline basement

Measured sections of the following tectonic slices:

1. Tectonic slice of Chamues-ch
2. Tectonic slice of Stevel
3. Lower tectonic slice Piz Mezzaun
4. Tectonic slice of Medras
6. Lower tectonic slice of Lavirum
8. Section in the northern Piz Alv area
9. Piz Alv north side
10. Piz Alv south side



Arrow indicates direction of main transportation from the horst into basin.
A: Erosion of Triassic dolomites on the horsts.
B: Erosion of crystalline basement on the horsts.
Symbols for rock types: see Figure 17.

Figure 18: Palinspastic map of the lower subdivision in the Piz Alv, Valle del Monte and Piz Mezzaun areas.

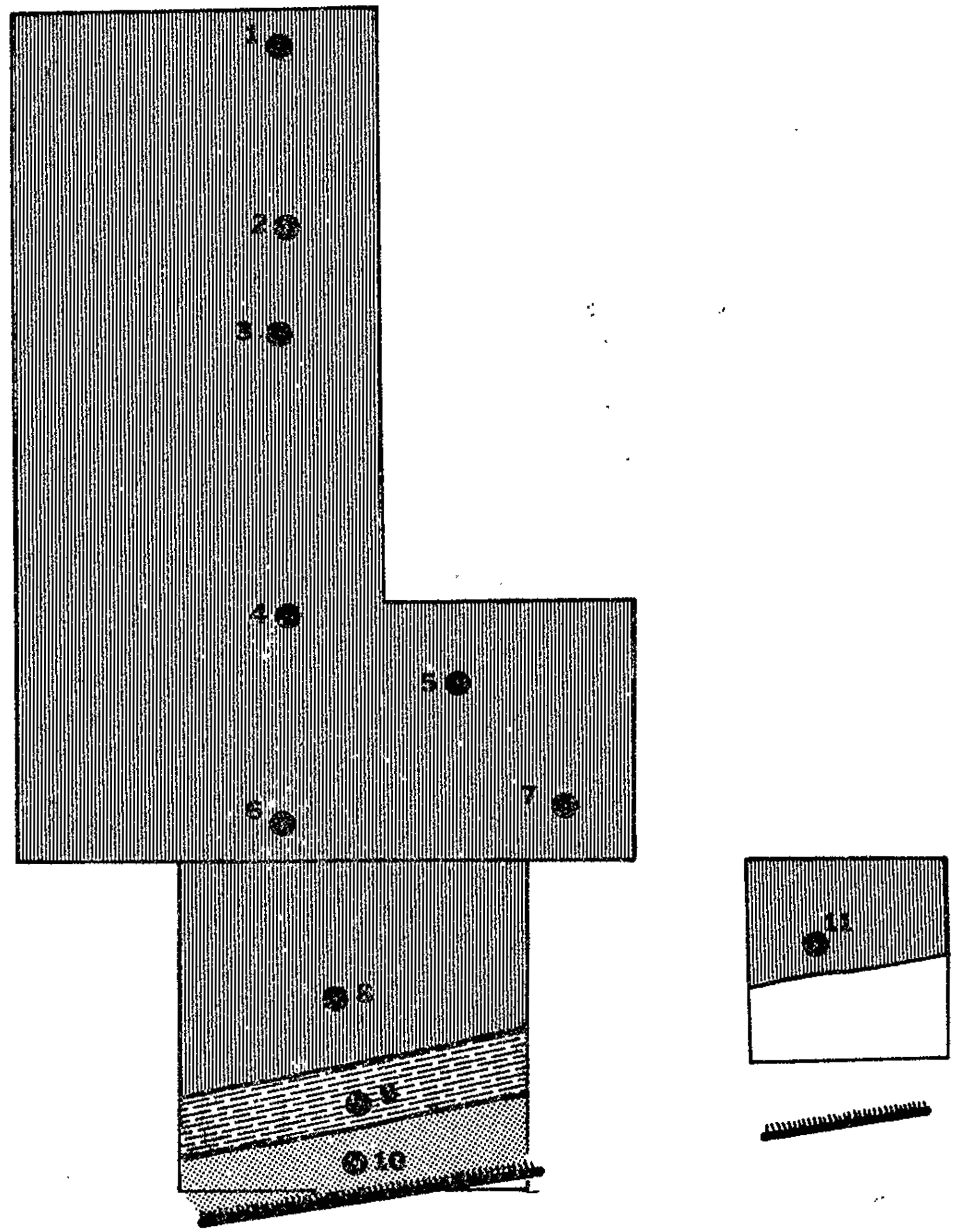


Figure 19: Palinspastic map of the upper subdivision in the Piz Alv,Valle del Monte and Piz Mezzaun areas.

Lithologic symbols and locations of sections: see Figures 17 and 18.

EXAMPLE OF A MARGINAL CRATONIC BASIN

(a) INTRODUCTION

The Permo Pennsylvanian Pedregosa Basin was investigated as an example of a marginal cratonic basin.

It occurs in Chihuahua, southwest New Mexico and southeast Arizona. The best outcrops of its eastern margin occur in the Big Hatchet mountains (Hidalgo County, southwestern New Mexico). The basic stratigraphy and structure of this area were worked out by the late R. Zeller. Basinal sections do not crop out in the Big Hatchets, but were drilled by an Exxon well (Humble no. 1 State "BA"). The cuttings and logs were made available by the Bureau of Mines and Mineral Resources of New Mexico. Core chips were made available by Exxon Co., Midland Texas. Some of the sections of the Big Hatchets were measured by C. Jordan, J. L. Wilson, and J. Tovar. Numerous sections in the field and detailed studies of the well were made by the author.

Special interest was focused on the formation of the megabreccias as a slope indicator.

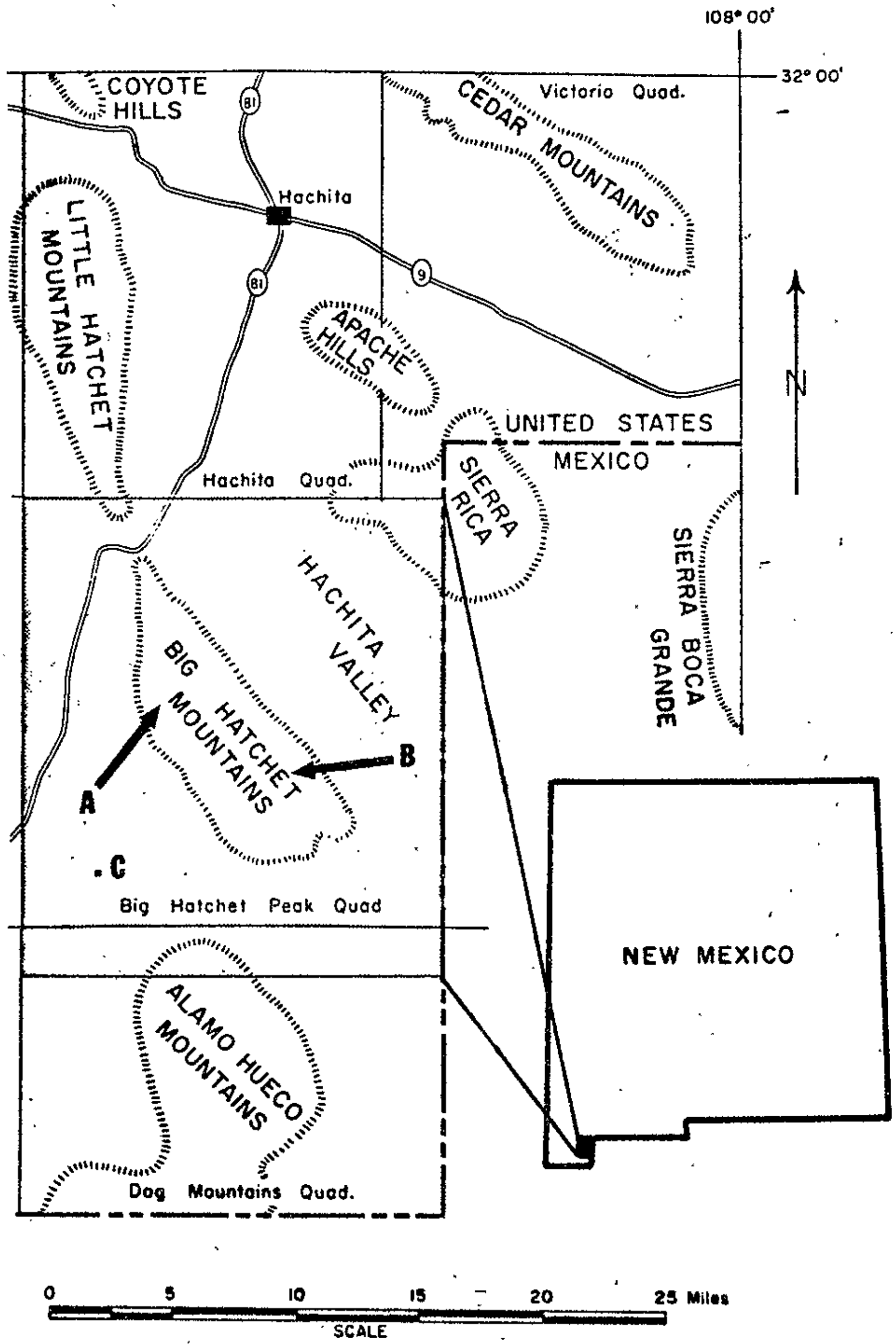


Figure 20: Index map.
A: Cement Tank Canyon
B: Sheridan Canyon
C: Exxon no.1 N.M. "BA".

(b) GEOLOGICAL SETTING

The upper Pennsylvanian-Wolfcampian Pedregosa Basin was mainly in Chihuahua and did not reach far into New Mexico and Arizona (Figure 21). The direction of the basin axis is about northwest to southeast. The northern end of the basin grades into the San Pedro Outer Shelf (C. A. Ross, 1973). This Outer Shelf is flanked on the north side by the Mogollon Inner Shelf and on the south side by the Papago Inner Shelf. Both shelves can be traced southward where they flank the Pedregosa Basin.

Parts of the shelf area were islands with no sedimentation but erosion of pre-Pennsylvanian sediments. The Florida Island or High was located northeast from the Big Hatchets (Figure 21). Northwest of the Florida Island was a large complex of Islands: Zuni, Defiance, and Kaibab Highs. The relationships between these Highs and the Florida Islands is obscured by the Mesozoic Burro uplift which had more or less the same position as the late Paleozoic Islands. Erosion stripped off all pre Mesozoic sediments on the Burro uplift. East of the Florida Island was also a shelf, and farther east the Orogrande Basin, also bordered by a shelf area (Diablo platform) at its eastern side.

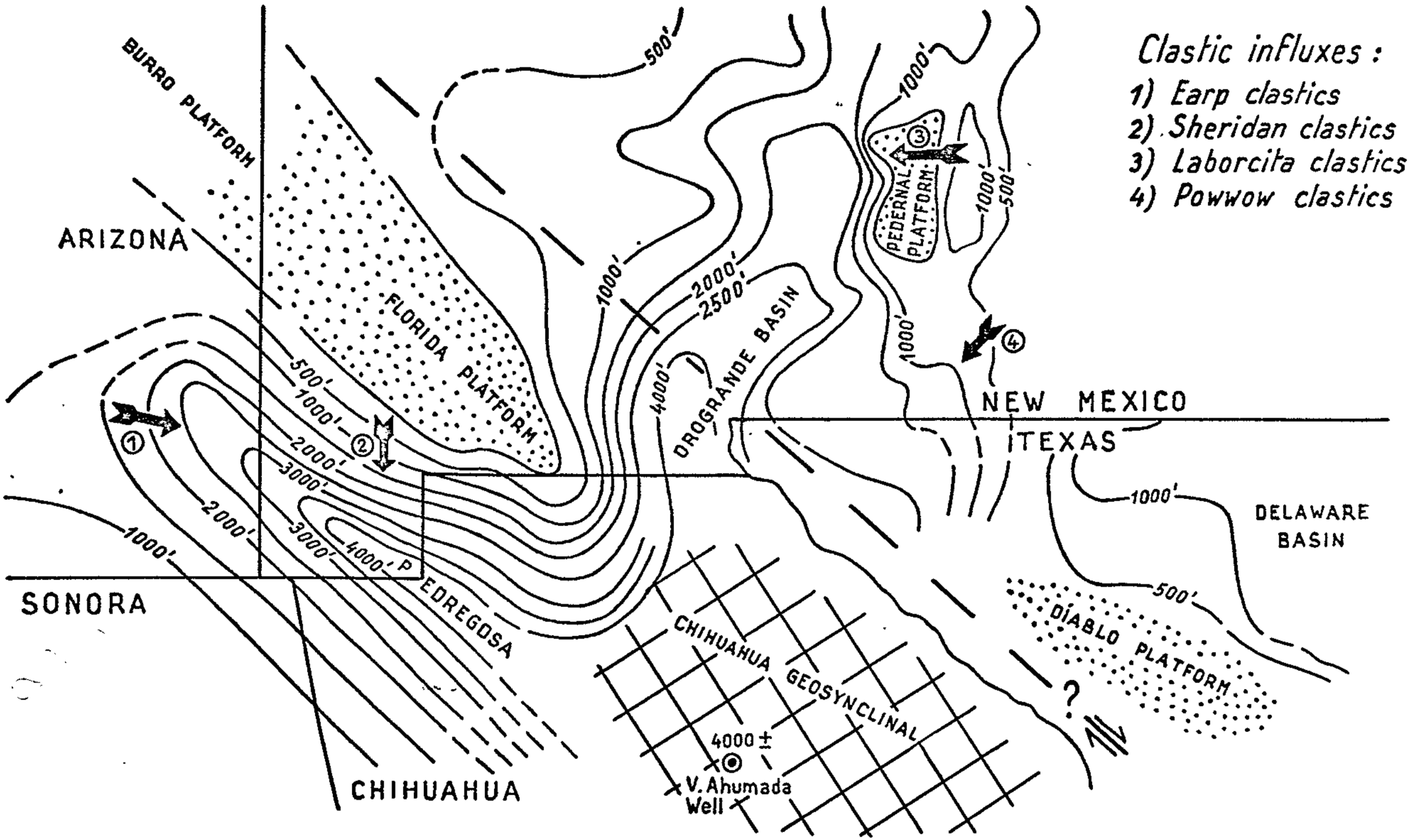


Figure 21 : Wolfcampian isopach map of Pedregosa and Oro Grande Basins
From J.L. Wilson, 1970 and C.F. Jordan, 1971.

The platforms and basins in southwest New Mexico and southeastern Arizona have a lineation whose orientation is northwest-southeast. This lineation might be due to basement faulting.

The Pedregosa and Orogrande basins grade southward into the Chihuahua Trough. This trough contains thick sections of Permian flysch. It is, however, not known at present if this trough was a foredeep of the Ouachita-Marathon Geosyncline or even part of this geosyncline. In the latter case its position is too much north of the Ouachita-Marathon Geosyncline (Figure 21). This implies large right lateral strike slip faults. The strike slip faults would be located at the eastern margin of the Chihuahua Trough right on the Texas Lineament. L. L. Corbitt and L. A. Woodward (1970) discuss the evidence for thrusting in southeastern Arizona and southwestern New Mexico. Studying their map shows a northward thrusting generally north of the Hatchet Gap and southward thrusting south of there. R. Zeller's map (unpublished) shows the same situation. Wrench faults are characterized among other things by steep lower parts and flat thrust like upper parts of the faults (Wilcox, Harding, and

Seely, 1973). The thrust directions of the upper parts of the fault are directed away on each side from the main strike slip direction. The geometry on a map would look like the thrust distribution in the Big Hatchet area.

But the lateral movements might also be contemporaneous with the sedimentation and of Permo-Pennsylvanian age. This implies a transform fault. The direction of this transform fault would be the same as the orientation of the platforms and basins north of the Chihuahua Trough.

(c) STRATIGRAPHY

This study deals only with the upper Virgilian and Wolfcampian part of the Horquilla Formation. This formation was defined by R. Zeller in the Big Hatchet area. It spans practically the whole Pennsylvanian and part of lower Permian. The upper contact of the Horquilla Formation in the Big Hatchet area is within the middle and upper Wolfcampian (C. Jordan, 1971). The top of the Horquilla Formation became laterally replaced towards the west by the shallow marine and continental Earp Formation, whose deposition ends the Pedregosa basin. In the Chiricuhua Mountains (southeastern Arizona), the top of the

Horquilla Formation reaches only into the Virgilian.

(d) GENERAL CHARACTER OF THE HORQUILLA FORMATION IN THE BIG HATCHETS.

The division of the Horquilla Formation into a lower subdivision (Morrowan - middle Desmoinesian) and an upper subdivision (middle Desmoinesian - middle-upper Wolfcampian) was made by R. Zeller (1965). The lower subdivision consists of oolitic and other shallow marine carbonate deposits; slightly deeper water carbonates consist of crinoidal and fusulinid rich regularly stratified limestones, which may contain chert. The Pedregosa Basin was not yet developed during the lower subdivision of the Horquilla Formation.

But the Pedregosa Basin began to develop during the upper subdivision. First it developed slowly. Desmoinesian and Missourian strata do not have many biohermal buildups. For example, thin bioherms (30 feet thick) grade basinwards into 7 foot thick Syringopora beds indicating that at this time not much of a slope was present. In the lower Borrego section (age: Missourian-Virgilian; locality: Sheridan Canyon) occur shallow carbonates interbedded with deeper water carbonates.

The real basin with biohermal buildups, at the shelf edge, breccias on the slope, typical basinal facies (turbidites and deeper water limestones) and terrigenous clastics developed during upper Virgilian and lower-middle Wolfcampian time.

Later, the typical basinal and shelf environments became covered with the shallow marine-continental clastics of the upper Wolfcampian Earp Formation.

(e) FACIES PATTERN ON THE SHELF EDGE AND UPPER SLOPE (EXAMPLE: CEMENT TANK CANYON)

(1) Introduction

The following facies definitions are partly based on a thesis by C. Jordan and the author's work, but they are mainly based on numerous reports and papers of J.L. Wilson.

The shelf edge could be divided into four environments: (1) shelf, (2) biohermal mounds, (3) flanking beds, (4) slope (Plate 3)

(a) Shelf

The sediments of the shelf area consist dominantly of open marine shallow water carbonates with some scattered small biohermal mounds and few oolitic bars. The matrix is usually micrite; some dolomitization occurs.

The scattered mounds are of wackestone with platy algae, Tuberitina, Tetratataxis and some Tubiphytes. Dasycladacean algae may have grown on the top of the bioherms if the water was shallow enough (less than 15 feet).

The normal open-marine shelf limestone consists of wacke- to packstone with the typical open marine faunal elements like echinoderms, brachiopods and fusulinids. Platy algae and Tubyphytes are present but in smaller amounts than in the bioherms. Smaller amounts of tubular forams, Girvanella, Paleotextularia and Globivalvulina are also present. Some concentrations of "Osagia" onkoids may occur in shallow lagoons behind the biohermal buildups. Tubular forams, Tetrataxis and Tuberitina were usually attached to platy algae and occur with the latter.

Pelecypods and ostracods are not bound to any specific environments. The oolitic bars are of grainstone with typical shallow water biota such as dasycladacean algaes.

Bedding in the shelf environment is usually well developed, except for the lensoid mounds.

(b) Biohermal Mounds

The microfacies of the bioherms consists of wackestones with abundant platy algae; less important are Tubiphytes. Tuberitina and Tetrataxis are much more abundant than Paleotextularia and Globivalvulina because they are attached to the platy algae. "Osagia" onkoids and Girvanella may be present in minor amounts.

The bioherms have a mound like shape and are elongated along the shelf edge. They grew below the wave base. When the bioherm reached the wave base, fauna and texture changed. Grainstones with abundant worn and broken tubular foraminifera mark this higher energy zone. Also present are Globivalvulina, Paleotextularia, crinoidal debris, brachiopods and fusulinids. This grainstone bed caps the top of the bioherm and is usually only a few feet thick.

The platy algae which build up the bioherm could presumably not live deeper than 120 feet, because of photosynthetic activity. The bioherms are usually 60-70 feet thick. Wave base must have been at a depth of about 60 feet, because the bioherm quit growing at the wave base, but started out at a depth of 120 feet, assuming a stable sea level after the beginning of the bioherm growth. It is possible that the platy algaes could live deeper than 120 feet. Halimeda, also a green algae, was found in water as deep as 200 feet. Applying this number for the platy algaes: the thickness of the biohermal buildup would imply that the wave base was at a depth of 130 feet instead of 60 feet.

(c) Flanking Beds

The bioclastic debris produced on top of the mound while the capping bed is forming gets shed on either side of the bioherm and builds up the flanking beds. These have typical foreset bedding. The foreset beds dip basinwards and sometimes shelfwards.

Tubiphytes can be an important constituent, but is never dominant as in the flanking beds of the lower Wolfcampian bioherms in the Kemnitz Field (Wilson, 1974; Malek-Aslani, 1972).

The flanking beds are cemented by a clotted microspar (irregular shaped patches of micrite in a microspar; "structure grumuleuse" after Cayeux).

The foresets are well-bedded and have a depositional dip of 25°. Their vertical thickness is usually around 30 to 60 feet. The foresets grade basinwards into the slope sediments.

(d) Slope Sediments

The slope sediments are best described as "deeper water" carbonates in the sense defined by J. L. Wilson (1969). The fauna consists of open marine organisms like fusulinids, brachiopods and echinoderms. Cementation is by microspar which is occasionally clotted.

The shelfward part of the slope sediments is strongly influenced by bioclastic debris swept down from the flanking beds.

(2) Cement Tank Canyon

Cement Tank Canyon is an unofficial geographic name; the location of the canyon is shown on Figure 20.

This canyon has been chosen for detailed studies, because the canyon axis is more or less parallel to the depositional dip of the paleoslope, and tectonic complications are only at the canyon head and at the lower end. The south side of the canyon wall has good undisturbed outcrops for about one mile. These are necessary to obtain detailed insight of the complicated facies pattern dominating the shelf edge.

Soft weathering sediments and hard weathering ledges are interbedded on the south wall of Cement Tank Canyon (Plate 1). The soft weathering sediments are either sandstones or deeper water carbonates; the hard weathering ledges are either bioherms, foreset beds or carbonate megabreccias.

One section has been measured up the side of the whole canyon; two ledges and two soft weathering intervals were carefully traced and studied in detail (Plate 1).

(a) Ledge No. 1

A bioherm is at the eastern end of ledge no. 1. The bioherm changes laterally (westward) into foreset beds (Plate 1,A). The latter could be traced westwards. The top of the foreset beds is flat indicating the wave base. The bedding thickness varies between 1 and 3 feet. The microfacies of the foreset beds is well-sorted packstone with occasionally clotted microsparry cementation. The fauna is made up of abundant tubular foraminifera and accessory echinoderms, pellets (worn foraminifera?), Globivalvulina, and Paleotextularia. Minor amounts of "Osagia" onkoids, bryozoa, ostracods, Tubiphytes, platy algae occur.

The foreset beds stopped building outwards (westwards). A small amount of slope sediment (deeper water carbonates) onlaps the last foreset bed (Plate 1,B). The microfacies of these deeper water limestones are wackestones and packstones with the following faunal composition:

Wackestones contain fusulinids, brachiopods, and tubular foraminifera.

Packstones contain abundant sponge spicules and accessory fusulinids, echinoderms, pellets, small pelecypods, gastropods, and Tuberitina.

Above the slope sediments but still on the foreset beds a later bioherm grew (Plate 1,C). A lateral coarse facies of the bioherm could be distinguished (D). This coarse grained facies is time equivalent to the growth of the bioherm below wave base. The fine grained foreset beds (Plate 1,E) are time equivalent to the capping bed, i.e. the part of the bioherm which grew into the wave base. The coarse grained lateral facies of the bioherm is a mixture between the biohermal facies and the slope facies. It is less sorted, coarser grained and bioclasts are less abundant than in the fine grained foreset beds. The fauna consists of tubular foraminifera, echinoderms, Tubiphytes, brachiopods, Paleotextularia, "Osagia" onkoids, and Tuberitina. Minor amounts of Globivalvulina and Paleotextularia occur.

The stopping of the foreset bedding (A) and the onlap of the slope sediments (B) on the foreset beds mark a sharp rise of the sea level and wave base. The deposition of the bioherm (C) is related to a later drop of the sea level.

Above this bioherm (C) is another one. Both of them could be traced basinwards (Plate 1); they change into foreset bedding which is slumped and brecciated (Plate 1,F).

(b) Ledge No. 3

A bioherm is at the eastern end (shelfwards) of ledge no. 3. This bioherm grades laterally (westwards) into a breccia which forms the rest of the ledge. Breccia ledges like this one are very common on the shelf edge and slope of the Pedregosa Basin. They always occur in front of the bioherms. The breccia ledge no. 3 was traced for 1.3 miles toward the basin where it is cut off by a fault at the western end of the canyon. The breccia extended originally further into the basin.

Six detailed sedimentological sections were measured in different places along the ledge. The components of this breccia (clasts) have a size between 1/8 inch and 2½ feet. No grading, either vertical or horizontal was found.

Under the microscope it could be observed that the components were lithified, half-way lithified or soft. Few of the components are coated and some look like "armored mud balls." The boundary between the components and the matrix is in places stylolitic. In only a few examples matrix is lacking and all the boundaries between the clasts are stylolitic. The components derive from the shelf, bioherm and slope environments. Dolomitization has affected mostly the upper part of the ledge.

Burrowing organisms occur sporadically on the surface of the ledge of breccia.

The stratigraphic sequence through the bioherm at the eastern end of the ledge no. 3 offers a clue to the event forming the breccia (Plate 1). This section shows a grainstone just above the bioherm. Above this grainstone are rather shallow marine deposits; but probably these formed in water still deep enough to be below wave base. These deposits do not have typical biohermal characteristics. Above these sediments is a short section of breccia.

The bioherm grew into the wave base which led to the formation of the grainstone above the bioherm. Then sea level rose for a short period and slope sediments were deposited. Later sea level dropped and erosion occurred on the bioherm and the shelf. The Bugle Ridge section has shallow water sediments, grainstones and red stained thin conglomerates (Wilson, oral comm.), possibly indicating erosion on the shelf. Most of the erosion was probably submarine because no real subaerial indications were found.

Tidal currents and wave action could have ripped off previously lithified sediments to form clasts (Plate 3). Currents might have transported them with bioclasts to deeper water, where they were dropped on the foreset bed. Overloading of the foreset bed (depositional slope 25°) and the upper slope resulted in a huge slump of this pile of submarine rubble which also involved the frontal part of the bioherm and the beds above the bioherm. Also the bedded slope sediments on which these masses were dumped became involved and slid down. During transportation, the hard and soft rocks became brecciated. The slope sediments acted partly as a matrix and partly became brecciated.

The clasts in the lower part of the breccia ledge come from the slope and bioherm; the ones from the upper part of the ledge from the bioherm and shelf. This is best explained by slumping of shelf and bioherm masses onto the slope sediments.

(c) Ledges No. 4 and 5

Ledges 4 and 5 are good examples of the distribution of the breccias (Plate 1). The breccias thin out towards the bioherm (onlap) and are underlain by grainstones. This distribution leads again to the previous conclusion about the formation of the breccias, that they are caused by drops of sea level resulting in erosion of the bioherm top. The sea level drop is also the reason for some grainstones on top of the breccia ledge no. 3.

(d) Deeper Water Limestone

The deeper water limestones are in addition to the sandstones the soft weathering intervals of Cement Tank Canyon. Three detailed sections were measured between ledges no. 1 and 2 in the deeper water limestones.

Bed thickness of these alternating limestones, marls and shales varies between 1 inch and 1 foot. Grading and parallel lamination are observed; cross bedding is rare. One bed may contain several intervals of grading. The grading of these intervals is usually incomplete upward, never reaching the pellitic grainsize. Bases of these graded intervals are usually erosional. The fauna consists of in place elements such as sponge spicules, brachiopods and fusulinids and of transported elements like "<u>Osagia</u>" onkoids, pelecypods, platy algae, tubular foraminifera, <u>Tuberitina</u> and <u>Tubiphytes</u>. The size of the bioclasts varies between 5 mm and 0.015 mm. Lithoclasts are rare; their maximum size is about 4 mm. They were derived from sediments with shallow water facies types (grainstones).

The sedimentary structures, the transported faunas and the lithoclasts with shallow water facies types indicate transportation by means of turbidites.

(e) Sandstones

Sandstone is the other lithological type responsible for the soft weathering intercalations in the section. Only one sandstone intercalation occurs in this section.

The sandstone body is channel-like because its top and base are erosional (unconformable) contacts. The erosion of the carbonate beds "on top" of the sandstone channel is due to the fact that the wall of the Cement Tank Canyon is slightly oblique to the channel axis and the channel walls are steeper than the Cement Tank Canyon wall. The carbonate beds "on top" are in fact the ones in which the sandstone filled channel was eroded. The sandstone has interbedded silt and shale. The grain size distribution, sedimentary structures and the environmental interpretations of the channel will be described in the chapter about the Sheridan Canyon which discusses another much larger channel.

(f) Summary and Conclusions of the Facies Pattern on the Shelf Edge and Upper Slope

Plate 1 is a facies cross section through the shelf edge and upper slope in Cement Tank Canyon. The plate is based on a panorama photograph.

Bioherms, flanking beds and deeper water limestones are the normal facies pattern on the shelf edge. The bioherm formation adjusts to minor sea level fluctuation and deposition of them moved up and down the slope. Larger sea level fluctuations had more drastic influences on the facies pattern. Rising sea level results in an onlap of the slope sediments (deeper water carbonates) on to the flanking beds and bioherms. A dropping sea level results in the formation of the megabreccias.

Special sea level conditions resulting in clastic influx will be discussed in the next chapter.

(d) FACIES PATTERN IN A SHELF EDGE LOCATED EMBAYMENT; EXAMPLE: SHERIDAN CANYON

The correlation between the Cement Tank Canyon and the Sheridan Canyon (Plate 2) is based on fusulinids and lithological correlation. The fusulinids were identified by G. Wilde (Exxon Co., Houston, Tex.). The upper Borrego section was described by C. Jordan (1971); it has been slightly revised for this type of work. Also the Bugle Ridge section, which was measured by J. L. Wilson, has been incorporated. Three more sections were measured in addition in Sheridan Canyon. Breccia ledges, deeper water carbonates, etc., were walked out and sampled. A geological map was constructed.

The sections could be divided into seven units (A-G):

<u>Unit A</u>. About 30 feet thick, consists of sandstones and conglomerates with point bar structures. The environment was interpreted as a shallow marine meandering channel. The possibility of a river was excluded, because no effect of subaerial exposure could be found in the surrounding carbonates.

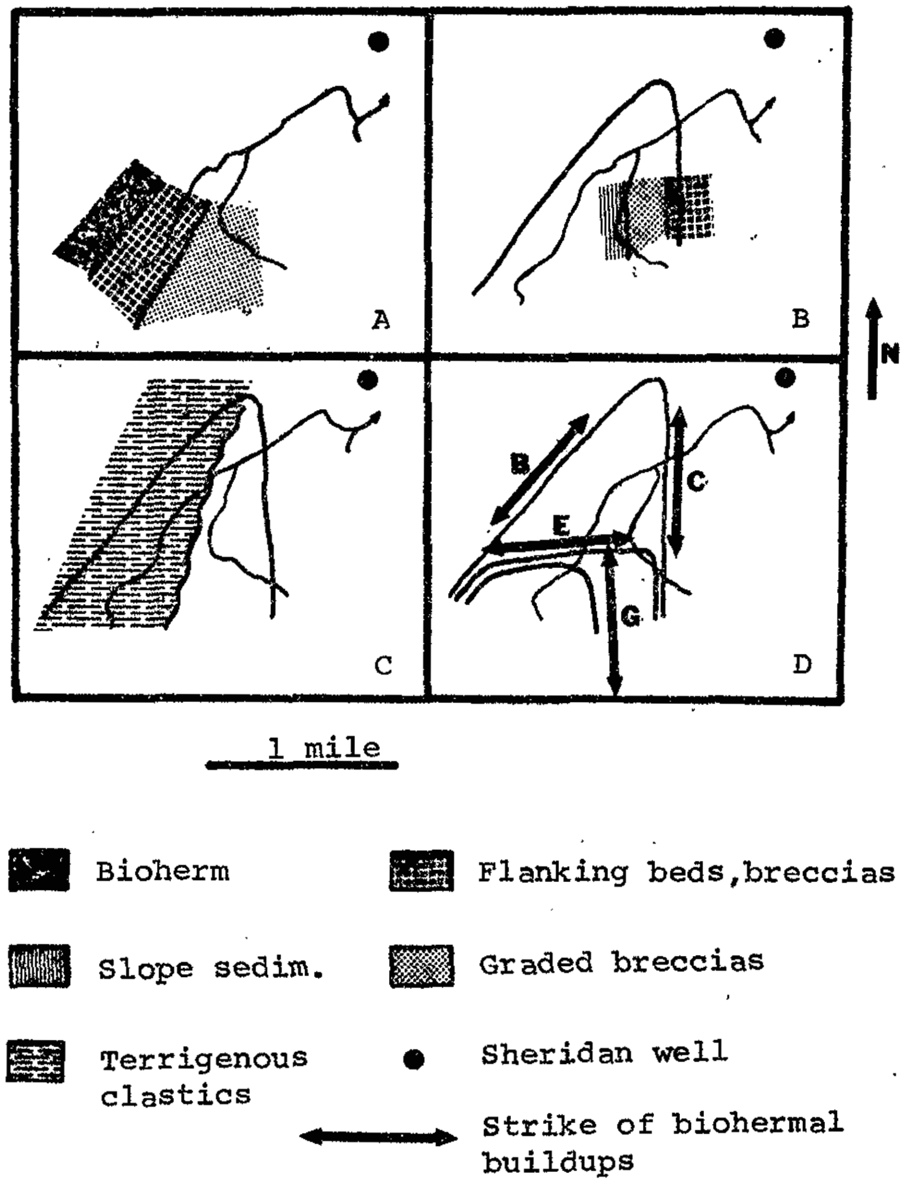


Figure 23: Paleogeography in Sheridan Canyon

A: During stratigraphic unit B.
B: During stratigraphic unit C.
C: During stratigraphic unit D.
D: Strike of biohermal buildups and shape of embayment during stratigraphic units B,C,E and G.

Unit B. About 400 feet thick. Two sections were measured in this unit (sections no. 2 and 3 in Sheridan Canyon). Section no. 3 (Plate no. 2 and Figure 23) consists of interlayered bioherms, foreset beds (flanking beds) and a few deeper water limestones. More deeper water limestones and more and thicker breccias and no bioherms exist in section no. 2, indicating that the regional paleoslope dips to southeast. The same paleoslope direction is given by the orientation of the bioherms in section no. 3: SW-NE (long axis of bioherms).

Some of the breccias in section no. 2 have an uppermost graded layer. The clast size in this graded bed varies between coarse sand and silt. Breccias with a graded layer on top are explained as "debris flows" which generated a turbidite. The turbidites continue further downslope than the breccias. This mechanism can explain at least some of the turbidites on the slope and the basin. It is assumed that the breccias with a graded layer on top travelled for a longer distance than the one without this layer. If the breccias with a graded layer on top in section 2 have come from the eastern end of the embayment, the transportation distance would be about 2 miles. The breccias without the graded layer might come from the northern and southern edge of the embayment, and their travel distance would be only 1 mile.

Unit C. 300 feet thick. The unit consists of well-bedded, dark deeper water limestones. These are graded and may have up to 15% of quartz silt. Parallel lamination is common, in contrast to the rare cross bedding. Some of the beds are extremely burrowed. The cementation is microspar which is occasionally clotted; dolomite rhombs are rare.

A breccia with a graded layer on top occurs in the middle part of the unit. The breccia could be traced eastwards into the equivalent bioherm. The embayment edge therefore lay east of section no. 2 (Figure 23).

The upper part of the unit contains additional deeper water limestones, which are coarser than the ones from the lower part of this unit. Beds with oriented fusulinids and rich in crinoidal debris are common here.

Unit D. This unit contains terrigenous sandstones and siltstones with the same sedimentary structures as the sandstones in the channel of Cement Tank Canyon. The bedding is 1-2 inches thick. Each bed contains either cross bedding or parallel lamination. Seldom do both structures occur in one bed; in the rare cases in which they do, the parallel lamination is always lower and the cross bedding lies above. This combination of sedimentary

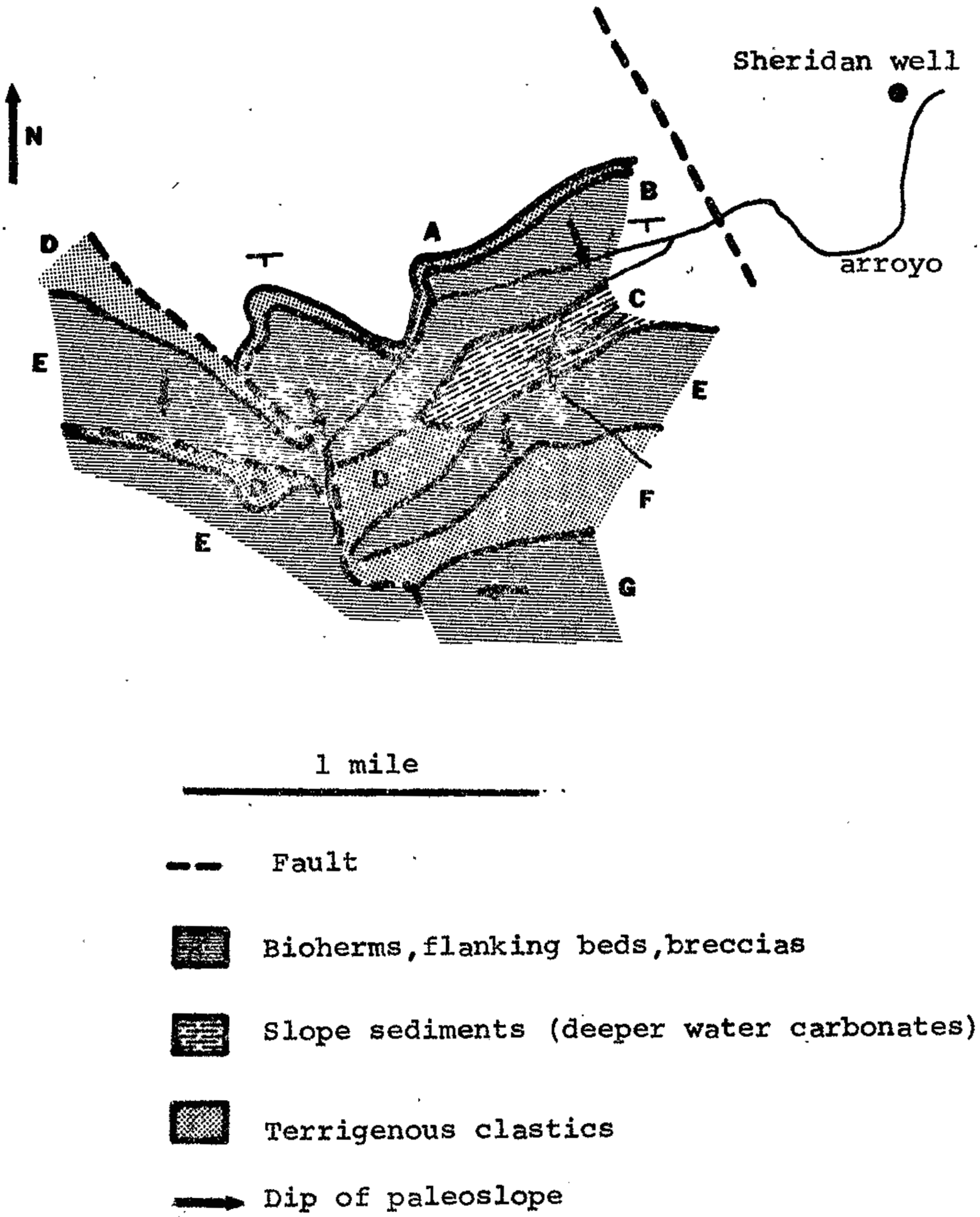


Figure 24: Geological map of the Sheridan Canyon area.

structures is similar to those in modern delta front or prodelta environments.*

The shape of the sand body is that of a channel (Plate 2 and Figure 23). The channel is eroded into unit C and partly into unit B.

Unit E. This unit is a bioherm. The orientation of the bioherm is E-W (Figure 23).

Unit F. This unit (Figure 24) is a terrigenous clastic section, like unit D. The unit wedges out towards the shelf; the base is not eroded. The fine grained texture (siltstone and shales) and the geometry indicates that the channels funneling the clastics for this unit were someplace else on the shelf edge and not in the Sheridan Canyon area.

Unit G. This unit (Figure 24) is also a bioherm. The orientation is N-S (Figure 23).

(1) Conclusions Concerning the Facies Pattern in the Shelf Edge Embayment at Sheridan Canyon

The rock types and the formation of the breccias are the same as on a normal shelf edge (Cement Tank Canyon). The influx from the breccias was from all three sides of the

* Oral communication by C. Campbell, Esso Production Research Co., Houston, Texas.

embayment, because the embayment was surrounded by bioherms. Owing to the location of the measured sections one may observe some breccias which traveled farther downslope than in the Cement Tank Canyon. These more distal breccias have a graded layer on top of them.

Sheridan Canyon also gives a clue to the understanding of the terrigenous clastics. In times of relative low sea level stand the clastics had time enough to bypass the shelf in shallow water channels (unit A); only a coarse lag deposit stayed behind on the shelf. Channels were eroded at the shelf edge and became pathways for the clastic influx which funneled down the slope and was shed into the basin (unit D in Sheridan Canyon and the sandstone body in Cement Tank Canyon). The clastics in the basin are described in the following chapter concerning the well, "Exxon no. 1 N. M., "BA." The influx of terrigenous clastics occurred only during low stands of sea level, as did the formation of the breccias. Theoretically one should find carbonate breccias just below the sandstones. However, the sandstones never crossed the shelf during times when bioherms were at the shelf edge, but always after times of deeper water deposition on the shelf edge (onlap of slope sediments on bioherms and flanking beds). The evidence bearing on this is limited to a small number of examples of clastic influx observed on the shelf edge.

The timespan of low sea level which led to the formation of the carbonate breccias was never long enough for the clastics to cross the shelf. The clastic sediments probably derived from the Burro-Florida uplift lying east of the Big Hatchets. However, only one example of lag deposits on the shelf was observed (unit A plate 2). No other lag deposits were observed in all the other sections. Three possibilities seem to explain this:

(1) Strike slip faults offset the shelf from the shelfedge (see p. 84).

(2) The sections on the shelf failed to cross a clastic deposit because of the scarcity of sections and the stringlike configuration of clastic deposits.

(3) The clastic influx may have come from a northerly direction instead of easterly.

The last explanation seems to be the most logical one because the canyon in the Sheridan embayment (unit D, and arrow, plate 2) has a northerly direction. Small channels (unit A, plate 2) might also have had a northerly direction originally but turned at the shelf edge to flow parallel to the dip of the slope.

Figure 23 shows the paleogeography and filling of the Sheridan embayment. The embayment is a preferred location for the clastics to cross the shelf edge and slope (units D,A).

The filling of the embayment by clastics as well as by carbonates results in a swing of the bioherm orientation from the different directions along the embayment edges back to the regional direction (Figure 23).

(e) BASINAL FACIES (EXXON NO. 1 NEW MEXICO STATE "BA")

In 1958 Exxon Co. (then Humble Oil and Refining Co.) drilled a wildcat well about 7 miles southwest from Sheridan Canyon (plate 2). Electric logs and cuttings were made available by the Bureau of Mines and Mineral Resources of New Mexico. In addition, core chips were provided by Exxon Co. (Midland). The well penetrated the whole Horquilla Formation. Detailed petrographic studies were executed between 7250 and 9000 feet. The well was correlated with the measured sections of the Big Hatchet area with fusulinids (determined by Garner Wilde) and with lithological units (plate 2).

Detailed petrographic studies reveal that there are carbonate turbidites containing fragments of shallow water facies types (grainstones) as well as thick intervals of terrigenous clastics such as silt, very fine sandstones and shales.

(f) CONCLUSIONS AND INTERPRETATIONS OF THE MARGINAL CRATONIC BASIN (PEDREGOSA BASIN)

The geographic position of the shelf edge is defined by the position and orientation of the bioherms and their flanking beds.

Typical slope indicators are the carbonate breccias which extended at least for 2 miles into the basin at Sheridan and Cement Tank canyons. Their distal portions have graded layers at their tops. The most interesting phenomena on the slope are the erosive channels which are backfilled with sandstones and silts having delta front type structures. In contrast, the basin contains thin units of carbonate turbidites and thick units of terrigenous clastics like shales, silts, and fine sandstone.

The normal facies pattern consists of a bioherm at the shelf edge, foreset beds (flanking beds) basinwards and in places shelfwards of the bioherms and deeper water sediments on the slope and basin (plate 3). When sea level rose, the deeper water sediments lapped onto the flanking beds and the bioherm. As sea level dropped, the carbonate breccias formed. When sea level remained low for some time, terrigenous clastics had time to cross the shelf and erode the channels as they were shed into the basin. When sea level began to rise again or when the basin was about to become filled, the channels filled up backwards with sand and silt having delta front structures.

In summary, minor sea level fluctuations resulted in shifting of the bioherm and flanking bed complex up and down the slope for a distance of at least one mile. Major sea level fluctuations resulted in:

(1) formation of the carbonate breccias during short periods of low sea level stands;

(2) deposition of terrigenous clastics during long periods of low sea level stand;

(3) onlap of the slope sediments onto the bioherm and flanking beds during rise of sea level.

The deposition of clastics and carbonates was always separate and occurred at distinct time intervals. The carbonate sedimentation is characterized by thick sedimentation on the shelf and shelf edge and thin sedimentation in the basin. The geometry of the clastic sedimentation is reversed: thick in the basin and thin or absent on the shelf (plate 4); i.e., the basin deepened and developed during carbonate sedimentation and filled up during the clastic sedimentation.

Subsidence in the basin had to exceed subsidence on the shelf: Periods of clastic deposition completely filled the basin and at the end of such a time little relief existed between the shelf and basin. During subsequent carbonate sedimentation, deeper water sediments (dark shale and limestone) consistently deposited in the basin and shallow water sediments on the shelf. This persistent facies pattern was controlled by subsidence which was stronger in the basin. Subsidence prevailed also during terrigenous clastic periods but the sedimentation rate exceeded subsidence and the basin filled.

RESERVOIR ROCKS

(1) Pedregosa Basin

There are three obvious possibilities for reservoir rocks:

(a) Channeled sandstones on the slope

Probably the best reservoir rocks in the Horquilla Formation are the channels with the terrigenous clastics on the slope. The channels are filled with fine sandstones; sometimes the sandstones grade into silt in the upper part of the channel. The channels are eroded into deeper water limestones which do not have any porosity in the Big Hatchet area. The channels are covered by the next depositional sequence which is a bioherm-flanking bed complex with scattered carbonate breccias. None of these rock types has any porosity except for the bioherms (see next paragraph). If the bioherms have porosity the reservoir will be expanded.

The channel pinches out on the shelf edge or grades into a thin string of coarse lag deposits with point bar structures.

Due to the slowly rising sea level the channels filled in backwards (i.e., from the basin to the shelf). The basinal section in the Exxon well has relatively few sandstones. It is, therefore, assumed that during the deposition of the terrigenous clastics during low sea level stand, a cone of sandstone formed at the distal end of the channel, at the foot of the slope, and did not extend far into the basin.

The two channels observed on the slope are 50 to 400 feet thick, at least a mile wide, and several miles long.

Similar reservoir types occur in the Midland and Delaware Basin. The channels are potentially good reservoirs (Galloway and Brown, 1973).

(b) Leached Bioherms

The porosity in bioherms is due to subaerial exposure of the bioherms— leaching and collapse brecciation at an early stage in their history. This type of reservoir rock exists in the Ismay Field (Colorado, Utah). The bioherms in the Big Hatchets area are well cemented; if original or secondary porosity was present, it has been cemented, probably during Tertiary uplift by phreatic water. Porosity due to leaching could have been preserved elsewhere along the Pedregosa shelf edge in the subsurface.

(c) *Tubiphytes* Flanking Beds

The porosity in these flanking beds is due to internal openings in the *Tubiphytes*. The Kemnitz oil field has this type of porosity (Malek-Aslani, 1970). *Tubiphytes* beds are not common in the Horquilla Formation in the Big Hatchets area, except for unit G.

(2) *Piz Alv—Piz Mezzaun Area*

Possible oil reservoirs could occur in similar rock sequences, but only in less deformed areas. Possible reservoir rocks are:

(a) *Fractured radiolarian chert*.

(b) *Leached breccias and horsts*. Subaerial leaching was observed in the breccias of the Alv Formation and is, of course, obvious in the horsts. However, all leached cracks were cemented early (vadose and submarine cementation; see p. 15) in this region.

(c) *Turbidites*. A possible reservoir rock could be a series of coarse grained and matrix-poor turbidite wedges at the base of the slope where only rare pelagic sedimentation occurs.

Source rocks for all reservoirs would be the shale in the distal parts of the basin or the shales in the upper part of the stratigraphic section.

COMPARISON BETWEEN THE MARGINAL CRATONIC BASIN AND THE MARGINAL GEOSYNCLINAL BASIN

The marginal cratonic and marginal geosynclinal basins are compared (or contrasted) in terms of parameters which control sedimentation. These parameters are:

(1) Megatectonic environment which controls geomorphology and rate of subsidence.

(2) Climate.

(3) Sea level and its fluctuations.

Sedimentation patterns are considered as follows:

(1) Basin geometry.

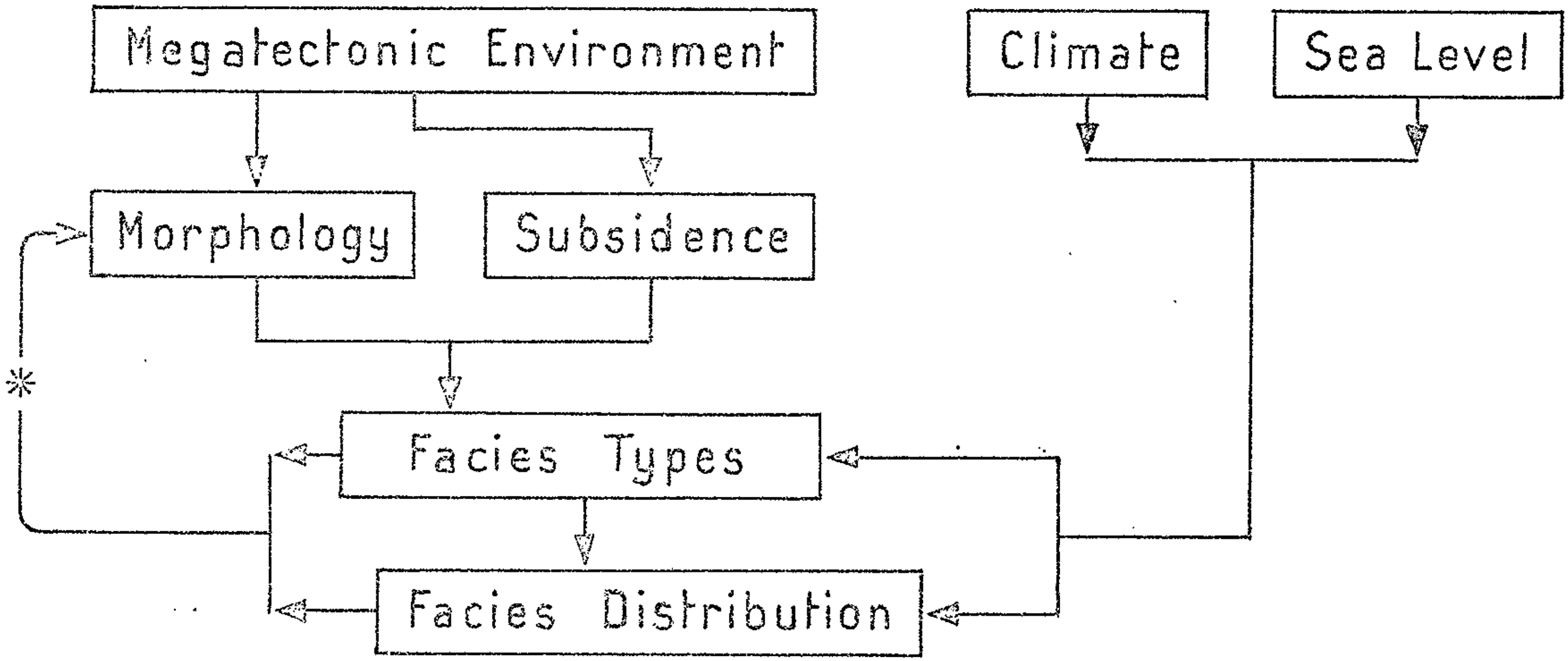
(2) Sedimentation and filling of the basins ?

(3) Geometry of the facies types.

A comparison between the two basin types is demonstrated in Table I. Differences in similar rock types are added in a special table (Table II). These differences between similar rock types, as well as the general differences mentioned in this chapter, are important to distinguish these two basin types when working with only limited subsurface samples.

(a) MEGATECTONIC ENVIRONMENT

(1) <u>Piz Mezzaun-Piz Alv Area (Lower East Alpine Nappe)</u>



* Modifications of morphology by facies types and their distribution (as example reef growth)

Figure 25 : Flow diagramm showing influence of the megatectonic environment, climate and sea level on sedimentation.

The north Atlantic opened first in early Jurassic time (Pitman and Talwani, 1972), and this opening resulted in divergent, left, lateral wrench faulting between Europe and Africa. The wrench faults passed either through the future Western Alps or south of Italy (Hsu, 1972). The latter case would involve a rotation of Italy which would result in secondary divergent wrenching in the Western Alps.

Divergent wrench faults have a component of extension which is similar to pull-apart structures. This extensional component moves apart the two sides of the fault, and this produces a tension, which in turn results in normal faulting. A recent example of a pull-apart structure is the Red Sea, which has been described by J. B. Lowell and G. J. Genik (1972). They show normal faults and horst and graben structures at the newly formed continental edges and the formation of oceanic crust in the "graben" of the Red Sea.

Wide spread shallow water carbonate deposition on continental crust persisted during the Triassic in southeastern Switzerland. During early Jurassic time geosynclinal conditions began to "form". Figure 1 is a late Jurassic-early Cretaceous paleogeographic reconstruction. One continental block is present to the north (Helvetic realm), one to the south (East Alpine and South Alpine Realms) and a small continental block is in the middle (Brianconnais). These three blocks rifted apart during the Jurassic and formed "grabens" (geosynclinal basins) between them. The block boundaries are characterized by normal faults. It is commonly believed that all newly formed continental edges are characterized by normal faults.

The conclusions of Talwani and Pitman, the facies development in the Lower East Alpine realm over platforms as horsts and basins as grabens, and the synsedimentary structures, i.e. normal faults, separating the two, suggests that the Jurassic tectonics in the Western Alps can best be described as a pull-apart structure related to the opening of the Atlantic. It is not yet possible to demonstrate whether or not lateral displacement is present in the western Alps.

The basin of the Piz Mezzaun-Piz Alv area is on the edge of the southern continental block (East Alpine and South Alpine realm). The basin can be termed either a marginal geosynclinal basin or a basin marginal to a pull-apart structure.

The Jurassic sediments in the Piz Alv-Piz Mezzaun area are related only to the tensional phase and not to the later compressional phase which lead to the Alpine orogeny.

(2) <u>Pedregosa Basin</u>

The Pedregosa Basin is a marginal cratonic basin. The entire basin is still on continental crust and connects to the problematic Chihuahua Trough to the south. This trough is a foredeep of the Ouachita Marathon geosyncline or even the external part of it. In the latter case it might be offset laterally to the northwest either by post Paleozoic wrench faults or by late Paleozoic transform faults (Figure 21). The Pedregosa Basin, as well as the other basins and platforms in this area, are approximately parallel to each other but oblique to the Ouachita Marathon trend. The parallelism might be due to basement faulting and perhaps to possible late Paleozoic transform faults, which had a subparallel orientation.

(b) SYNSEDIMENTARY TECTONISM

There might have been basement faulting in the Pedregosa Basin which, however, did not offset younger sediments (see last chapter) although it controlled rate of subsidence and types of sediments deposited. The Piz Alv area is characterized by active normal faulting <u>during</u> sedimentation; it is possible that strike slip faults also existed, but these are no longer recognizable in the field. Vertical offset on some of those faults must have exceeded 1500 feet. There was also a rapid up and down movement of some horsts; one of them was eroded to the crystalline basement and overlain by turbidites, with no intermediate facies. The Alpine basin, regionally an area of almost geosyclinal subsidence, displays clearly stronger tectonic activity.

(c) CLIMATE

The climate in the region of the Piz-Alv-Piz Mezzaun Basin was semi-arid with seasonal rains. An indication of aridity is the red colored (oxydized) dolomitic matrix of the subaerial breccia (Alv Formation). Vadose leaching and cementation are evidence for seasonal rain fall. The Liassic equatorial belt trends through southernmost Europe and Morocco.

The climate in the Big Hatchet Mountains was similar: Evaporites in the Earp Formation indicate aridity. Caliche crusts (vadose cementation) were discovered by P. Winchester (oral communication)in Wolfcampian beds of the Sacramento Mountains (southeast New Mexico). The climate in the southern part of the United States changed during the depositional time of the Horquilla Formation.

Pennsylvanian time is characterized by a warm and humid climate, but Upper Wolfcampian by an arid climate. The climate during the intervening late Virgilian and early Wolfcampian time was in general probably simi-arid with seasonal rains.

(d) SEA LEVEL AND ITS FLUCTUATIONS

Sea level fluctuations are an important parameter in the sedimentation of the Pedregosa Basin:

(1) Minor fluctuations produce shifting of facies types up and down the slope.

(2) Major fluctuations cause formation of breccias, onlap of deeper water carbonates onto the shelf edge and deposition of the clastics in the basin.

Sea level fluctuations were not an important feature in the Piz-Alv— Piz Mezzaun Basin. Only the breccia of the Alv Formation might suggest sea level fluctuation (leaching in the vadose zone, cementation in the phreatic zone), but this effect is probably due mainly to vertical tectonic fluctuations.

Evidence of sea level fluctuations may well be obscured by uplift and subsidence in a basin with active tectonism. Shelves which might be affected by sea level changes are narrow and inconspicuous. Sedimentation generally occurs in basin, subsiding so rapidly and with water so deep that eustatic sea level changes did not affect sedimentation very much.

The uppermost Triassic deposits were shallow but open marine sediments. The Piz-Alv— Piz Mezzaun Basin formed by tension which resulted in graben structures (p. 117). The horsts should still have shallow marine sediments on top of them. However, many of them reached above sea level. A drastic down drop of sea level simultaneous with the forming grabens is responsible for this subaerial exposure of the horsts.

(e) BASIN GEOMETRY

As mentioned above, the shape of the Pedregosa Basin is elongate, more or less parallel to the other marginal basins, but oblique to the Ouachita Marathon geosyncline.

In the Alps, the zone of Jurassic tectonism was parallel to the Alpine trend.

(f) SEDIMENTATION AND FILLING OF THE BASIN

The Pedregosa Basin is characterized by a reciprocal sedimentation of clastics and carbonates. Carbonates build up thick sequences on the shelf, at the shelf edge (bioherms) and thin sequences in the basin (deeper water carbonates, turbidites). The difference in elevation between the basin and the shelf was probably near 400 feet when carbonate sedimentation ceased.

The clastic deposition, which always alternated with carbonate deposition, is characterized by very thin or no deposits on the shelf and thick deposits in the basin. The terrigenous clastics filled the basin, and no topographic difference existed between the shelf and the basin after clastic deposition and before the next carbonate deposits were laid down. The water depth after the carbonate deposition was approximately 500 feet in the basin, i.e. the approximate thickness of intervening terrigenous units between the limestones. During transgressions when deeper water carbonates were deposited on the shelf the water was of course deeper. Tidal flat sediments (Earp Formation) lie above the basinal sediments of the Pedregosa Basin.

The process of reciprocal sedimentation did not occur in the Piz-Alv-Piz Mezzaun basin. Deposition of all facies types occurred simultaneously. But the sequence is characterized by a general deepening of the sea and decreasing influx of detritus. This is reflected in the sediments by a vertical change from turbidites into radiolarian cherts.

The water dapth of radiolarian chert is assumed to be approximately 12,000 feet for recent conditions, but was probably much less in Jurassic time. A thin layer of Cretaceous sediments (deep sea marls; Couches Rouges Formation) overlies the radiolarian chert. Upper Cretaceous (?) — Lower Tertiary Flysch deposits lie above the Couches Rouges in the Piz Alv-Piz Mezzaun basin. These Flysch deposits belong to another tectonic phase (compressional phase and not tensional rifting) of the Alps.

The two basins differ significantly in their sedimentary history. The Pedregosa Basin became filled several times during its existence (always at the end of a period of clastic sedimentation); it also became completely filled at the end of its existence and it is overlain by continental sediments, documenting that in Middle Wolfcampian time the basin had filled up and that subsidence had almost ceased.

The Piz Alv-Piz Mezzaun basin on the other hand became continuously deeper and never filled up, and the turbidites grade upward into radiolarian chert.

(g) GEOMETRY OF THE FACIES TYPES

(1) <u>Pedregosa Basin</u>

The shelf edge is characterized by bioherms (about 70 feet thick, 100 feet wide). The shape of the bioherms is generally that of a bread loaf, with the long axis parallel to the shelf edge. In front of the bioherms are flanking beds with depositional surface dips of about 25°. These have the same thickness as the bioherms and extend for about 1000 feet basinwards. The breccias, which are also in front of the bioherms, are about 30 feet thick and extend for at least 2 miles into the basin. The lateral extension of the breccias is unknown because of lack of control, but is probably in the order of ¼ to ½ mile. The bioherms and the related facies types are seen to be stacked upon each other vertically. Bioherms can migrate basinward due to outbuilding of sediments or due to lowering of sea level (true regression). Bioherms might also migrate up dip during small transgressions. Larger transgressions are characterized by deposition of deeper water sediments on the slope and shelf edge and a lack of bioherms. Sand occurs in thin string-like deposits on the shelf, in channels (50-400 feet deep and 1 mile wide) on the slope and in cone-like buildups at the foot of the slope. Silt and shale are the predominant particle size filling the basin.

(2) <u>Piz Alv-Piz Mezzaun Basin</u>

The breccias of the Alv Formation built cones along the horsts boundaries. The lower part of the cones extended into the marine environment. The breccia of the Fain

Formation (which is the marine equivalent of the Alv Formation) has a more sheet-like form. The turbidites are also sheet-like. Submarine fans of the "Californian" type are missing, because no point source is present. Sources are scattered all along the continental edge (horsts).

(h) CONCLUSIONS

Three factors control the depositional history of basins:

I. Megatectonic environment

II. Climate

III. Heights of sea level

The flow diagram of Figure 25 explains how the three factors control sedimentation (facies types and their distribution). Morphology is sometimes modified by sedimentation as, for example, in the case of reef growth.

The difference of the depositional history of the two basins in this report is a result only of the different megatectonic environment and the height of sea level and its fluctuations. The climatic factor is eliminated because the climate was the same in both regions.

This thesis hopefully describes the rules applicable to understanding and predicting sedimentation in different basins by varying the three factors. An example might illustrate this: The megatectonic environment of the present Red Sea is the same as in the Piz Alv-Piz Mezzaun area, but the difference is negligible. However, the height of sea level was drastically different: The opening of the Red Sea was "dry". Either the Red Sea trough was considerably above sea level or the trough was partly blocked off from the seas. The basin of the Piz Alv— Piz Mezzaun region opened "wet"; with the exception of some horsts the newly formed basin was under water. In this comparison megatectonic and climatic factors are the same; only the sea level factor was different. But the differences in this single control had a strong influence on the sedimentation: Turbidites, submarine breccias and some few subaerial breccias characterize the "wet" Piz Alv— Piz Mezzaun Basin. Evaporites, continental clastics with some few marine incursions characterize the "dry" Red Sea Basin. The modern Red Sea rift is now partly open to the sea and filled with water.

GENERAL COMPARISON (Table I)

	Big Hatchet Mts. (New Mexico)	Piz Alv-Piz Mezzaun Area (SE Switzerland)
Basin	Pedregosa Basin (permo-Pennsylvanian)	Southern margin of Piemontais Geosyncline (Jurassic)
Megatectonic environment	Marginal cratonic basin	Marginal geosynclinal basin
Synsedimentary tectonism	Basement faults (?), no offset of Permo-Pennsylvanian sediments (parallel to Pedregosa Basin, but oblique to the Ouachita-Marathon orogenic trend)	Active normal faults (parallel to the basin and geosyncline)
Climate	Semi-arid, seasonal rainfall	Semi-arid, seasonal rainfall
Sea level and its fluctuations	Major influence on sedimentation: shifting of facies. Formation of breccias. Sedimentation of clastics.	Minor influence

Table I, cont'd

	Big Hatchet Mts.	Piz Alv-Piz Mezzaun Area
Basin geometry	Marginal cratonic basins in this area are parallel to each other, but oblique to the Ouachita-Marathon trend.	Basin is probably parallel to geosyncline and Alpine trend. Lateral extent of basin is unknown.
Sedimentation and filling of the basins	Thick carbonates on the shelf and shelf edge, thin carbonates in the basin. Clastics are absent or sparse on the shelf and thick in the basin. Basin formed during carbonate deposition, but filled up during clastic deposition. On top of the basin: near-shore continental sediments.	One sequence of sedimentation. Basin never filled up. On top: deep radiolarian chert.

TABLE I, cont'd

	Big Hatchet Mts.	Piz Alv-Piz Mezzaun Area
Geometry of facies types	Bioherms (70 ft thick, — 100 ft wide, and probably several hundred feet long). Flanking beds have same thickness as bioherms, extend about 1000 ft basinward. Breccias (30 ft thick) extend for at least 2 miles basinward; lateral extension ¼ to ½ mile. Sandstone channels on the shelf are 50-400 feet thick, 1 mile wide, several miles long. Sandstone on shelf is in stringlike meandering belts.	The subaerial breccia occur in cones or strings along the horsts. Submarine breccia occur in bands adjacent to the subaerial breccia. The turbidites are sheet-like.

SIMILAR ROCK TYPES (Table II)

	Marginal Cratonic Basin (Pedregosa Basin, N.M.)	Marginal Geosynclinal Basin (Piz Alv-Piz Mezzaun Area, Switzerland
Breccias	Submarine slump breccias. Marine fauna intermixed. Components: contemporaneous with formation of breccia. Size: sand to several feet across; components are hard or soft.	Subaerial breccias: vadose and submarine leaching and cementation. Submarine breccias are bedded, are of only a few components; marine fauna. Components: older than formation of breccia. Size: sand to megablock size (1200 ft long).
Turbidites	Bouma's "c" and "d" intervals. Clasts derive from contemporaneous sediments; contemporaneous fossil debris. Bedding about 1 inch.	Bouma's "a" and "b" intervals. Clasts derive from older formations; contemporaneous fossil debris. Bedding 1 foot and thicker. Sandstone turbidites: delta front structures in channels on the slope (grain-size; fine sandstone). Silt and very fine sand dominate in the basin.

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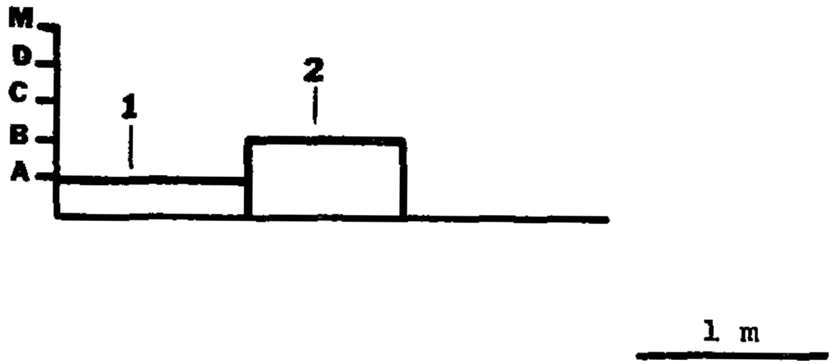
KEY TO PLATE D



On the vertical axis are the interval types;
on the horizontal axis is the thickness of the interval.
As example:

1. Interval of type A, 1 m thick.
2. Interval of type B, 0.8 m thick.

Upwards in the section is from left to right in each row. The highest row starts at the base of the section (or unit); the lowest row ends with the top of the section (or unit).

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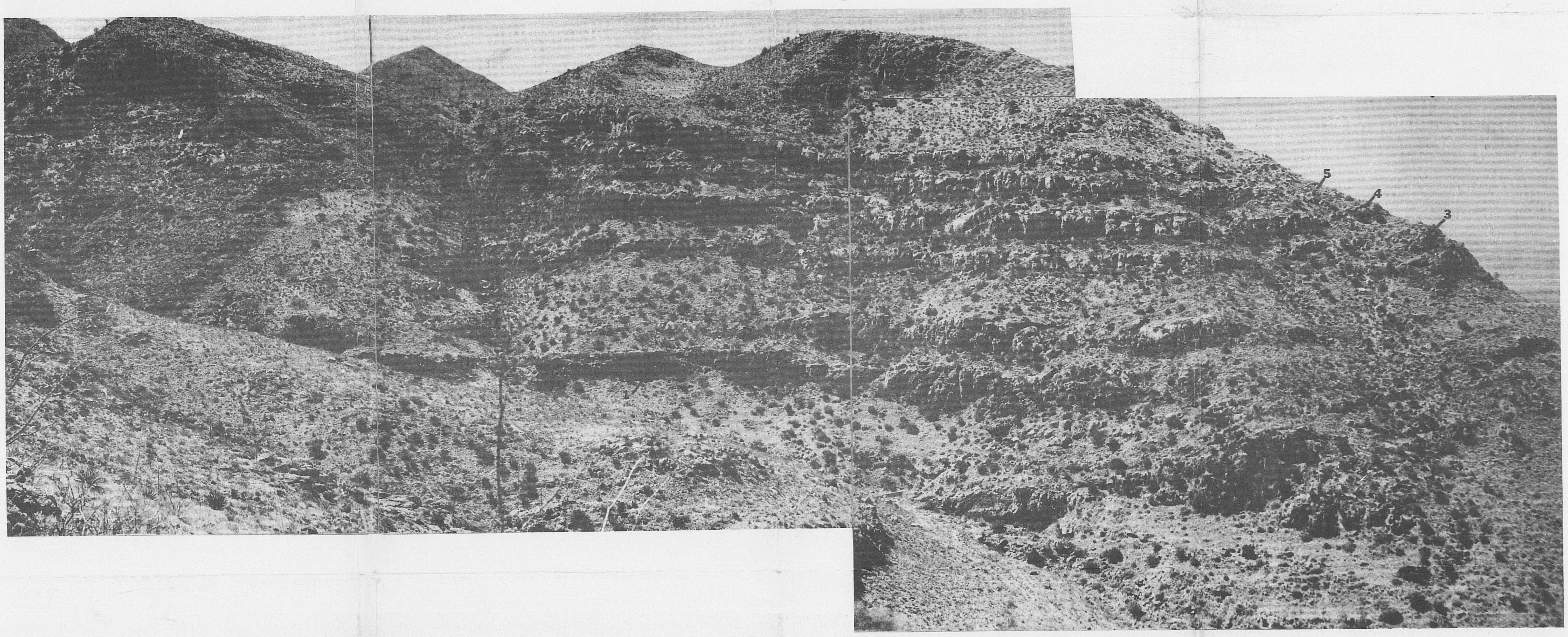
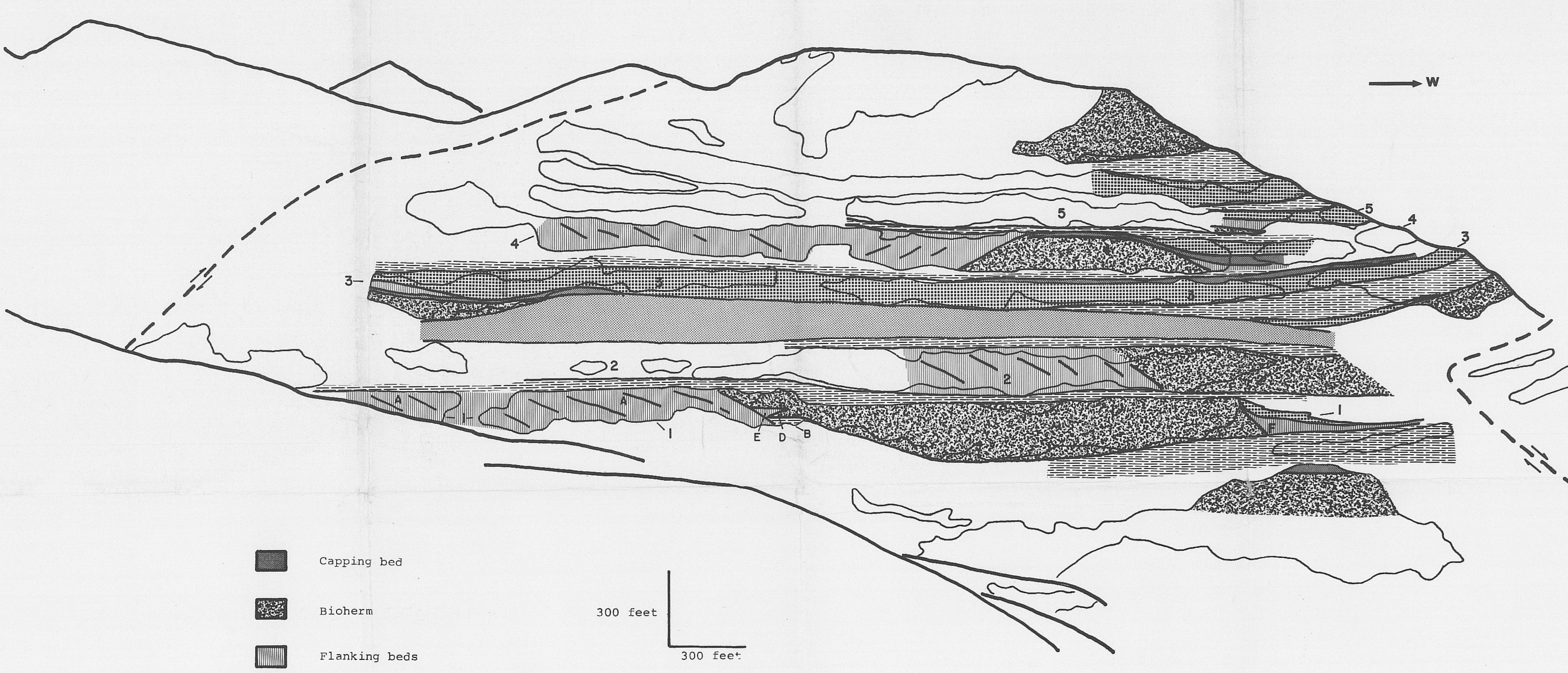
Plate 1

FACIES DISTRIBUTION OF THE SHELF EDGE, CEMENT TANK CANYON.

(UPPER VIRGILIAN-LOWER WOLFCAMPIAN)





Capping bed

Bioherm

Flanking beds

Slope sediments

Carbonate breccia

Terrigenous clastics

Fault

The Perdregosa Basin is on the right (west) and the Florida Shelf to the left (east).

The ledges,which can be observed on the photograph,are outlined on the diagram below. The numbers refer to certain ledges mentioned in the text. The letters are features in ledge no.1 and are explained in the text.

Plate 1.

FENCE DIAGRAM OF UPPER VIRGILIAN AND LOWER WOLFCAMPIAN STRATA IN THE BIG HATCHETS

PLATE 2

M.A. SCHUEPBACH

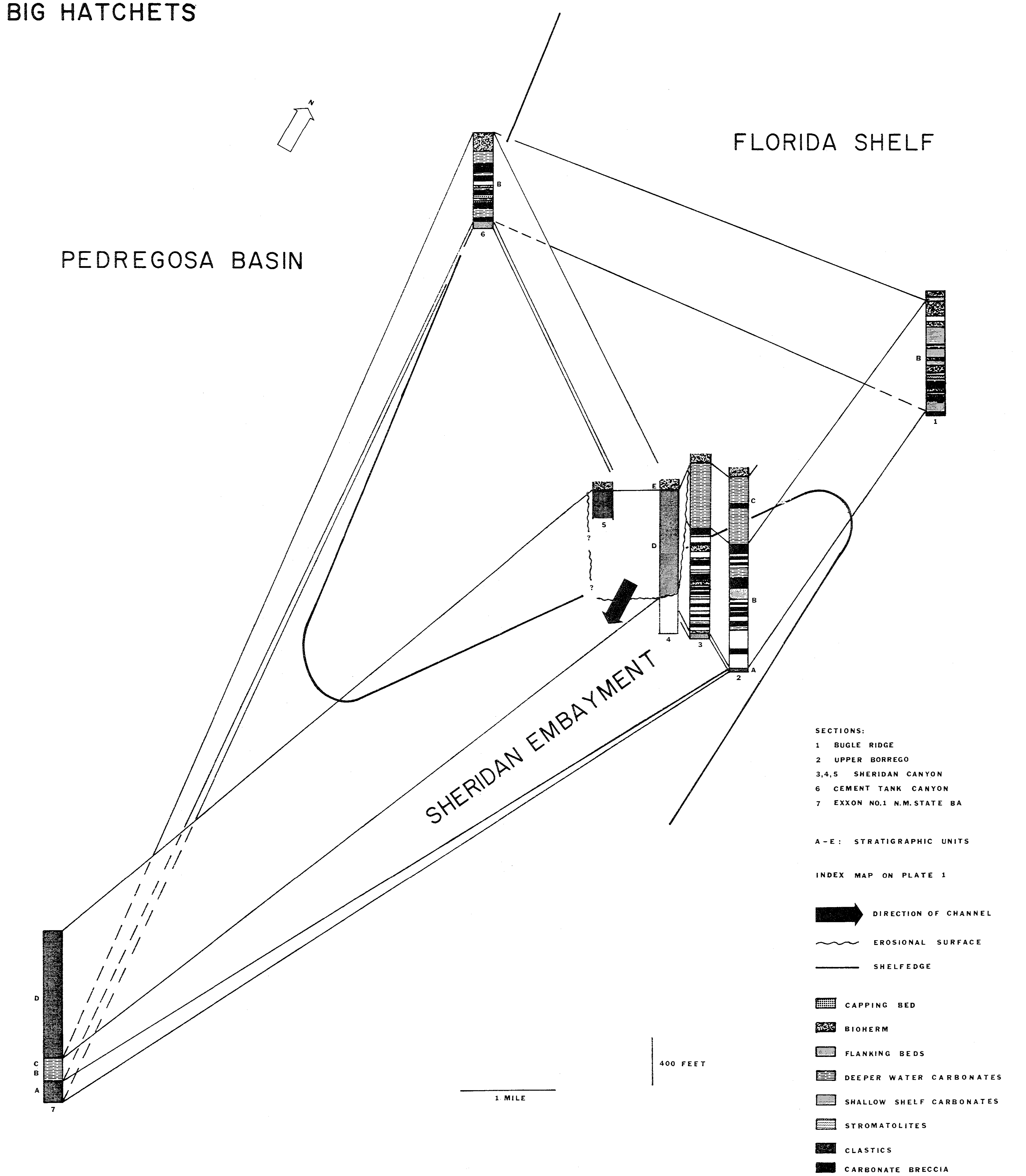


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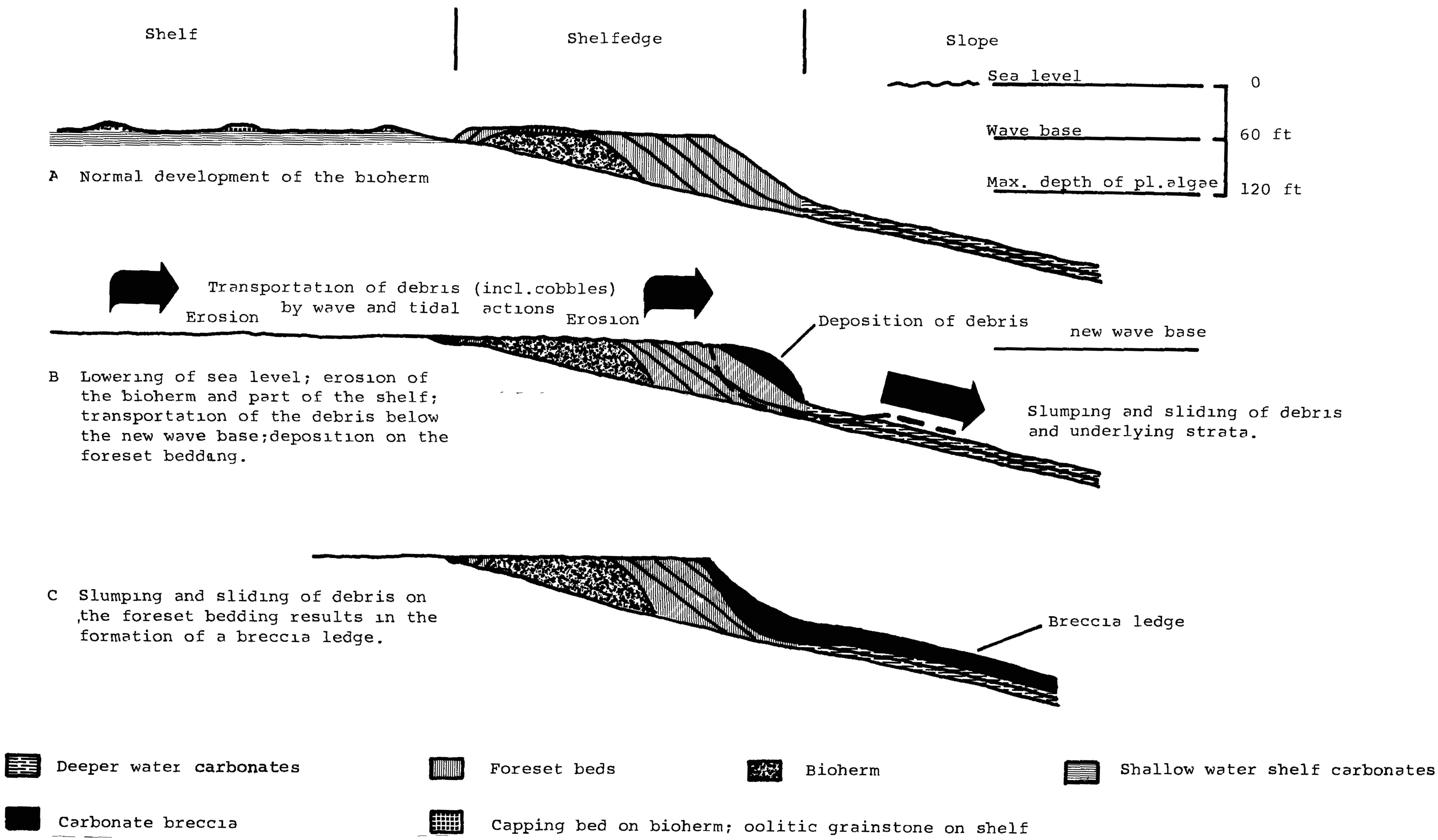
Plate 2

FORMATION OF BRECCIA LEDGES

PLATE 3



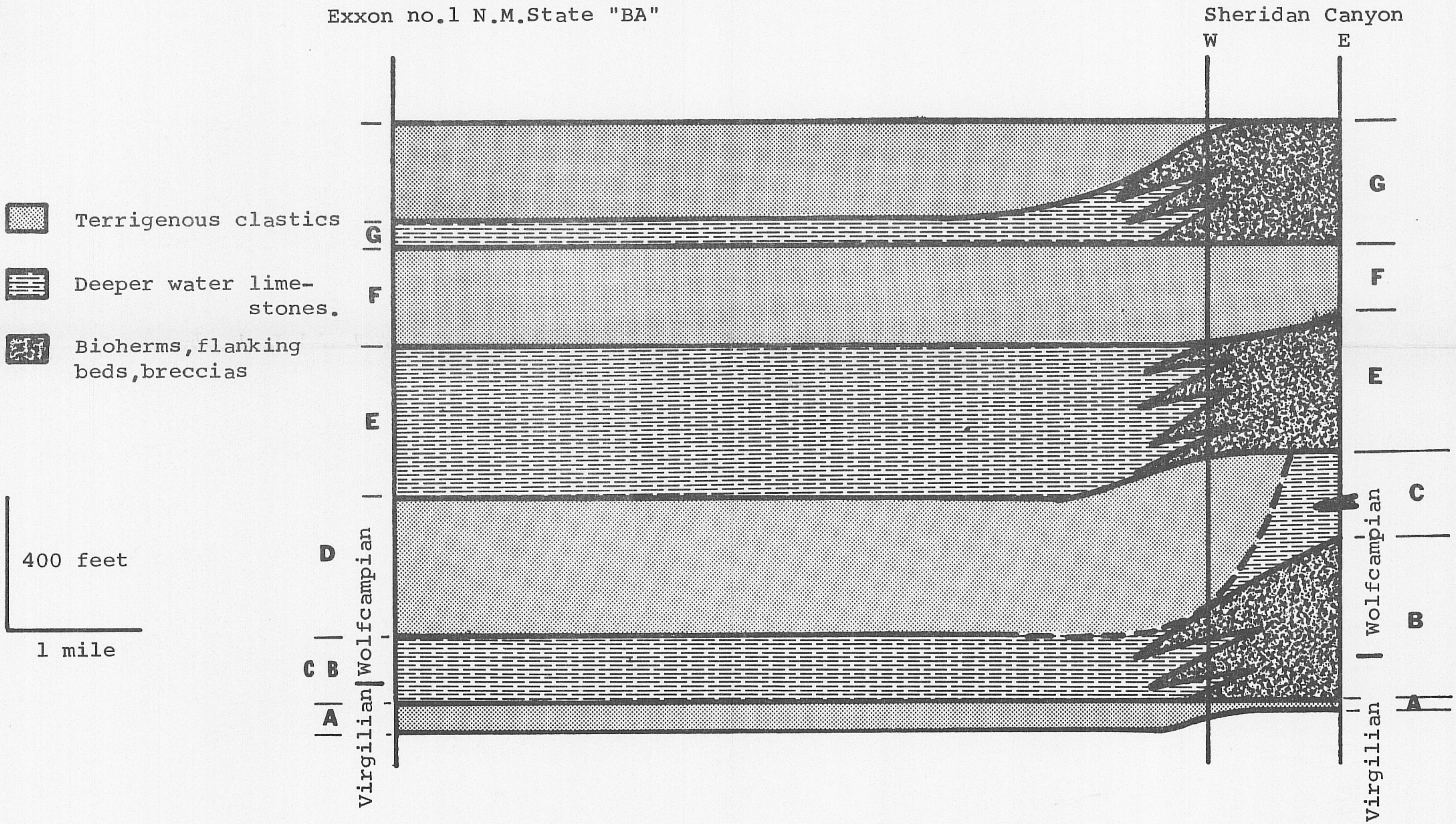

Exxon no.1 N.M.State "BA"
Sheridan Canyon
W
E
Terrigenous clastics
Deeper water lime-stones.
Bioherms, flanking beds, breccias
400 feet
1 mile
G
F
E
D
C B
A
Wolfcampian
Virgilian
G
F
E
C
B
A
Wolfcampian
Virgilian

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PLATE A

GEOLOGICAL MAP OF THE EAST SIDE OF THE ENGADIN VALLEY BETWEEN ST.MORITZ AND ZUOZ.

PLATE A

Sedimentary zones of Bernina nappe:

1: Piz Alv
2: Monte Garone – Valle del Monte
3: Gessi
4: Plaun da Staz
5: Piz Mezzaun
6: Bugliaun

A: Bernina nappe (crystalline core)

B: Stretta fold (crystalline of Bernina nappe)

C: Languard nappe

D: Seja slice (Languard nappe)

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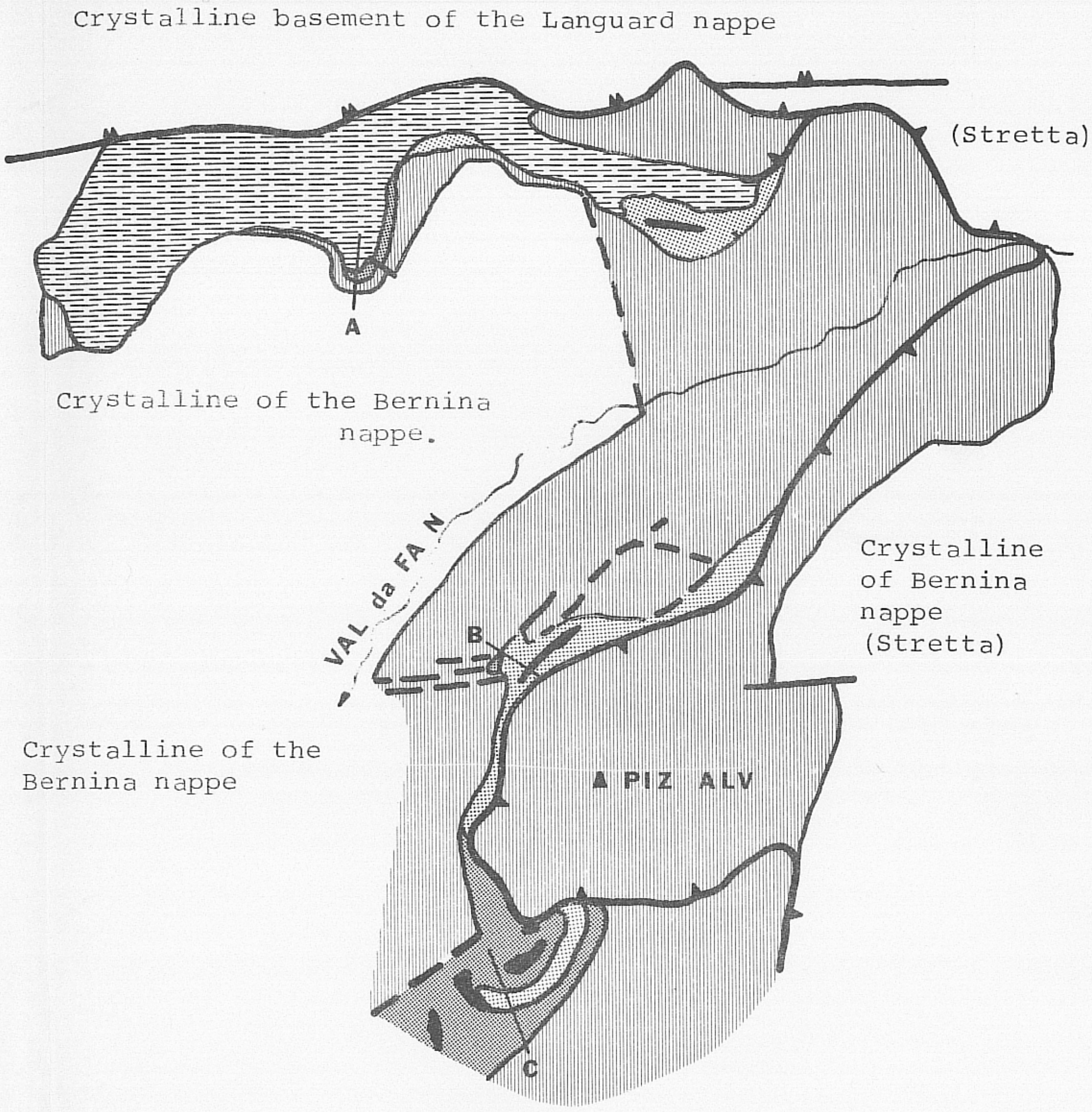
PLATE B

GEOLOGICAL MAP OF THE PIZ ALV AREA.



Scale: 1:25 000

- Mezzaun Formation
- Fain Formation
- Alv Formation
- Triassic formations
- Sedimentary blocks
- Alpine faults
- Alpine thrusts
- Jurassic faults
- Stratigraphic sections

A,B,C

Stratigraphic sections

PLATE B

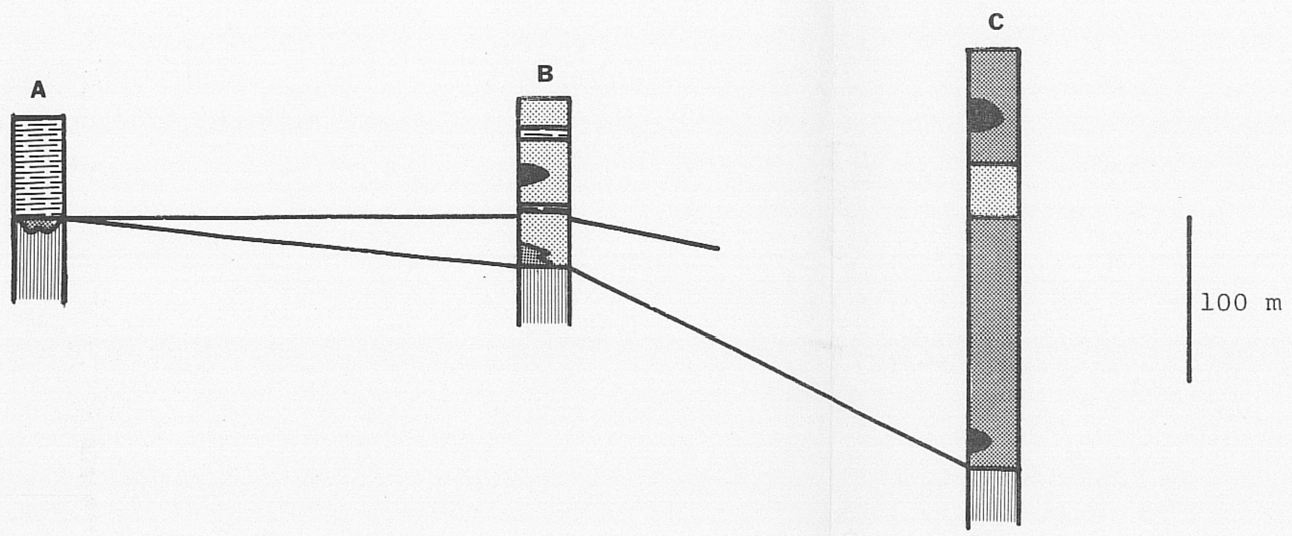


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PLATE C

GEOLOGICAL MAP OF THE PIZ MEZZAUN AREA.

Scale 1 : 25 000

Jurassic formations. Mezzaun Formation, except in the following cases:

Tectonic slice of Il Corn: Mezzaun Formation
Fain Formation
Alv Formation

Lower tectonic slice of Lavirum: Mezzaun Formation.
Alv Formation.

Tectonic slice of Medras: Radiolarian chert above
Mezzaun Formation.

Tectonic slice of Stevel: Mezzaun Formation
Alv Formation.

Tectonic slice of Chamues-ch: Mezzaun Formation
"Steinsberg Lias"

C Triassic formations.

B Crystalline of the Bernina nappe

A Crystalline of the Languard nappe

The sedimentary formations belong all to the Bernina nappe

1: Tectonic slice of the Il Corn
2: Upper tectonic slice of Lavirum
3: Lower tectonic slice of Lavirum
4: Upper tectonic slice of the Piz Mezzaun
5: Tectonic slice of Medras
6: Tectonic slice of Stevel
7: Lower tectonic slice of the Piz Mezzaun
8: Tectonic slice of Chamues-ch

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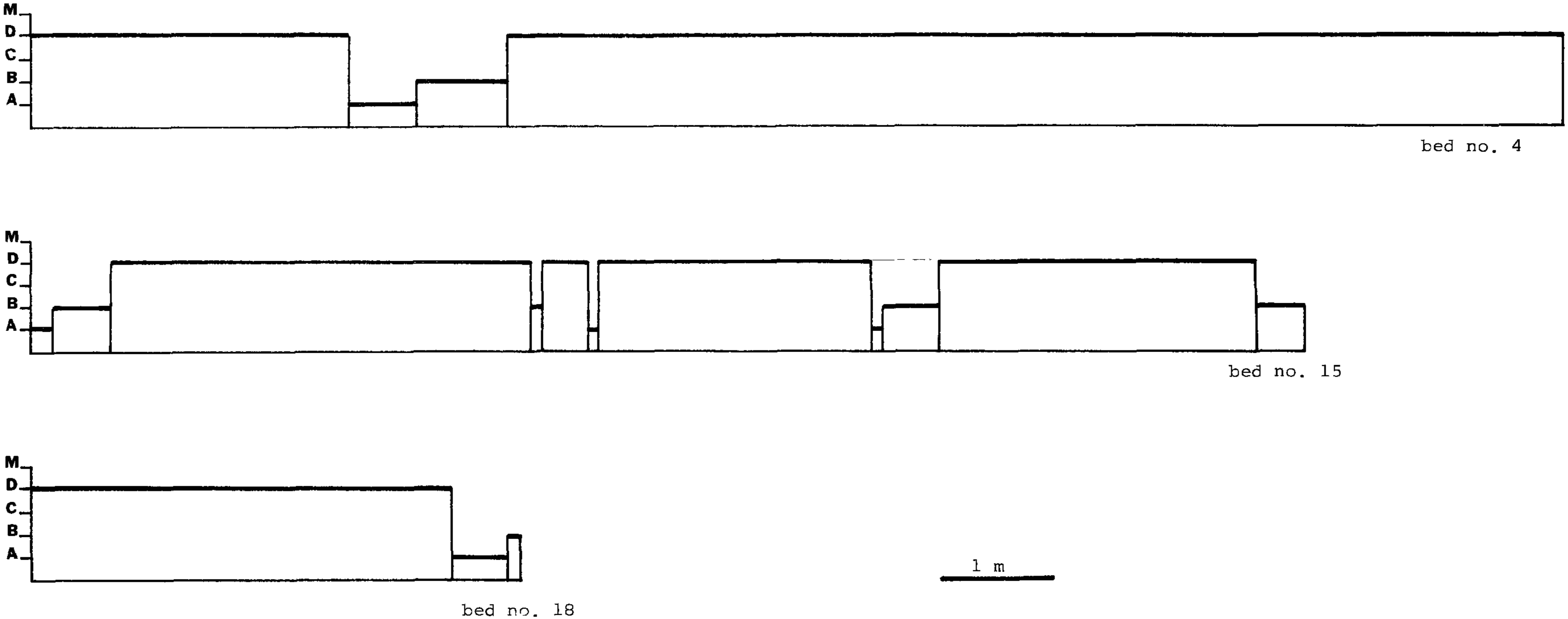
PLATE D - 1

MEASURED SECTION OF THE MEZZAUN FORMATION (TECTONIC SLICE OF STEVEL)

PLATE D

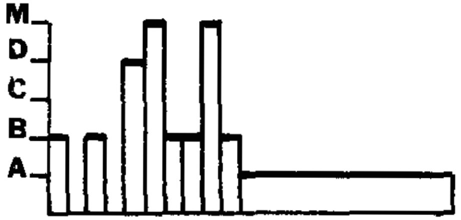


MEASURED SECTION OF THE MEZZAUN FORMATION (TECTONIC SLICE OF MEDRAS)

PLATE D

Unit 4

bed 205



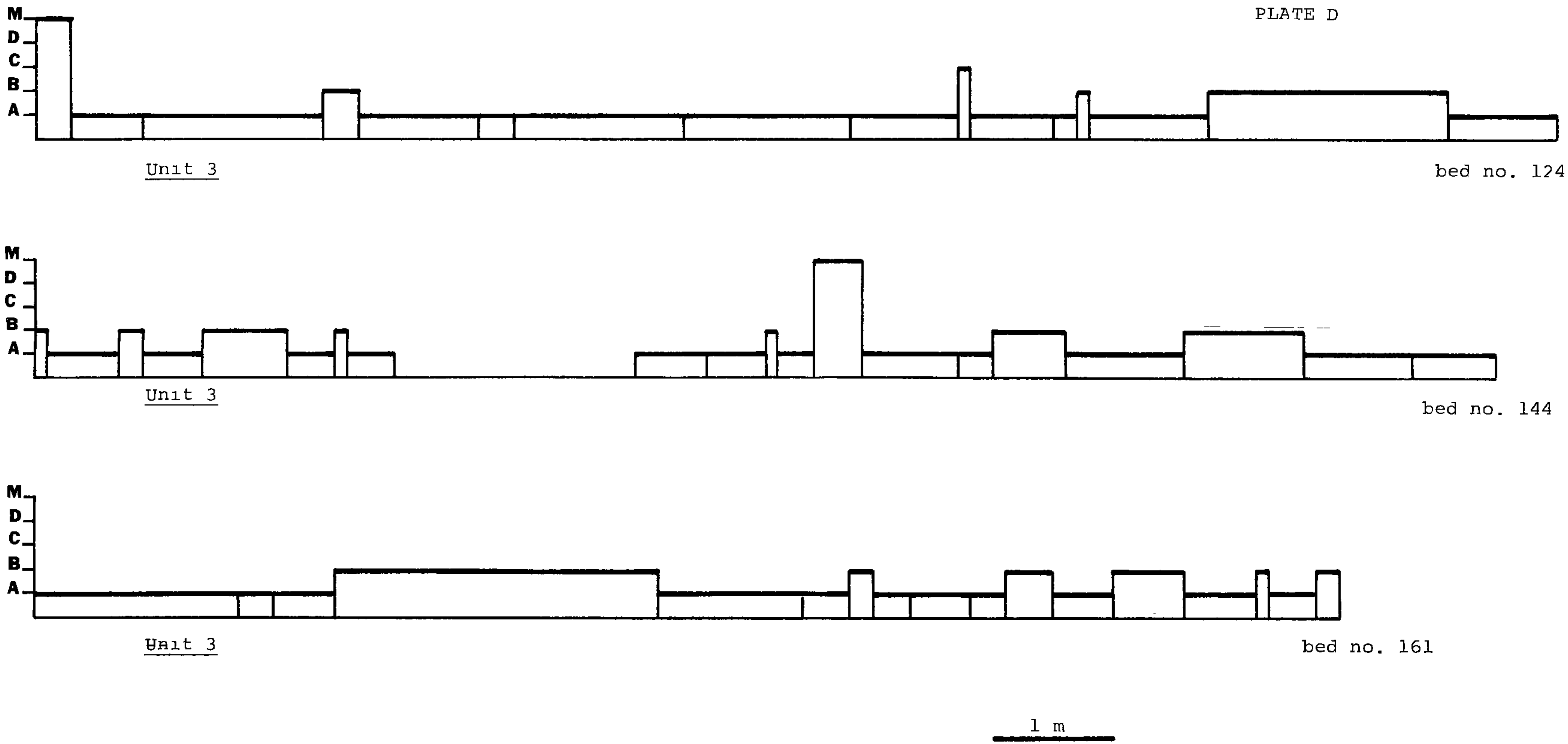
Unit 4

bed no. 214

1 m

MEASURED SECTION OF THE MEZZAUN FORMATION (TECTONIC SLICE OF MEDRAS)

PLATE D



MEASURED SECTION OF THE MEZZAUN FORMATION (TECTONIC SLICE OF MEDRAS)

PLA D

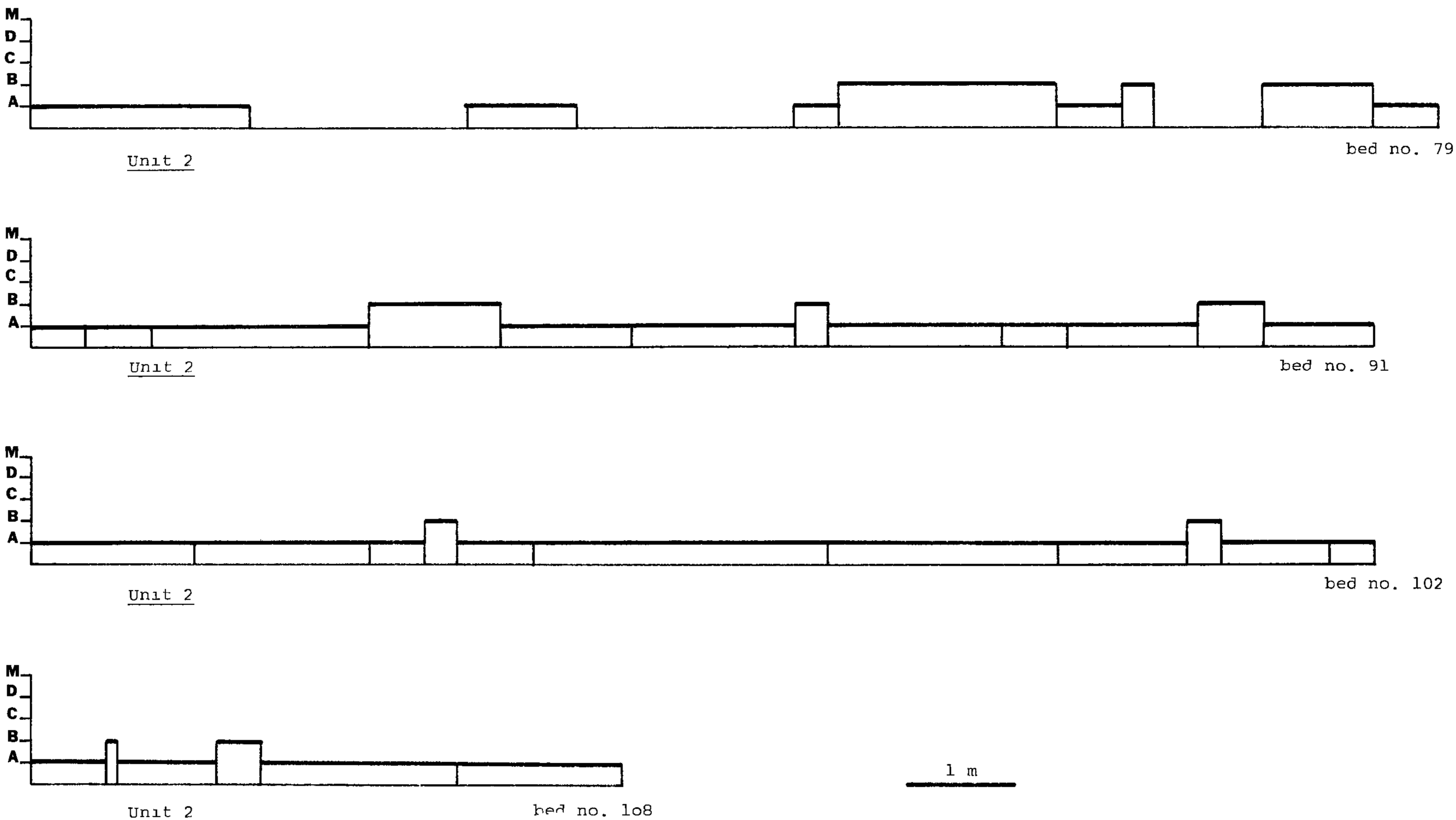


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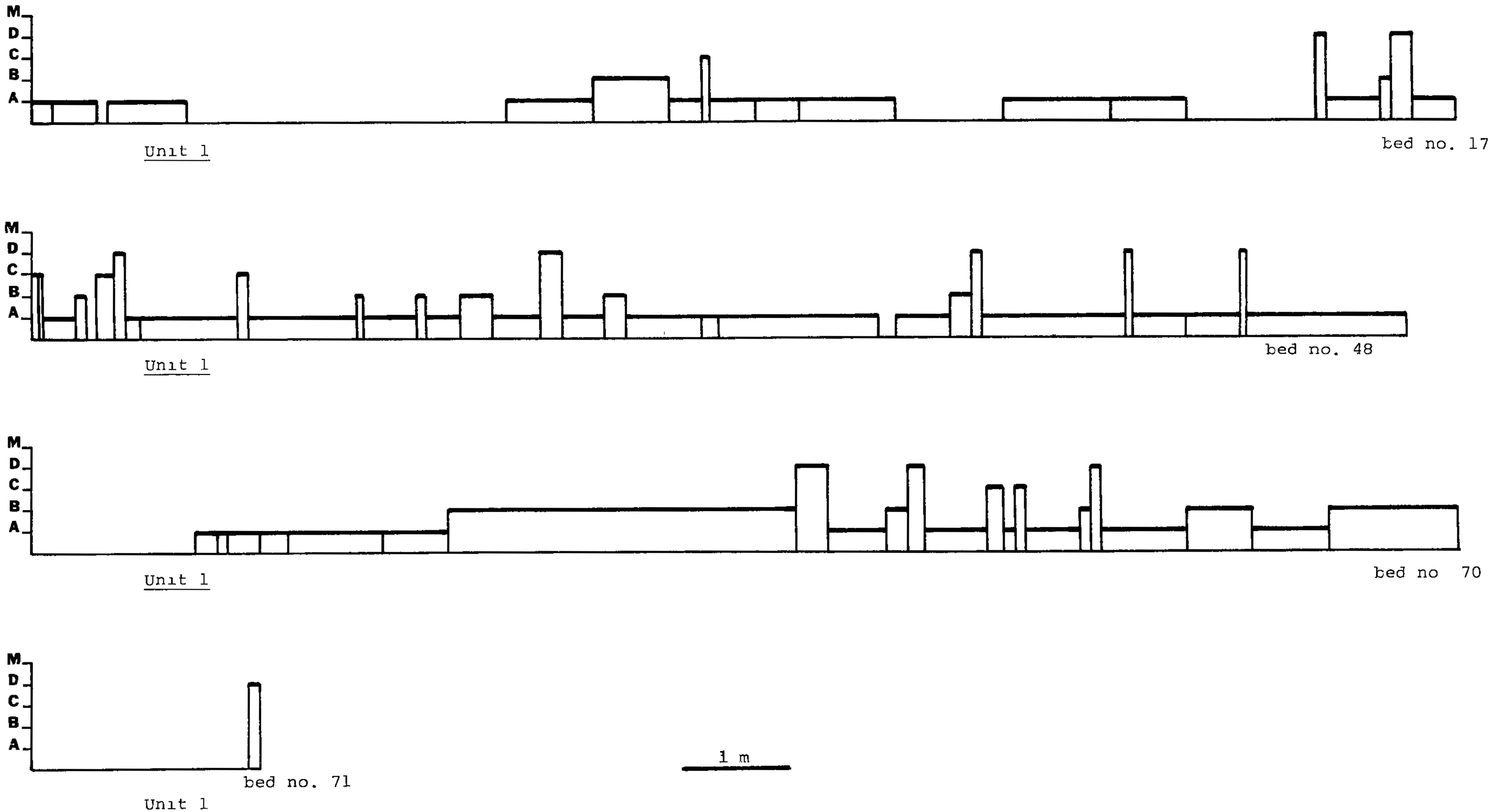
PLATE D - 5

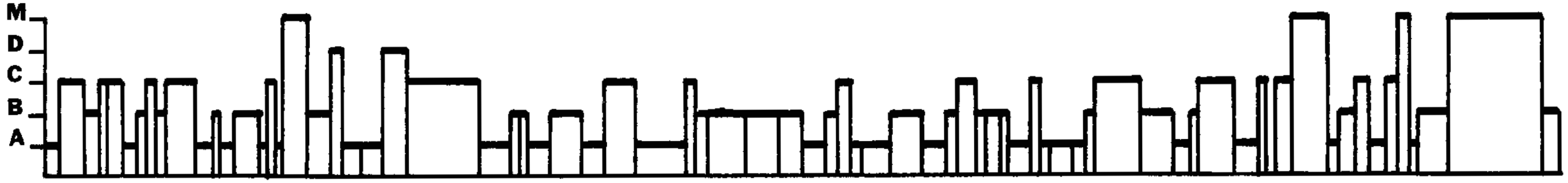
PLATE D

MEASURED SECTION OF THE MEZZAUN FORMATION (TECTONIC SLICE OF MEDRAS)



Measured section of the Mezzaun formation (Lower tectonic slice of Piz Mezzaun)

PLATE D

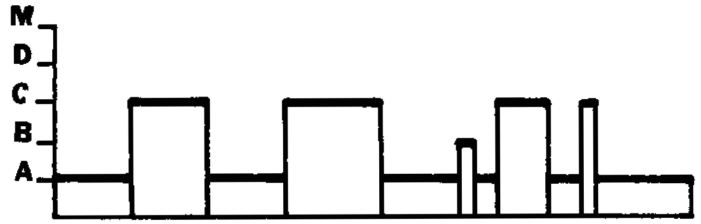


bed no. 73

Unit 1

bed no. 49

Unit 2



1 m

bed no. 60

Unit 2