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(with 50 figures, 4 tables, 10 plates and 1 geological map)

PADOVA
SOCIETÀ COOPERATIVA TIPOGRAFICA
1991

MEMORIE DI SCIENZE GEOLOGICHE
già *Memorie degli Istituti di Geologia e Mineralogia dell'Università di Padova*

Vol. XLIII, pagg. 17-123, 50 figg., 4 tabb., 10 tavv. e 1 carta geologica
Padova, maggio 1991

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EARLY MESOZOIC EXTENSION AND ALPINE SHORTENING IN THE WESTERN SOUTHERN ALPS: THE GEOLOGY OF THE AREA BETWEEN LUGANO AND MENAGGIO (LOMBARDY, NORTHERN ITALY)

CONTENTS

Key words: Southern Alps, extensional tectonics, Alpine tectonics, subsidence, fault rocks.

ABSTRACT

The Southern Alps represent a segment of the Mesozoic South-Alpine passive continental margin which was involved in Alpine shortening in Tertiary times. During extension swells with reduced sedimentation and strongly subsiding basins developed. In this work the large-scale tectonic and sedimentation pattern of two of the most important extensional structures, the Lugano swell and the M. Generoso basin, are studied. The regional results are then integrated with data obtained from the literature in order to discuss the extension of the South-Alpine margin at a lithospheric scale.

The crystalline basement of the studied area consists of paragneisses and subordinate orthogneisses and amphibolites of Variscan age. The Upper Permian to Carnian sediments were deposited in continental to shallow marine environments and show no abrupt thickness or facies variations. Beginning with the Norian, normal faulting caused dramatic differences in sediment thickness and, sometimes, facies. It is in Norian time that the Lugano swell and the M. Generoso sedimentary basin were individuated. The two domains were separated by the E-dipping Lugano normal fault. Normal faulting was going on also inside the basin itself. The Norian extensional pattern goes on until the Middle Rhaetian when most of the faults inside the basin were disactivated. Strong subsidence of the M. Generoso basin continued up to the Toarcian along the Lugano normal fault. In the Toarcian normal faulting comes to an end.

Tertiary Alpine tectonics caused the imbrication and the steepening of the northern part of the basin. Also pre-Alpine structures like the northern segment of the Lugano normal fault were steepened. The Lugano normal fault, after erosion, is now exposed at the surface and coincides with the E-W striking M. Grona line of previous authors. The central and western segments of the M. Grona line represent the contact between basement in the north and sediments in the south. Bedding in sediments close to the line is gener-

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The Swiss National Fond (grant 4.903-0.85.20) is gratefully acknowledged for financial support during the field work. The University of Zürich, the Rudolf STAUB Fonds and the MEYER-KEYSER Legat have financed the publication of this Thesis.

ally oblique with respect to the line; the sediments become younger from E to W. The eastern part of the M. Grona line is intracrystalline. Along the M. Grona line fault rocks are exposed which formed during the Mesozoic normal faulting. Only brittle deformation is found south of the line (hangingwall of the Lugano normal fault). North of the line the cataclastic rocks found in the west are progressively substituted by lower greenschist mylonites towards the east. The mylonites are overprinted by cataclastic deformation moving from N to S towards the M. Grona line.

On the base of sediment thickness and of the analysis of the fault rocks, the geometry of the Lugano normal fault before Alpine steepening can be reconstructed. The fault was steep (50°-70°) from the surface to a depth of ca. 5-7 km. It attained then gentler dips of 10°-20° till a paleo-depth of ca. 15 km. Deformation is always limited to a well definable zone.

A similar geometry is shown by the Lago Maggiore fault which controlled the evolution of the M. Nudo basin, west of Lugano. A stretching factor of 1.5 can be reconstructed along a W-E striking profile from the Canavese to the Albenza Plateau. The analysis of subsidence as derived from sedimentary record, suggests that the tectonically-derived crustal stretching factor applies also to the underlying lithospheric mantle.

RIASSUNTO

Le Alpi Meridionali rappresentano un segmento del margine continentale passivo della placca (o promontorio) adriatica (apula) successivamente coinvolto nel raccorciamento alpino. Durante la distensione si formarono bacini subsidenti e zone caratterizzate da sedimentazione ridotta.

In questo lavoro, basato sul rilevamento di terreno e sullo studio sedimentologico-strutturale della porzione settentrionale del bacino del Monte Generoso e della soglia di Lugano (Alpi Meridionali centro-occidentali) vengono investigati i problemi legati alla distensione mesozoica e alla sua influenza sulla sedimentazione. I risultati dello studio vengono poi integrati con dati dalla letteratura per discutere il problema della distensione litosferica nel Sudalpino.

Il basamento cristallino della regione studiata è costituito da paragneiss e, subordinatamente, da ortogneiss in facies anfibolitica. La principale fase di metamorfismo è ercinica. I sedimenti dal Permiano superiore fino al Carnico sono stati depositati in ambiente da continentale a marino e non mostrano variazioni rilevanti di facies e/o spessore tra la zona del Luganese e quella ad est della linea di Lugano. È invece a partire dal Norico che l'attivazione di importanti faglie normali determina drammatiche variazioni di spessore e, talvolta, anche di facies. È nel Norico che cominciano ad individuarsi i due domini che caratterizzano la regione fino al Lias: la soglia di Lugano e il bacino fortemente subsidente del Monte Generoso. All'interno di quest'ultimo si impostarono altre faglie normali. Il *pattern* distensivo norico continuò pressocché immutato fino al Retico medio quando gran parte delle faglie normali attive all'interno del bacino del Generoso si estinsero. La subsidenza del bacino continuò però ad essere forte lungo la faglia di Lugano che si identifica quindi come una *master fault* crostale immergente verso est. Le faglie normali e quindi la distensione terminano nel Toarciano.

La tettonica alpina terziaria ha causato, oltre alla formazione di due scaglie nord-vergenti, anche la subverticalizzazione dei sedimenti e della parte settentrionale della faglia di Lugano che affiora adesso alla superficie e corrisponde alla linea del Monte Grona con andamento E-W. La parte centro-occidentale della linea rappresenta il contatto tra basamento a nord e sedimenti a sud; nella sua parte orientale la linea diventa intracristallina.

Lungo la faglia sono esposte rocce formatesi durante i movimenti lungo la faglia (*fault rocks*). A sud della linea (*hangingwall* della faglia di Lugano), la deformazione è solo fragile. A nord della linea (*footwall*) è visibile una varietà di *fault rocks* disposte sistematicamente secondo due *trends*: uno W-E ed uno N-S. Nella parte occidentale della faglia si hanno solo brecce cataclastiche. A partire dalla Val Sanagra invece compaiono miloniti in facies di scisti verdi che diventano meglio sviluppate e più potenti verso est. Lungo un profilo N-S le miloniti vengono interessate progressivamente da deformazione a temperature decrescenti con formazione di ultramiloniti e poi brecce cataclastiche.

Sulla base delle microstrutture delle *fault rocks* e con il controllo dei sedimenti è possibile ricostruire la geometria profonda della faglia di Lugano prima del raccorciamento alpino. La faglia di Lugano era ripida (50°-70°) dalla superficie fino a 5-7 km di profondità; a profondità maggiori, fino a circa 15 km, l'inclinazione era di 10°-20°. Fino a questa profondità la faglia rimane comunque ben individuata. Una geometria simile è mostrata dalla faglia del Lago Maggiore che ha controllato l'apertura del bacino del Monte Nudo. Costruendo un profilo dal Canavese al margine est del bacino del Monte Generoso si può ricostruire uno *stretching factor* per la crosta del bacino lombardo occidentale pari a 1.5. L'analisi della subsidenza dedotta dai sedimenti permette di avanzare l'ipotesi che la distensione osservata nella crosta intermedia e superiore sia compensata a livelli litosferici più profondi sulla stessa verticale: il mantello litosferico sotto il bacino lombardo si sarebbe quindi esteso con una geometria generale di taglio puro.

ZUSAMMENFASSUNG

Die Südalpen stellen ein in die alpine Verkürzung miteinbezogenes Fragment des mesozoischen passiven Kontinentalrandes der Apulischen (Adriatischen) Platte dar. Während der mesozoischen Streckung entstanden subsidierende Becken und durch geringere Sedimentation charakterisierte Hochzonen. In dieser strukturgeologischen und sedimentologischen Studie werden die Lugano-Schwelle und das M. Generoso-

Becken und ihre Bedeutung im Rahmen der mesozoichen Dehnung untersucht. Auf Grund der Resultate dieser Studie und der Daten aus der Literatur wird das Problem der lithospherischen Dehnung im Südalpin diskutiert.

Das kristalline Grundgebirge des untersuchten Gebietes besteht aus Paragneissen und untergeordneten Orthogneissen und Amphiboliten herzynischen Alter. Die permischen bis karnischen Sedimente wurden in kontinentalen und marinen Bereichen abgelagert und zeigen keine abrupten Fazies- und Mächtigkeitsänderungen, die man auf aktive Brüche zurückführen kann. Im Nor verursachte die Aktivierung von normalen Brüchen (Abschiebungen) auffallende Mächtigkeits- und Fazies-Unterschiede zwischen den auf der Hochzone und den im Becken abgelagerten Abfolgen. Im Nor entwickelten sich die zwei Hauptdomänen der Region: die Lugano—Schwelle und das M. Generoso—Becken. Abschiebungen wurden auch innerhalb des Beckens entwickelt. Das norische Dehnungsmuster bestand praktisch unverändert bis zum mittleren Rhät, während dem meisten der sich im Becken befindenden Brüche ausklingen. Die Subsidenz des M. Generoso—Beckens dauerte trotzdem bis in den späteren Lias fort und wurde von dem nach Osten abtauchenden Lugano-Bruch kontrolliert.

Die tertiäre alpine Tektonik verursachte eine nordvergente Verschuppung und die Steilstellung der Sedimente und der prä-alpinen Strukturen besonders des nördlichen Segmentes des Lugano-Bruches. Dieser ist jetzt an der Erdoberfläche aufgeschlossen und koinzidiert mit der W-E verlaufenden M. Grona-Linie früherer Autoren. Die westlichen und zentralen Teile der M. Grona-Linie trennen die metamorphen Gesteine im Norden von den Sedimenten im Süden. Die Sedimente sind diskordant in Bezug auf die M. Grona-Linie und werden jünger von Osten nach Westen. In ihrem östlichen Teil ist die M. Grona-Linie intrakristallin.

Dem Bruch entlang sind Gesteine aufgeschlossen, die in Zusammenhang mit den mesozoischen tektonischen Bewegungen entstanden sind. Die Verformung südlich der Linie (hangingwall des Lugano-Bruches) ist nur spröd. Nördlich der Linie (footwall) zeigen die fault rocks zwei Trends: von W nach E werden kataklastisch deformierte Gesteine durch schwach grünschieferfazielle Mylonite ersetzt; die Mylonite werden von N nachs S spröd überprägt.

Auf Grund der Sedimentmächtigkeiten und der Analyse der fault rocks kann die Geometrie des mesozoischen Lugano-Bruches vor der alpinen Steilstellung rekonstruiert werden. Von der Oberfläche bis zu einer Tiefe von 5-7 km war der Lugano-Bruch steil (50° - 70°). Er wurde dann flacher und hatte eine Neigung von 10°-20° bis eine Tiefe von ca. 15 km. Bis in diese Tiefe ist die Verformung auf eine gut definierbare Zone beschränkt.

Eine ähnliche Geometrie zeigt der Lago Maggiore-Bruch, der das westlich von Lugano liegende M. Nudo-Becken begrenzt. Entlang einem E-W streichenden Profil vom Canavese bis zum Albenza Plateau kann ein Dehnungsfaktor von 1.5 rekonstruiert werden. Die Analyse der von den Sedimenten abgeleiteten Subsidenz läßt vermuten, daß der tektonisch rekonstruierte Dehnungsfaktor auch für die ganze Lithosphäre gilt.

1 INTRODUCTION

1.1 Objectives, general background and the model

The main topic of this thesis is the kinematics of extension of a fossil continental margin as exemplified by the Lombardian zone of the Southern Alps.

It is by now widely accepted that substantial extension leading to the formation of sedimentary basins takes place during rifting between two lithospheric plates preceding the opening of an intervening ocean (Sleep, 1971). Although the geodynamic context is generally accepted, it is still poorly known how the extension actually occurs: mode and geometry of extension in the different crustal levels and in the lithosphere are presently highly debated. This is above all due to the fact that most passive continental margins are presently either not directly accessible to observation or, when involved in subsequent compression, highly disrupted. In the latter case, most of the extension-related normal faults are reactivated as thrusts and horizontal as well as vertical relationships are difficult to reconstruct.

Extensional structures are, however, generally quite well preserved in the Southern Alps; these represent one of the best preserved segments of fossil passive continental margins exposed at the earth surface. Stratigraphers and sedimentologists have long been studying the sedimentary successions of the Southern Alps and since at least 20 years the importance of Early and Middle Mesozoic extension in controlling sedimentation patterns is generally accepted (Bernoulli, 1964; Bosellini, 1973; Gaetani, 1975; Winterer and Bosellini, 1981).

Tectonicians and structural geologists in contrast, have devoted surprisingly little attention to the study of the structures associated with extension, rather concentrating on the Alpine features.

This study is mainly an attempt to look in a mountain chain for extension-related structures and to use them to demonstrate how extension occurred at different structural levels.

The Southern Alps, stretching from Yugoslavia to the western Alps south of the Periadriatic Line (*sensu* Schmid *et al.*, 1989) are a segment of the Adriatic passive continental margin which un-

derwent extension in the Early and Middle Mesozoic due to the moving apart of Adria from Europe preceding the opening of the Jurassic Tethys ocean (BERNOULLI and LEMOINE, 1980). During extension the margin was cut by generally, but not always, N-S striking normal faults separating basins and highs. As a result, paleogeographic domains were identified which were typically elongated in a N-S direction. From W to E the major domains are: the Canavese zone, the Gozzano swell, the Lombardian basin (comprising Monte Nudo basin, Arbostora swell, Monte Generoso basin, Albenza zone, Sebino trough, Botticino high), Trento plateau, Belluno trough and Friuli platform (e.g. AUBOUIN, 1963; GAETANI, 1975; WINTERER and BOSELLINI, 1981) (Fig. 1) (1).

Alpine orogeny in Late Cretaceous (?) to Tertiary times caused mainly N-S directed shortening in the western Southern Alps, i.e. perpendicular to the Early Mesozoic extension (LAUBSCHER, 1985; DOGLIONI and BOSELLINI, 1987). As a consequence, the major N-S striking normal faults were only partially, if at all, reactivated as strike-slip faults, so that E-W relationships between the different features are substantially preserved. Also due to Alpine shortening, formations exposed at the surface become younger from the axial part of the chain in the north to the foreland in the south (Po Plain). This offers a very good opportunity to study pre-Alpine structures at different crustal levels.

Particularly favorable in this respect is the Lago di Como-Lago di Lugano region (Fig. 1) which is the object of this study. The area lies north of some of the sedimentologically and stratigraphically best known extensional structures of the Southern Alps: the *Monte Generoso basin* and the adjacent *Lugano swell*; the two structures are separated by the *Lugano line*. The dramatic differences between the successions east and west of Lugano line have long been recognized. After many decades of debates, the Lugano line is now generally accepted to be the present day expression of a Mesozoic E-dipping normal fault (the

(1) The paleogeographic features are generally described on E-W trending profiles because the Mesozoic sediments are presently found along a relatively narrow E-W oriented stripe. As a consequence, paleogeographic control is good in the E-W dimension but limited in the N-S one. However, variations in the N-S direction often controlled by inherited ENE-WSW trending lineaments were important and should not be neglected (KÄLIN and TRÜMPY, 1977; CASTELLARIN, 1982).

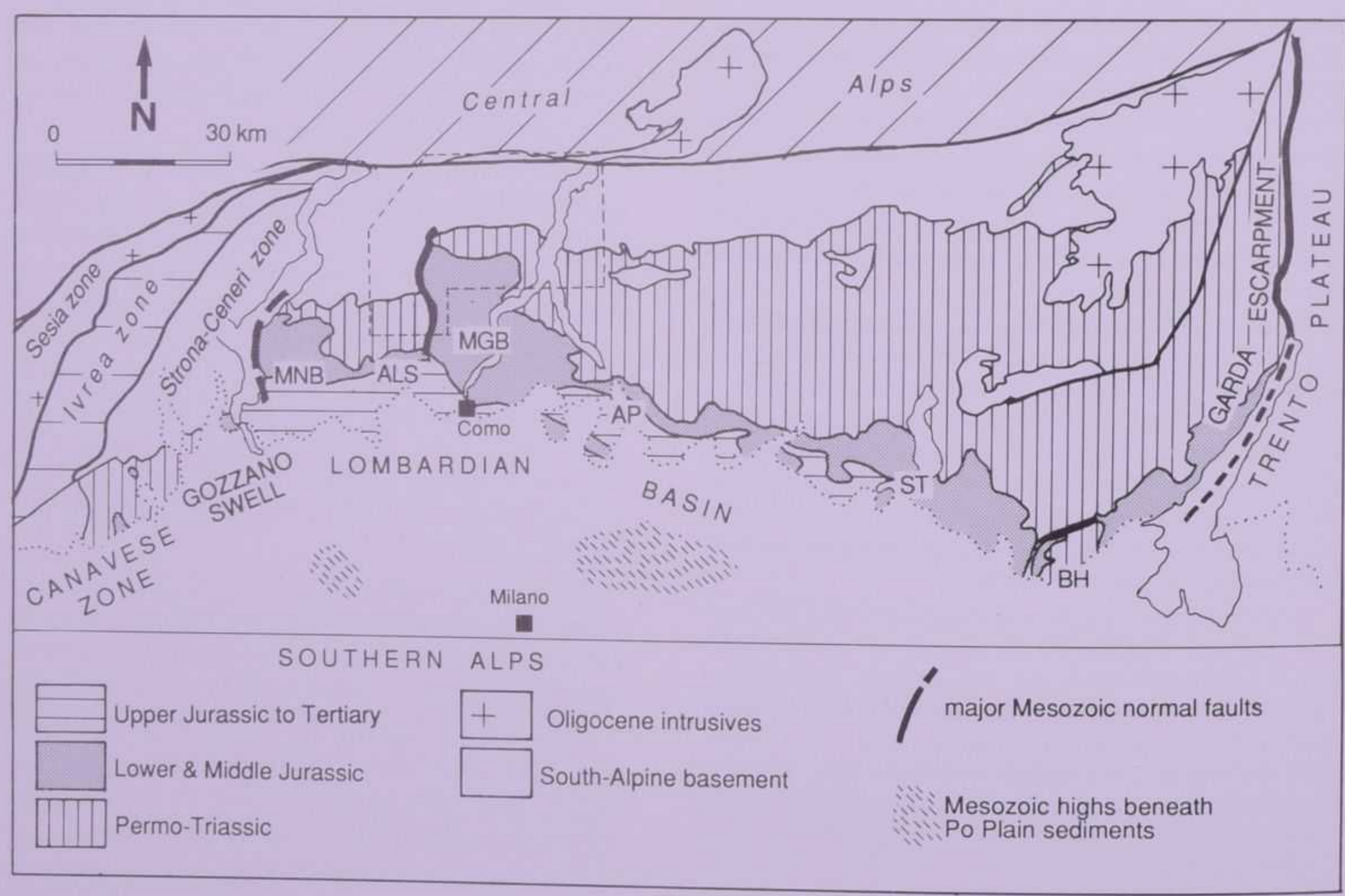

FIG. 1 - General map of the western Southern Alps. Dashed line outlines the study area (Fig. 2). GS = Gozzano swell; MNB Monte Nudo basin; ALS = Arbostora-Lugano swell; MGB = Monte Generoso basin; AP = Albenza plateau; ST = Sebino trough; BH = Botticino high.

Lugano normal fault) separating the Lugano swell in the west from the Generoso basin in the east (WIEDENMAYER, 1963; BERNOULLI, 1964).

BISTRAM (1903), who produced a beautiful map of the Lake Como-Lake Lugano area, first interpreted the Lugano line as a Liassic normal fault. His interpretation was however ignored and a number of geologists tried to interpret the Lugano line as a Late Cretaceous normal fault (FRAUENFELDER, 1916) or even as an Alpine thrust (DOEGLAS, 1930; STAUB, 1949; BURFORD, 1951 among others). The interpretation of the Lugano line as an Alpine reactivated Liassic normal fault was proposed again by VONDERSCHMITT (1940) and generally accepted after the two contributions by WIEDENMAYER (1963) and BERNOULLI (1964).

During Alpine shortening sediments, basement and pre-Alpine structures of the northern parts of the Generoso basin and of the Lugano swell have been brought into a steep, S-dipping position. Over a few km the complete sequence from the metamorphic rocks beneath the basin to the Liassic syn-rift sediments is observed (Fig. 2). Because of the steep position of bedding, the present day map-view is actually a relatively undisturbed paleo-depth section of the Monte Generoso basin (Fig. 3). This idea represents a kind of red thread throughout my study.

The present day setting allows a study not only of the syn- and pre-rift successions and of the basement but also of the structures which controlled at depth the evolution of the basin. Particularly significant is that also the Mesozoic Lugano normal fault controlling the development of the Generoso basin has been passively folded and is presently exposed at the surface (BALLY *et al.*, 1981). I identify the steepened part of the Lu-

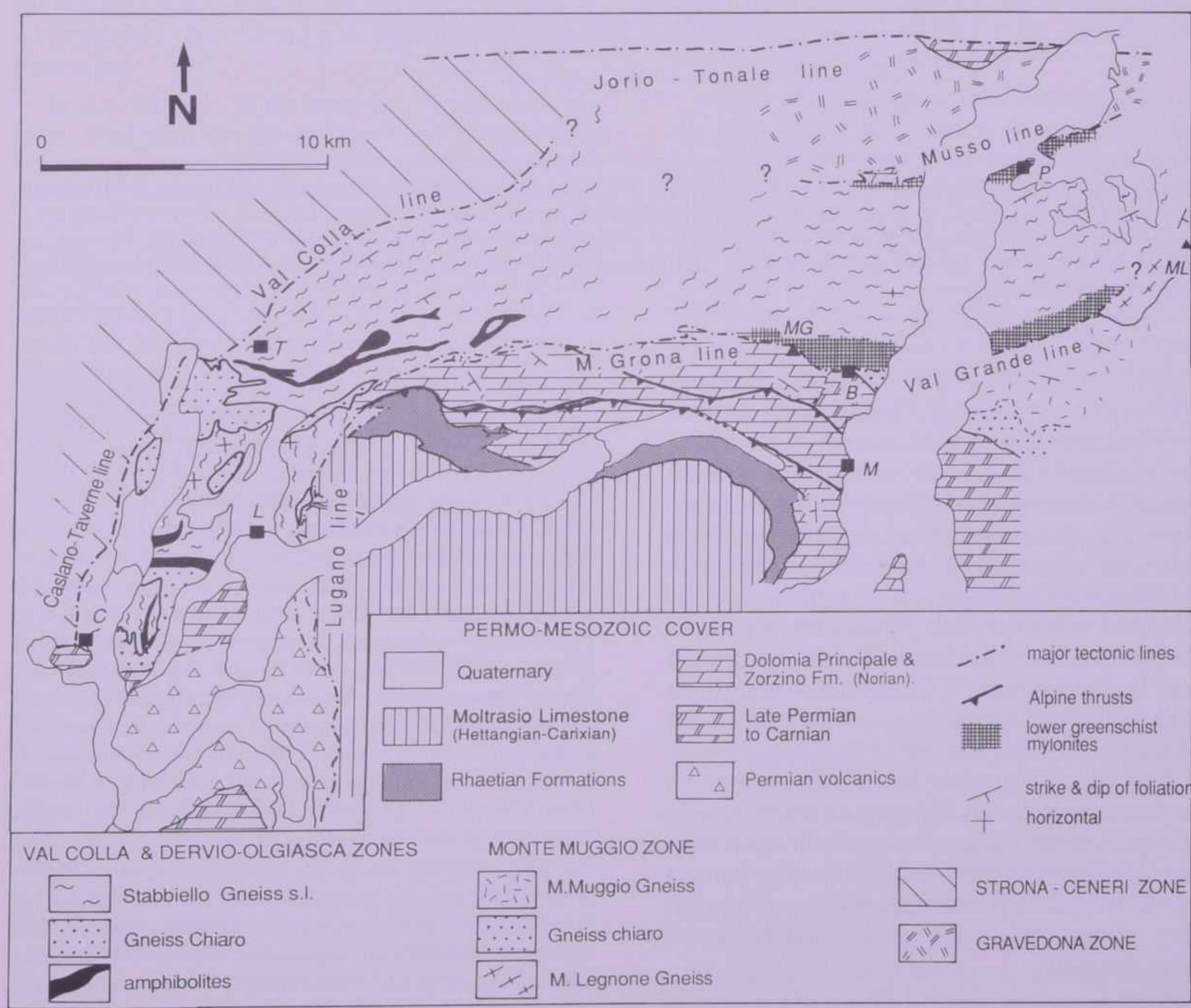

FIG. 2 - Geologic map of the study area and adjacent regions. After own data and maps by REINHARD (1964); BERNOULLI (1964); EL TAHLAWI (1965); FUMASOLI (1974). C = Caslano; L = Lugano; T = Tesserete; MG = Monte Grona; B = Breglia; M = Menaggio; P = Piona; ML = Monte Legnone.

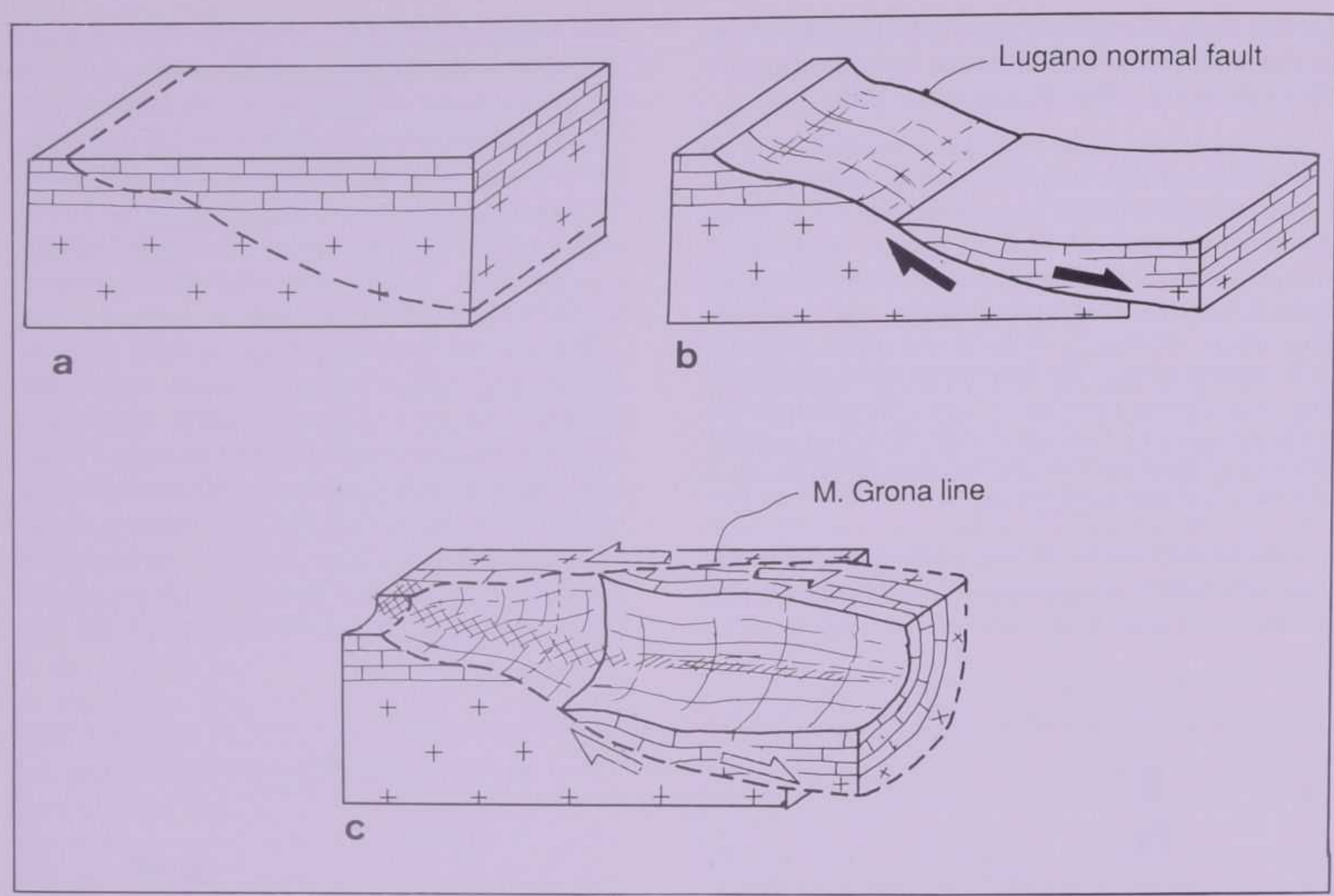

FIG. 3 - Cartoon illustrating the model proposed in this study. a) situation before the onset of rifting; b) situation during rifting: development of the basin is mainly controlled by the Lugano normal fault; c) Alpine shortening brings the northern part of the Generoso basin and of its substratum into a steep position. Mesozoic movement along the normal fault appears as sinistral after Alpine steepening. Syn-rift sediments and internal thrusting have been removed for clarity.

gano fault with the *Monte Grona line* (LEHNER, 1952), a tectonic feature which till now has not received convincing explanations. The Monte Grona line (Fig. 2) represents the sediment/basement tectonic contact from the Tesserete region in the west to Breglia in the east. The line is generally oblique with respect to sedimentary bedding. Across the line a stratigraphic cut-off is observed with the whole Triassic missing in the W and the lower to middle Triassic missing in the E (Breglia region). Further to the east the Monte Grona line becomes intracrystalline. All these features find a comprehensive explanation in the proposed model (Fig. 3). The kinematic and metamorphic characteristica of the fault rocks along the Monte Grona line perfectly agree with the hypothesis of the Monte Grona line being a northern and originally deep segment of the Lugano normal fault passively steepened during Alpine shortening.

The first geologic maps which have been made in the Lake Lugano-Lake Como area (NEGRI *et al.*, 1869; TARAMELLI, 1890 and 1903; SCHMIDT and STEINMANN, 1890) all show the middle Triassic dolomites ("Muschelkalk" of the Swiss authors and "Dolomia inferiore" of Taramelli) lying stratigraphically on the basement east of Lugano. Only at the beginning of the century it was recognized (REPOSSI, 1902; BISTRAM, 1903) that most of the dolomites mapped as middle Triassic were instead of Norian age: a relevant stratigraphic omission was therefore present along important segments of the contact. The anomalous nature of the contact between basement and sediments W of the Monte Grona area was progressively accepted. Both REPOSSI (1902) and VON BISTRAM (1903) interpreted the Monte Grona line as a fault lowering the sediments of the southern block with respect to the metamorphic rocks of the northern one. The fault would continue E of Breglia along the basement/Verrucano and Verrucano/S.Salvatore Dolomite contacts both of which were interpreted as tectonic. It is however not clear how these authors could kinematically explain the observed stratigraphic cut-off with a normal fault which was supposed to be parallel to bedding. SEITZ (1917) on the base of the stratigraphic nature of the basement/Verrucano and Verrucano/S.Salvatore Dolomite contacts demostrated by TORNQUIST (1902) during the construction of the new road along Lake Como rejected the hypothesis by BISTRAM and REPOSSI. He, however, accepted the tectonic nature of the Monte Grona

line east of Monte Grona; how he could kinematically link the two domains remains unknown. Probably because of these kinematic problems, the tectonic interpretation of the line was abandoned.

Around the half of the century, MAGNANI (1944) and VENZO and MAGLIA (1947) rejected the tectonic interpretation of the Monte Grona line. In order to explain the incomplete sedimentary succession, the authors postulated a morphologic basement high onto which the Carnian sea transgressed. LEHNER (1952) on the base of breccias found along the sediment/basement contact (which he first named Monte Grona line), breccias which he interpreted as of sedimentary origin, and on the base of the thin tectonized outcrops of Verrucano and Raibl Beds found along the Monte Grona line accepted the idea of a morphological high persisting throughout the Early and Middle Trias. LEHNER at the same time accepts the possibility of a tectonic reactivation of the Monte Grona line in relationship with movements along the Jorio-Tonale line. In his sketch-map he recognized for the first time an important E-W trending fault W of Acquaseria and implicitly interpreted it as the E-ward prolongation of the Monte Grona line.

In the '60 and '70 no important contributions appear concerning the Monte Grona line: authors working in adjacent areas (*e.g.* BERNOULLI, 1964; GNACCOLINI, 1965a) generally adopted the "stratigraphic" interpretation.

The "stratigraphic interpretation" has recently gained new popularity in a somewhat modified and extended form (GIANOTTI, 1984; GAETANI *et al.*, 1986; GIANOTTI and TANNOIA, 1988). According to these authors the basement N of the Monte Grona line acted as a morphological high of regional importance (reaching the Musso area) from the late Permian to the Early Carnian when the Raibl Beds and then the Dolomia Principale were deposited directly on the basement. Their interpretation is based on the same "evidence" proposed by LEHNER. In addition they describe an horizon of breccias which they interpret as "transgressive horizon" (Sasso di Cusino) and mention Carnian palynomorphs from the fault zone (Val Sanagra). Both LEHNER's breccias and the "transgressive horizon" turn out to be, at a closer study, cataclastic rocks so that they cannot prove the sedimentary nature of the Monte Grona line. The mentioned middle Triassic ages are also of little use since they come from horizons of tectonic breccias containing clasts of different ages. The stratigraphic hypothesis has therefore no compelling evidence.

BALLY *et al.* (1981) discussing listric faults in general, briefly speculated that the Monte Grona line could represent a northern and deeper segment of the Lugano normal fault. They thus anticipated the interpretation presented in this study.

LAUBSCHER (1985) was the first author to newly stress the tectonic relevance of the Monte Grona line which he interprets as a back-thrust of Alpine age. His interpretation however does not take into account the pre-Alpine events and fails therefore to explain important features such as the obliquity between Monte Grona line and bedding.

1.2 METHODS AND ORGANIZATION OF THE STUDY

A detailed mapping has been carried out in the region lying between Lugano, Menaggio and N of the Val Menaggio (Fig. 1). Basement lithologies and evolution are described in chapter 2.2. Since the basement except for its southern and northern boundaries does not play a central role in my work, this chapter is mainly a review of the few data available from the literature integrated with personal observations.

Stratigraphy and sedimentology of the sedimentary units are described in chapter 2.3. The emphasis is laid on the sediments as tracers of the Late Permian to Middle Mesozoic paleotectonic evolution of the area, rather than on sedimentology. For this purpose I will often compare the sedimentary successions in the area of study with those of adjacent areas.

Alpine structures have been reconstructed on the base of mapping and macroscopic analysis (chapter 3). This allows for the reconstruction of Alpine deformation in the Lago di Lugano-Lago di Como region. This is useful because geometry of Alpine shortening W of the Grigna group (LAUBSCHER, 1985) is poorly understood. The knowledge of Alpine structures is also necessary for the further development of this study in order to obtain a picture of the pre-Alpine situation.

Once the effects of Alpine shortening are filtered out, the features related to Mesozoic extension can be observed. Extensional features have been studied in the sedimentary cover (chapter 4.2) and in the basement (chapter 4.3). A number of normal faults which strongly influenced sedimentation patterns has been recognized in the sedimentary units. This allows for a control on the superficial response to the deep seated extension. The fault rocks outcropping along the Monte Grona line and its eastern prolongation, the Val Grande line, have been studied in detail in order to test the hypothesis that the Monte Grona line is a partially reactivated northern segment of the Lugano normal fault and to constrain kinematic and metamorphic conditions of deformation. Microscopic analysis was necessary for this purpose.

On the base of these data and restoring the sediments with associated Mesozoic extensional structures to their pre-Alpine position, the structure of the Monte Generoso basin could be reconstructed at a crustal scale (chapter 4.4). The re-

sults of the study in the Monte Generoso basin have been then integrated with data from other basins of the western Southern Alps in order to model the extensional evolution of the western segment of the Lombardian basin. On the base of this reconstruction, a stretching factor has been derived for this segment of the passive continental margin (chapter 4.5).

Subsidence as derived from the sedimentary record is then discussed and compared with the crustal vertical movements predicted by the tectonic-derived stretching factor (chapter 5). The sedimentary and tectonic data sets are used to constrain subsidence curves for the Lombardian basin and then discuss possible geometries of Mesozoic extension of the South-Alpine lithosphere (chapter 5.2). The philosophy of this approach is that extensional faulting in the upper crust and vertical movements are two consequences of the same phenomenon, namely the stretching and extension of a lithospheric plate. How superficial faulting and upper crustal vertical movements are related to each other depends on the mode of extension in the lower crust and in the lithospheric mantle (pure shear vs. simple shear geometries). My prejudice is that only the simultaneous use of tectonic-derived and stratigraphy-derived data can allow a realistic hypothesis on how extension of a passive continental margin occurs.

What is presented here is only a tentative reconstruction of South-Alpine subsidence: I am fully aware of the incompleteness of the tectonic reconstruction, of the stratigraphic columns as well as of the simplicity of the adopted subsidence models. This study is however the first attempt to integrate the different data sets and is worth trying in spite of the risk of being "corrected" in the near future.

2 REGIONAL SETTING, FORMATIONS AND LITHOLOGIES

2.1 Regional setting of the study area

The mapped area is located in the Lombardian Alps and extends from NE of Lugano to the region east of Lake Como. It lies mainly N of the E-W trending Val Menaggio (between Lake Lugano and Menaggio) and terminates towards the N in the crystalline basement (Figs. 1 and 2).

The area is geologically located in the northern part of the sedimentary succession of the Monte Generoso basin (Fig. 2). This is limited towards the west by the Lugano line. The sediments in the area of study are late Permian to Liassic in age and represent the pre- and syn-rift successions of the Monte Generoso basin (hangingwall of the Lugano normal fault). The sediments of the basin assume towards the N a steep S-SW dipping position so that units of increasing age are found moving in the same direction. Two major Alpine N-vergent thrusts cause repetitions of the succession. The contact with the basement is represented by the roughly E-W trending Monte Grona line.

Sedimentary bedding generally dips to the SW-SSW and is therewith oblique with respect to the S-dipping Monte Grona line so that this line cuts upsection moving from E to W. These geometric relationships go hand in hand with the observed stratigraphic omission which increases moving from E to W. While E of Breglia the Monte Grona line is intracrystalline, the sediment/basement contact is stratigraphic and no omission is observed, in the Tesserete region the Triassic is missing and the Early Liassic Moltrasio Limestone lies in direct contact with the crystalline basement (Figs. 2 and 3).

The western part of the Monte Grona line interferes with the Lugano line in the tectonically disrupted and badly exposed region of Tesserete. From this region to Breglia (Fig. 2), the Monte Grona line marks the contact between basement and sediments. East of Breglia the Monte Grona line leaves the sediment/basement contact and becomes intracrystalline. The contact between metamorphic basement and late Permian Verrucano conglomerates in this region is stratigraphic. The Monte Grona line can be followed towards the east to the western shore of Lake Como. The continuation of the Monte Grona line east of the lake is the *Val Grande line* (El Tahlawi, 1965): the two segments of the line show perfectly comparable fault rocks with the same kinematic and structural features.

The interpretation of the Val Grande line being the continuation of the Monte Grona line east of Lake Como has already been proposed by Gianotti and Montrasio (1981) and Gianotti and Perotti (1986). However no evidence has been presented by these authors supporting their interpretation. Other authors have speculatively correlated the Monte Grona line with the Orobic thrust (Rossi, 1975) and with both the Orobic thrust and the Val Grande line (Mottana *et al.*, 1985; Gaetani and Jadoul, 1987).

The crystalline basement is made up of schists and gneisses of different origin mostly related to the Variscan orogeny.

W of the Lugano line, *i.e.* in the footwall of the Lugano normal fault a succession substantially continuous from the late Permian to the Liassic is

preserved. The succession is much thinner than that deposited in the Monte Generoso basin and is characterized by limited subsidence.

2.2 The basement

Premise

It must be first said that the basement of the Lake Lugano-Lake Como region is very poorly known. The Swiss part of the area has been petrographically studied in the fifties and sixties so that good lithologic maps exist, however, with no modern interpretations of metamorphism and deformation. In the basement of the Lake Como region the only mapping and comprehensive study have been carried out by El Tahlawi (1965). In the last decade some contributions concerning mineralogical aspects have been published (Bocchio *et al.*, 1980). The whole area between Lake Como and the Swiss boundary has received practically no attention whatsoever. Because of this and because of different methodologies adopted by the different groups, metamorphic units are often not directly comparable across the whole area.

2.2.1 Lithologies

The basement of the studied region is subdivided in a number of units mainly on the base of tectonic lines and some, generally not well understood, lithologic differences. Two main units are defined north of Lugano: the *Strona Ceneri zone* in the NW and the *Val Colla zone* to the SE; they are separated by the *Val Colla line* and its possible prolongation towards the S, the *Caslano-Taverne line*. The Val Colla zone is continuous towards the east with the *Dervio-Olgiasca zone*. In the Lake Como transect, the Dervio-Olgiasca zone lies between the *Gravedona zone* in the north and the *Monte Muggio zone* in the south (Fig. 2).

The Strona - Ceneri zone

The Strona-Ceneri zone (Serie dei Laghi of Italian authors) is made up of schists and gneisses of sedimentary and magmatic origin; they are interpreted as the product of Variscan deformation and metamorphism affecting a polyphase pre-Ordovician basement with its terrigenous cover (Borghi, 1988 and references therein). Radiometric ages confirm this interpretation: K-Ar dating on micas generally cluster around 310-320 Ma (McDowell, 1970).

Towards the south, close to the Vall Colla line, the gneisses of the Strona-Ceneri zone become mylonitized; mylonites formed under upper greenschist conditions (Zingg, 1987; Bernoulli *et al.*, 1990). The hundreds of meters thick Val Colla fault zone gently dips to the SE. Nothing is known about the kinematics of movements across the line. A polyphase history of deformation is likely: while the mylonites are probably Paleozoic, an Alpine (?) reactivation at least of the Caslano-Taverne line is suggested by obliquity between the foliation of the Strona-Ceneri zone and that of the western Val Colla zone (Fig. 2).

The Val Colla line is generally considered to be of pre-Permian age (Reinhard, 1964 and many others). This because its southern termination (Caslano) (Fig. 2) was thought to be discordantly overlain by Permian conglomerates (Graeter, 1951). However the basal contact of the conglomerates is not exposed so that it cannot be said whether it is stratigraphic or (Alpine) tectonic.

The Val Colla zone

The Val Colla zone is mainly made up of a thick succession of gneisses and schists of sedimentary origin, metamorphosed and deformed under amphibolitic conditions (*Stabbiello Gneiss*) (Reinhard, 1953; Boriani *et al.*, 1974). The metamorphism does not affect the late Carboniferous Manno conglomerates and is therefore older. The only radiometric dating carried out on rocks of the Val Colla zone is based on K-Ar on hornblende and gives an age of 296 Ma (McDowell, 1970).

In spite of the name gneiss adopted in the literature, the typical lithology of the Stabbiello Gneiss is represented by irregularly folded schists. They consist mainly of quartz, plagioclase of intermediate composition (An_{20-30}), muscovite and biotite. While muscovite forms large crystals marking the schistosity, biotite is found in small widespread crystals. Garnet (Fe-rich almandin) is rarely missing. Retrograde (?) chlorite is common often growing in garnet fractures. Staurolite and andalusite are sporadically found pointing to the amphibolitic metamorphic grade of these rocks.

With decreasing muscovite content and increasing portion of felsic minerals, the Stabbiello Gneiss becomes more massive.

The presence of andalusite, staurolite and the high anorthitic content of plagioclase indicate a sedimentary origin of the protolith.

In the monotonous succession of the Val Colla zone leucocratic orthogneisses stand out which are referred to as *Gneiss Chiari*.

The Gneiss Chiaro (Bernardo Gneiss of Reinhard, 1953; Gneiss Chiari del Corno Stella of Liborio and Mottana, 1971) is a massive, coarse-grained leucocratic gneiss. A clear foliation is rarely developed; in some outcrops a strong linear fabric is observed. The mine-

ralogical assemblage is simple with quartz, albitic plagioclase (An<6-7%), K-feldspar and muscovite. The plagioclase typically shows nicely developed albite and Carlsbad twinnings. The K-feldspar also shows common twinning often with a microcline pattern. Biotite is rarely found. The microstructure is generally well equilibrated. Quartz grains show often undulose extinction and some deformation bands. Dynamic recrystallization is never observed. Cataclasis, in contrast, is very common.

The nature of the protolith is debated. Interpretations range from arkose (El Tahlawi, 1965) to Carboniferous plutonites (De Sitter and De Sitter-Koomans, 1949; Köppel and Grünenfelder, 1971) or rhyolites (Boriani and Colombo, 1979).

The Gneiss Chiari are mainly found in the southern part of the Val Colla zone and generally form flat, up to a few hundred meters thick tabular bodies lying on the Stabbiello Gneisses. The contact is at least tectonized (*e.g.* in the Piancabella region, in the upper Val Colla etc). Also the upper contact is often tectonic. The only preserved stratigraphic cover consists of Verrucano conglomerate (chapter 2.2.2). The Gneiss Chiaro is typically the highest crystalline unit.

The Dervio - Olgiasca zone

The Val Colla zone is continuous towards the east with the Dervio-Olgiasca zone defined by El Tahlawi (1965) along the Lake Como transect ("low-pressure unit" of Bocchio *et al.*, 1980). The Dervio-Olgiasca zone is made up of rocks mineralogically well comparable to the schists of the Val Colla zone. Both towards the south and towards the north, the Dervio-Olgiasca zone is bounded by two steeply dipping stripes of fault rocks ranging from lower greenschist mylonites to cataclasites. The southern band is associated with the Val Grande line, the northern one with the Musso line (Fig. 2).

The fault rocks associated to the Musso line are generally included in the Gravedona zone (El Tahlawi, 1965; Bocchio *et al.*, 1980). I consider them to be part of the Dervio-Olgiasca zone since they are very similar to those of the Val Grande line and play the same kinematic role (see chapter 3.3).

The Dervio-Olgiasca zone is considered to consist of schists and gneisses with a N-wards increasing metamorphic grade (Fig. 4) (El Tahlawi, 1965; Bocchio *et al.*, 1980). The sequence would begin in the south with the chlorite zone (following El Tahlawi) or with the biotite zone (Bocchio *et al.*, 1980). Two-micas, staurolite-bearing gneisses are found north of Dervio. Sillimanite appears moving northwards; there, the typical paragenesis consists, beside of sillimanite, of quartz, plagioclase (An_{20}), muscovite, biotite, garnet and staurolite.

At this stage, however, it is very difficult to know if the minerals used to separate the different zones are in fact related to the same metamorphic event since no microscopic structural analysis has been presented. The chlorite described by El Tahlawi in the southern Dervio-Olgiasca zone for instance, is clearly related to lower greenschist mylonitization in connection with Mesozoic movements along the Val Grande line (chapter 4.3.3). The Dervio-Olgiasca zone itself has apparently a metamorphic history much more complex than traditionally described. Mottana *et al.* (1987) report relic kyanite in a sillimanite-bearing rock. In the same rocks the presence of andalusite in quartz nodules has long been known (Repossi, 1910). Further, a post-kinematic event overprinting the Variscan foliation and paragenesis is shown by albitic rims overgrowing plagioclase of intermediate composition (Bocchio *et al.*, 1980). As a matter of fact, the structural geology and the petrography of the area need to be investigated and revised respectively.

The Dervio-Olgiasca zone is particularly important because it shows, in contrast to the other surrounding units, Late Permian to Early Jurassic radiometric ages (Mottana *et al.*, 1985 and references therein) (Fig. 4).

Pegmatitic dykes are found in the northern part of the Dervio-Olgiasca zone (El Tahlawi, 1965). The dykes often intruded the mylonites forming the northernmost part of the Dervio-Olgiasca zone and have produced Late Permian to Triassic radiometric ages (Hanson *et al.*, 1966).

The Gravedona zone

The *Gravedona zone* (Bocchio *et al.*, 1980; Domaso-Cortafo zone of Fumasoli, 1974; nördliche Phyllonitzone of El Tahlawi, 1965) lies north of the *Musso line* and is made up of a heterogeneous, highly deformed and poorly known complex of gneisses of amphibolite grade.

The micaschists and gneisses of the Gravedona zone macroscopically resemble the Stabbiello Gneiss. The mineralogical assemblage is made up by quartz, plagioclase (An_{20-30}), muscovite, biotite, garnet and staurolite. Kyanite is sporadically found (Bocchio *et al.*, 1980). The rocks of the Gravedona zone are typically highly deformed; hydrothermal alteration is common.

Since the relations between the Val Colla and Musso lines are unknown, the relationships between the Gravedona and Strona-Ceneri zones are unclear.

The Monte Muggio zone

S of the Val Grande line appears the wedge-shaped *Monte Muggio zone* ("high-pressure unit" of Bocchio *et al.*, 1980). The typical lithology is a

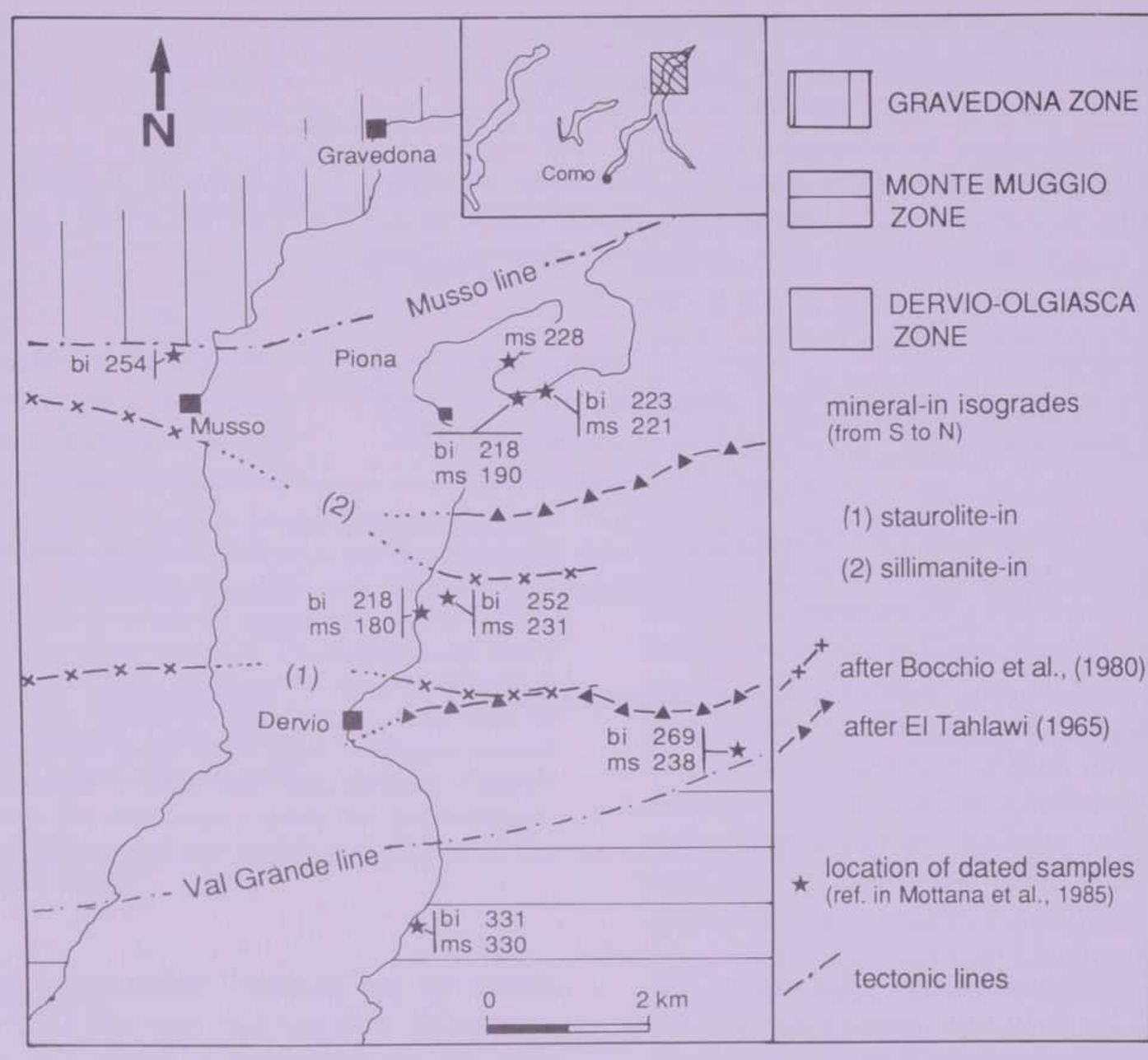

FIG. 4 - Metamorphic isogrades and published K-Ar radiometric ages in the northern Lake Como area. Note Mesozoic ages in the Dervio-Olgiasca zone juxtaposed to the Variscan ages of the M. Muggio zone. bi = biotite; ms = muscovite.

fairly massive, intensively fractured gneiss of amphibolite grade (*Monte Muggio Gneiss*). The petrographic similarity between the Monte Muggio Gneiss and the gneisses of the Gravedona zone has been emphasized by BOCCHIO *et al.* (1980).

The usual paragenesis of the Monte Muggio Gneiss consists of quartz, plagioclase, muscovite, biotite, staurolite and kyanite. The microstructure is generally well equilibrated. However in some samples near the Val Grande line, an incipient dynamic recrystallization is observed; undulose extinction and deformation bands are common. Characteristic is the ubiquitous brittle deformation.

Light-coloured orthogneisses perfectly identical to the Gneiss Chiari of the Lugano region and in a similar tectonic position are found E of Breglia. They tectonically overlie the Monte Muggio Gneiss and stratigraphically underlie the Permian conglomerates.

Also part of the Monte Muggio zone is a body of two-micas-orthogneisses limited to the NW by the Val Grande line (*Monte Legnone Gneiss*, EL TAHLAWI, 1965).

The Monte Legnone Gneiss outcrops along a NE-SW elongated zone running through Monte Legnone and is made up of coarse-grained augengneiss. Major constituents of the rock are quartz, plagioclase (An_{20-30}), K-feldspar, biotite and muscovite. Characteristic for the Monte Legnone Gneiss are the large augen of typically microcline twinned K-feldspars. Myrmekitic and perthitic structures are also very common. K-feldspars are often poikiloblastic with inclusions of muscovite, garnet etc. These features are of particular importance because very similar feldspars are found in the mylonites of the Val Grande fault zone. In fact, the Monte Legnone Gneiss is the protolith of the lower greenschist facies mylonites found along the Monte Grona/Val Grande fault zone (chapter 4.3.4). Mylonites parallel to the Val Grande line are mentioned by EL TAHLAWI (1965) in the western part of the body.

On the base of its large grain-size and of mineralogical assemblage the Monte Legnone Gneiss can be interpreted as of plutonic origin.

2.2.2 Summary of basement evolution

The main metamorphic event in the studied area, as generally in the western Southern Alps, is Variscan (BORIANI *et al.*, 1974). In some regions,

TABLE 1 - Main features of basement units (see text)

Val Colla zone	Dervio-Olgiasca zone	M. Muggio & Gravedona zones
- Variscan paragenesis substantially preserved.	- Variscan assemblage strongly overprinted: kyanite only as relic, albite overgrowth on plagioclase etc.	- Variscan paragenesis substantially preserved.
- Variscan microfabric is well preserved: only brittle deformation after Variscan metamorphic phase.	- Lower greenschist mylonitization affects the northern and southern parts of the zone (Musso and Val Grande lines).	- Variscan microfabric is well preserved: only brittle deformation is recorded after the main Variscan metamorphic event.
- Radiometric dating (K-Ar on hornblende) gives Variscan ages.	- Radiometric ages (K-Ar on micas) are Triassic to Early Jurassic.	- Radiometric dating (K-Ar on micas) produced only Variscan ages.
- Sedimentary cover only in the western part (Lugano, M. San Giorgio).	- No sedimentary cover.	- Sedimentary cover in both zones (Val Muggiasca and Sasso Pelo respectively).

although not in the studied area, a Caledonian relic paragenesis has been postulated (BORIANI *et al.*, 1974). In the study area the Variscan event is of amphibolite grade (stability field of sillimanite, andalusite, staurolite and intermediate plagioclase). The dominant schistosity and the widespread isoclinal folding are related to the Variscan orogeny. Not much is known about the major structures related to the Variscan orogeny. Horizons of high-temperature mylonites (upper greenschist facies) like those found along the Val Colla line (ZINGG, 1987), in the Val Colla zone itself and, possibly, along the Monte Grona and Val Grande lines (chapter 4.3.2) are probably of Variscan age. The distribution and kinematics of the mylonites are unknown.

The Variscan metamorphism ended before the Late Carboniferous when the Manno Conglomerate was deposited. At this time the western Val Colla zone, the Monte Muggio and the Gravedona zones were lying at or close to the surface. This is required by the stratigraphic contact with the overlying volcanics and sediments (1) and is compatible with the Variscan ages and with the absence of post Variscan thermal and deformational overprinting in the metamorphic rocks of the mentioned zones (see table 1).

In contrast to the units mentioned above, the Dervio-Olgiasca zone must have been lying at deeper crustal levels at the end of the Variscan as shown by the younger radiometric ages; this is compatible with the fact that the Dervio-Olgiasca zone has no sedimentary cover. On the base of these observations the position of the different units at the end of the Variscan orogeny can be reconstructed and is schematically represented in figure 5. The juxtaposition of the Dervio-Olgiasca zone with the Monte Muggio and Gravedona zones occurred during the Early Mesozoic along the Monte Grona/Val Grande fault zone (across which the age jump is observed) (further discussion in chapter 4.3.5). The kinematics required to achieve the juxtaposition is that of a normal fault. This is the Monte Grona/Val Grande fault zone, i.e. the northern and deep segment of the Lugano normal fault.

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(1) It is very interesting to note that the only stratigraphic base of the Verrucano conglomerates known in central and western Lombardy is made up of Gneiss Chiaro. This is true in the area of study, north of Lake Como (FUMASOLI, 1974) and also east of the lake (G. SCHÖNBORN personal communication, 1988; DALLAGIOVANNA *et al.*, 1986). It could also be the case in the central Austroalpine Ortler zone (N. FROITZHEIM, personal communication; DÖSSEGGER, 1974). This is compatible with the fact that clasts in the lower part of the Verrucano conglomerates are mainly made up of orthogneisses while paragneisses are rare. The extension of the area covered by Gneiss Chiari therefore must have been much larger than what is presently visible.

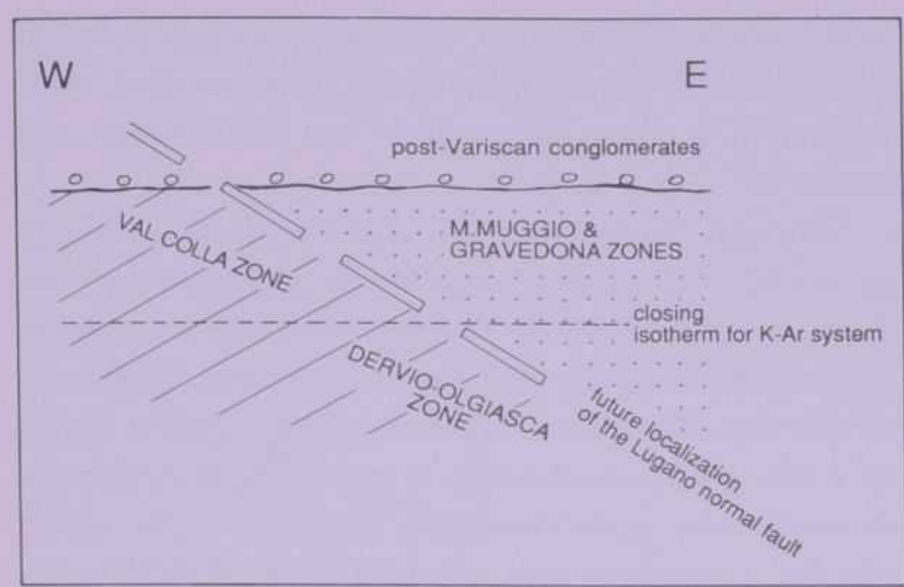

FIG. 5 - Schematic illustration of the relationships between the different basement units in the late Paleozoic after the end of the Variscan orogeny. The Val Colla, M. Muggio and Gravedona zones must have been at the earth surface since they are stratigraphically overlain by sediments. These zones show no post-Variscan overprinting. The Dervio-Olgiasca zone, lying east of the Val Colla zone, must have been at deeper crustal levels. Movements along the east-dipping Lugano normal fault in the Mesozoic caused the presently observed juxtaposition and the uplift and cooling of the Dervio-Olgiasca Zone.

2.3 THE SEDIMENTS

2.3.1 Summary of the sedimentary evolution of the Lombardian basin

The oldest sediments found in the western Southern Alps are the late Carboniferous Manno Conglomerate. They result from the erosion of the Variscan orogen. The deposits record continental environments and are found in limited and tectonized outcrops. Only in the Bergamasc Alps large NE-SW trending grabens accommodated huge thicknesses of slightly younger clastic and volcanic deposits (ASSERETO and CASATI, 1965).

The first sediments widespread over most of the South-Alpine domain are the clastics of the Servino-Verrucano Serie (LEHNER, 1952) and of the Bellano Formation (GAETANI *et al.*, 1986). The sediments are of late Permian to Early-Middle Triassic age and still record the erosion of the Variscan mountains; they were deposited on wide fluvial plains grading towards the east into marine environments (ASSERETO *et al.*, 1973). On the whole, these sediments become thicker from the west towards the east.

The studied area is localized in the transition zone between the western Lombardy/Piedmont area where Permian clastic sediments are thin to absent, and eastern Lombardy where the same sediments rapidly increase in thickness. The erosion of the Variscan basement went on until the late Anisian. At the same time the sea transgressed over the Southern Alps moving from the east to the west.

At the beginning of the Ladinian marine conditions were established all over the presently exposed Southern Alps: carbonate platforms with intraplatform basins are the typical paleogeographic features. BRUSCA *et al.* (1981) recognize for the Ladinian of the Southern Alps two major, E-W trending paleogeographic zones. The northern area, corresponding to the present-day outcrops is characterized by carbonate platforms and basins. The southern area, referred to as Southern mobile belt, is presently buried beneath the Po Plain sediments and was characterized by basement and volcanic complexes exposed at the surface. The E-W trend of the paleogeographic zones suggests that the N-S striking normal faults which will characterize the rifting stage, were not yet active. The volcanoes of the Southern mobile belt were the source for the acid to subalkalic tuffs found at different levels of the Ladinian carbonate succession (JADOUL and ROSSI, 1982). The geodynamic setting of this volcanism is far from being understood with interpretations ranging from a compressional context (CASTELLARIN *et al.*, 1988) to an extensional one (FERRARA and INNOCENTI, 1974) or transtensional/transpressive one (DOGLIONI, 1984).

There is no conclusive evidence demonstrating the existence of morphological highs with crystalline basement at the surface in the region between Lake Como and Lake Lugano as proposed by GIANOTTI (1984), GAETANI *et al.* (1986) and GIANOTTI and TANNOIA (1988). This hypothesis is mainly based on the absence along part of the Monte Grona line of Early to Middle Triassic sediments. However since the stratigraphic omission is of tectonic origin, it cannot be used to demostrate the existence of a morphological high (see introduction).

The Ladinian carbonate platforms became extinct in the Carnian, probably because of the onset of erosion in the Southern mobile belt. Large amounts of volcanic and basement-derived debris were shed towards the north forming large sandstone lobes interfingering with shaly and carbonate sediments (GARZANTI, 1985; GAETANI *et al.*, 1986). Evaporitic bodies are locally found.

The end of the erosion of the volcanic belt in the south allowed, at the beginning of the Norian, the reinstallation of a carbonate platform in the whole Lombardian basin (Dolomia Principale). Norian formations show, at the scale of the future margin, a thickness increase from the west, where they are even missing in the M. Fenera (RASETTI, 1897) and Gozzano areas (MONTANARI,

1969) to the east (see Fig. 1 in BERNOULLI *et al.*, 1979). At a more detailed look, however, the existence of domains with different subsidence rates is clearly recognized. The abruptness of the thickness changes indicates that they were fault-controlled. In the central Lombardy area, in which we are particularly interested, domains are identified with high subsidence rates persisting through the rest of the Triassic and the Liassic. We therefore consider the Norian as the beginning of rifting in the Lombardian zone of the Southern Alps. The major strongly subsiding zones in the Norian were the Monte Generoso basin and the Sebino trough. In both cases their development was controlled by roughly N-S striking normal faults. Some of these faults caused morphological and facies changes, others did not (1). In the former case, morphologic basins developed in which thick sequences of mainly platform-derived resedimented breccias to arenites and dark, often bituminous carbonates are found. Norian basins of this kind have been recognized not only in the area of study (LUALDI and TANNOIA, 1985; BERTOTTI, 1990) but also in the Bergamasc Alps (JADOUL, 1985), west of Lake Garda (PICOTTI and PINI, 1988) and, although only on the base of sediment facies and thickness, in the Lake Iseo area (CASATI, 1964).

The extensional tectonic pattern remained relatively constant in the Rhaetian. The sedimentological picture, on the contrary, changed dramatically: the introduction of very abundant terrigenous mud (sedimention rates exceeding 500 m/Ma are proposed by MASETTI *et. al.*, 1989 in the Lake Iseo area) caused the extinction of the pre-existing carbonate platforms. The abundant terrigenous input fossilized most of the pre-existing morphology. Only on a few morphological highs, peritidal carbonate sedimentation could continue. The transition between highs and lows was made up of very smooth and gentle ramps so that no coarse-grained resedimented deposits are found. Surprisingly enough, the cause of the dramatic change from the Norian to the Rhaetian has been fully overlooked in the literature. Large-scale climatic events have probably to be envisaged. Also unknown is the source area for the terrigenous material.

The Rhaetian west of the Giudicarie line is generally subdivided into three formations: the Riva di Solto Shale at the bottom, the Zu Limestone in the middle and the Conchodon Dolomite at the top. The Rhaetian formations in the deeper water successions show a quite regular increase in carbonate from the Riva di Solto Shale up to the Zu Limestone and to the Conchodon Dolomite which consists only of carbonates.

This overall trend of increasing carbonate content is repeated at a metric to decametric scale in cycles made up of a lower shaly portion, a central portion with marl-limestone couplets with the limestones thickening upwards, and a wholly carbonatic top (Plate I, Fig. 1). Limestones are considered to be allochthonous and derived from platforms east of the Giudicarie line. These cycles, which are typical for the transition from Riva di Solto Shale to Zu Limestone, are interpreted as resulting from the superposition of a low-frequency signal related to eustatic changes and a high-frequency signal of climatic origin (MASETTI *et al.*, 1989).

At the end of the Rhaetian, shallow water conditions were again established all over the Lombardian basin. Many, but by no means all of the Norian-Rhaetian faults seem to die out and the upper Rhaetian formations are laterally relatively continuous.

This has been interpreted by some authors (JADOUL, 1985; GAETANI *et al.*, 1986) as indicative of a quiescence phase separating two rifting pulses. Mainly on the base of my observations in the study area, I rather interpret these kinematic changes as due to a concentration of strain along the major faults (Lugano fault in the area of study) leading to the disactivation of the smaller, more superficial normal faults (chapter 4.2).

Extension went on in the Early Liassic and the well known thick succession of basinal siliceous limestones (Moltrasio Limestone) was deposited. Partly due to active normal faulting and partly to external factors the basinal areas are brought during the earliest Liassic below the photic zone. In the late Early to Middle Liassic, also the higher zones were drowned.

In the Pliensbachian the entire Lombardian basin was under deep waters. However the sea-floor still showed a varied morphology. Toarcian Rosso Ammonitico facies are found on the highs and their flanks while turbiditic deposits fill up the intervening lows (KÄLIN and TRÜMPY, 1977 for the Monte Nudo area; GAETANI, 1975 for central Lombardy). The Rosso Ammonitico and the turbidites seem to be laterally continuous and not displaced by normal

(1) It is unfortunately a still very common assumption that tectonic movements can, alone, cause the drowning of a carbonate platform. SCHLAGER (1981) showed on the contrary that a carbonate platform is capable to keep pace with the most rapid sea-level changes if ecologic conditions remain favorable. It seems that a platform can be drowned only if the ecological conditions are disturbed in such a way that organisms on the platform cannot react to the relative sea-level change. These disturbances can be associated to the fault itself, *i.e.* local, but can also be related to large-scale (climatic?) phenomena.

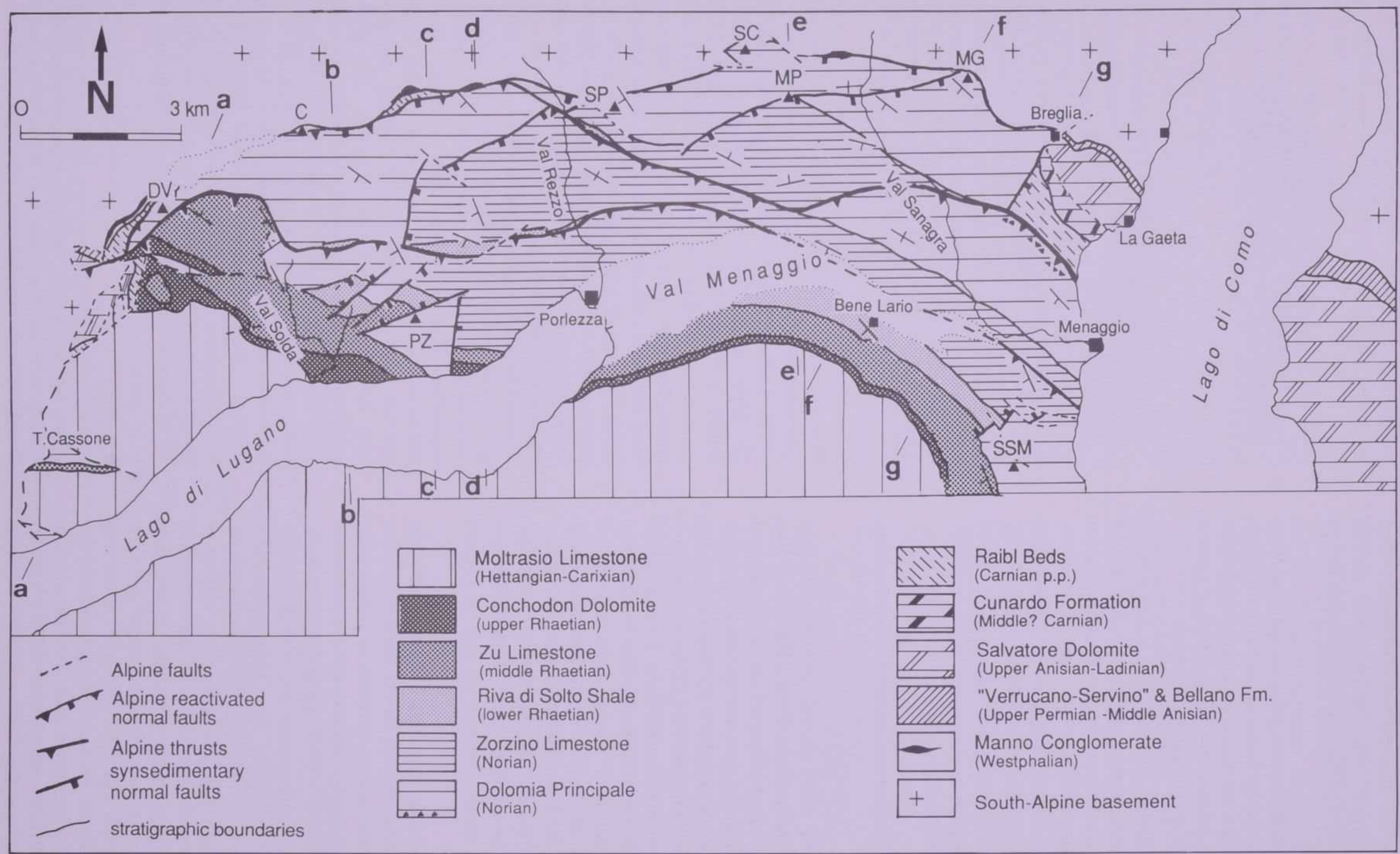

FIG. 6 - Geologic map of the study area. Triangles at the base of the Dolomia Principale N of Menaggio indicate the occurrences of Ligomena Breccia. DV = Denti della Vecchia; C = Catelina; PZ = I Pizzoni; SP = Sassi della Porta; SC = Sassi di Cusino; MP = Monte Pidaggia; MG = Monte Grona; SSM = Sasso di S. Martino. The trace of the geologic profiles of figure 16 is indicated.

faulting (see for instance GAETANI, 1975) (Fig. 45). This suggests that rifting-related normal faulting in the Lombardian basin ceased before the Toarcian.

In the Middle Jurassic, pelagic conditions were installed all over the Lombardian basin. Sedimentary successions are however variable with local condensed sequences (BERNOULLI, 1964; GAETANI, 1975; WINTERER and BOSELLINI, 1981).

In the Callovian, siliceous ooze is deposited all over the region west of the Giudicarie line; the radiolarites are followed by the Maiolica Formation (WEISSERT, 1979).

The passive margin evolution of the Southern Alps is interrupted in the middle Cretaceous by the onset of terrigenous sedimentation related to the first compressional episodes in the Austroalpine area (CASTELLARIN, 1976).

2.3.2 Stratigraphy in and around the area of study

General remarks

Sedimentary units in the area of study are found south of the Monte Grona/Val Grande line (Fig. 6). We have interpreted the Monte Grona/Val Grande line as a tectonic lineament and more specifically as the northern prolongation of the Late Triassic-Early Jurassic Lugano normal fault (Fig. 3). The Late Triassic/Early Jurassic age of the Monte Grona/Val Grande line implies that the Late Permian to Triassic sediments are allochthonous with respect to the basement north of the line. Palinspastically restoring the movements along the line, the late Permian to Middle Triassic succession outcropping along the western Lake Como shore comes to lie east of Lugano (Fig. 3).

Upper Permian to middle Triassic sediments are also found in tectonic slivers along the Monte Grona line. The slivers are made up of basement rocks (mainly Gneiss Chiaro) and Carboniferous to Middle Triassic sediments and have tectonic contacts with the basement and with the overlying Dolomia Principale. They form bodies elongated parallel to the Monte Grona line; bedding however dips to the SW and is therefore oblique to the line and, at the same time, parallel to bedding in the Dolomia Principale south of the line. These slivers are tectonic units ripped off from the hangingwall during its downward movement along the Lugano normal fault (extensional duplexes) and not transgressive deposits as interpreted by VENZO and MAGLIA (1949), LEHNER (1952) and others (Fig. 7) [1].

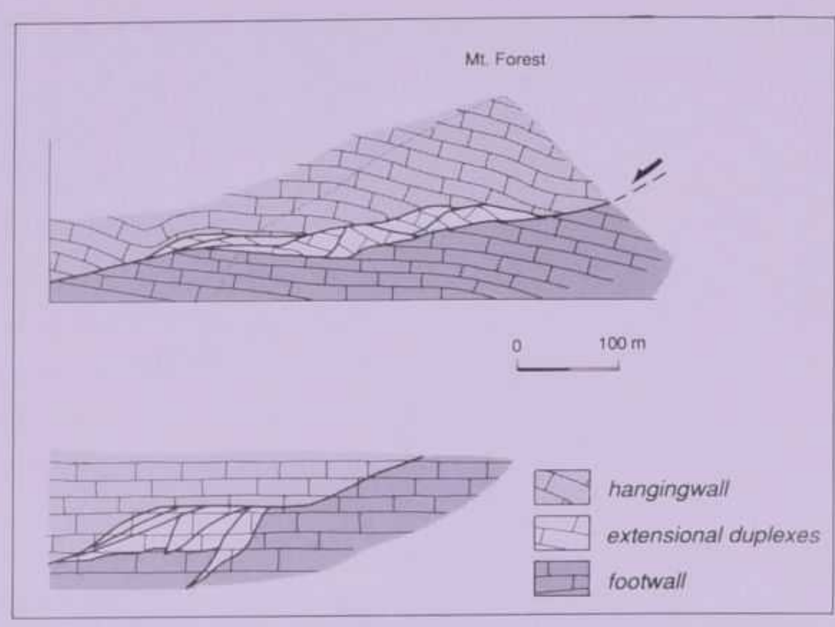

FIG. 7 - Extensional duplexes as observed by ROOT (1990) in the Purcell Mountains of southeastern British Colombia; below, their palinspastic reconstruction. Note how the duplexes form bodies elongated parallel to the fault; bedding however is clearly oblique to it and parallel to bedding in the hangingwall. The same features are observed along the Monte Grona fault zone (see for instance Fig. 41).

Manno Conglomerate (Westphalian)

The Manno Conglomerate (NEGRI and SPREAFICO, 1869) is the oldest sedimentary formation found in the western Southern Alps and is the product of erosion of the Variscan mountain chain. Outcrops west of the Bergamasc Alps are very scattered and usually pinched along tectonic lines. Because of the high content in plant remains and the absence of other fossils the Manno Conglomerate is considered to be deposited in a continental environment.

The Manno Conglomerate is made up of conglomerates and sandstones to siltstones. Conglomerates are poorly sorted and matrix- to grain-supported. The clasts are often rounded, cm-to dm-sized and consist only of metamorphic rocks (mainly leucocratic muscovite-bearing gneisses very similar to the Gneiss Chiari). The fine-conglomeratic to sandy matrix has a similar composition. The sandstone is relatively well-sorted and grain-supported. The clasts are typically 0.1-0.3 mm large and angular. They consist of quartz, feldspar, muscovite and subordinate biotite. The siltstones have similar composition and show abundant plant remnants. Bedding is often irregular and some channeling is observed. Because of

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[1] Other tectonic slices are found along the Monte Grona line which cause stratigraphic repetitions. Differently from the ones described above, they formed during Alpine shortening.

poor exposures it is however difficult to recognize in the Manno Conglomerate large-scale sedimentary structures.

Well preserved floras have been found in the Manno Conglomerate of the Val Colla and of the Val Sanagra (VENZO and MAGLIA, 1947), as well as at the type locality of Manno. A Stephanian age has been suggested by SCHMIDT (1920) for the outcrops of the Val Colla and of Manno. More recently a late Westphalian age was proposed for the floras of the Val Colla, of the Val Sanagra and of Manno (JONGMANS, 1960).

The Manno Conglomerate is found in the area of study in small, strongly tectonized outcrops along the Monte Grona line (VENZO and MAGLIA, 1947; LEHNER, 1952). The lower and upper contacts are tectonic in all outcrops.

Of the several outcrops described by VENZO and MAGLIA (1947) only those in Val Sanagra, south of Cimadera and north of the Denti della Vecchia are still accessible (Fig. 6). Outcrop conditions are however much worse than described by VENZO and MAGLIA (1947) and by LEHNER (1952). Some of the outcrops described by VENZO and MAGLIA (1947) as Manno Conglomerate, actually consist of a fine-grained fault gouge.

The thickness of the Manno Conglomerate varies from 0 to 15 m in the Val Sanagra (VENZO and MAGLIA, 1947), to 20-30 m in the Val Rezzo and in the upper Val Colla (LEHNER, 1952) to ca. 100 m beneath the Denti della Vecchia (LEHNER, 1952).

Servino-Verrucano Serie and Bellano Formation (Late Permian? - Late Anisian)

A mainly clastic sequence known as Servino-Verrucano Serie and Bellano Formation was deposited in the Lake Lugano-Lake Como region beginning from the Late (?) Permian. The clastic sequence underlies the Late Anisian - Ladinian carbonates.

The denomination *Servino-Verrucano* is generally used to indicate the predominantly clastic sequence beneath the Early Anisian dolomitic/evaporitic horizon of the Carniola di Bovegno. As recommended by the Lexique Stratigraphique International (7c/II p. 1076-1077) we use Verrucano and Servino only as informal lithologic terms. A stratigraphic differentiation between the Verrucano Lombardo and the overlying Servino is possible E of Lake Como but it is not applicable west of the lake where the two lithologies are irregularly interbedded (LEHNER, 1952). The only paleontological findings in the Servino-Verrucano succession come from its base and seem to indicate the Scythian (REICH, 1912; FRAUENFELDER, 1916). The Servino-Verrucano Serie terminates east of Lake Como with the dolomites and evaporites of the Carniola di Bovegno (Fig. 8). Above this horizon, a clastic succession recently named *Bellano Formation* (GAETANI, 1982; FARABEGOLI and DE ZANCHE, 1984; GAETANI *et al.*, 1986) is found. The succession is lithologically quite similar to the Servino-Verrucano Serie. Since the Carniola di Bovegno, which separates the two formations, is considered to be Early Anisian, the Bellano Formation should be Middle Anisian.

The clastics of the Servino-Verrucano Serie are derived from the erosion of the Variscan chain and of the volcanics of the Lugano region. At a regional scale, the Servino-Verrucano Serie represents a transgressive sequence lying on the Variscan basement and on Permian volcanics.

The Servino-Verrucano Serie is mainly made of coarse-grained conglomerates and sandstones deposited in alluvial plains with anastomosing systems (ASSERETO *et al.*, 1973). The conglomerates, informally called Verrucano, are typically reddish, matrix-supported and poorly sorted. Clasts are cm-sized, fairly well rounded and made up of lithoclasts of orthogneiss (Gneiss Chiaro)

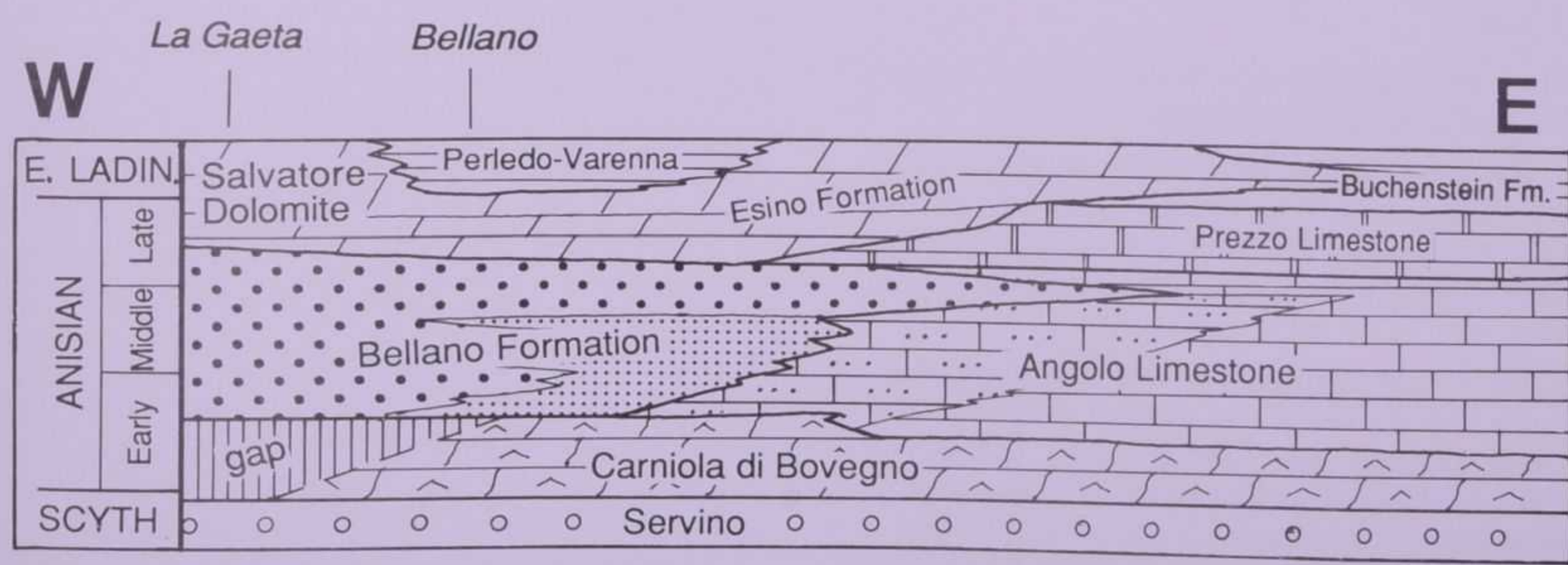

FIG. 8 - Stratigraphic relationships of the Scythian to Ladinian formations in the Lake Como region. Modified after GAETANI *et al.*, (1986). Note the two clastic successions separated by the Carniola di Bovegno. The upper succession, the Bellano Formation, interfingers towards the east with the Angolo Limestone.

and of volcanics, mainly rhyolites of the Lugano succession. The sandy matrix typically consists of quartz, feldspars and muscovite. The Verrucano conglomerates often show erosional bases. Associated to the conglomerates, reddish sandy marls are sometimes found.

The sandstones, informally known as Servino, are yellowish-weathering, well sorted and grain-supported. The clasts are angular, typically 0.1-0.3 mm large and made up by quartz with subordinate feldspars and muscovite. Bedding thickness is generally from cm to dm.

Intercalated at different levels of the Servino-Verrucano Serie, dolomitic horizons are found which mark short-lived marine ingressions. The most significant of these events is found at the top of the Servino-Verrucano succession and led to the deposition of the Carniola di Bovegno. This marine event, well recorded east of Lake Como separates the Verrucano-Servino succession from the Bellano Formation. In the studied area no Carniola di Bovegno is found; the two clastic sequences must then be separated by a major unconformity which however has not been identified with certainty (Fig. 8).

The Bellano Formation is mainly made up of conglomerates and sandstones lithologically similar to those of the Servino-Verrucano Serie. Also the components are comparable (GAETANI *et al.*, 1986). The depositional environment remains that of alluvial plains open to marine ingressions coming from the east. These events are again marked by dolomitic intercalations which become more common and thicker from the western shore of Lake Como towards the east. In the Grigna region, east of the lake, the Bellano Formation interfingers with the marine carbonates of the Angolo Limestone (GAETANI *et al.*, 1986) (Fig. 8).

The Verrucano-Servino Serie and the Bellano Formation in the area of study

The clastic succession has two main occurrences: along a stripe from Breglia to Lake Como and in slivers tectonically emplaced along the Monte Grona line (Fig. 6) (see general remarks to this chapter).

The best stratigraphic section is found S of La Gaeta (Fig. 6) along the western shore of Lake Como (ESCHER v.d. LINTH, 1853; TORNQUIST, 1902; LEHNER, 1952; GIANOTTI, 1968; GAETANI *et al.*, 1986).

The base of the succession is no longer exposed along this profile. It is, however, visible in several outcrops further to the west; it is always stratigraphic and marked by an erosional surface. The Gaeta profile begins with a few meters of Verrucano conglomerates, continues with some 30 meters of Servino sandstones and some 20 m of sandstones with dolomitic intercalations. This is considered by GAETANI *et al.* (1986) to be the Servino-Verrucano Serie, *i.e.* the Late Permian to Early Triassic part of the succession. The overlying 80-90 meters are attributed to the Bellano Formation and consist mainly of conglomerates which can be organized in fining-upward sequences (GAETANI *et al.*, 1986). The contact at the top with the well-bedded base of the Salvatore Dolomite is gradational but very rapid: within a few decimeters the quartzitic clasts disappear and peritidal carbonates are found.

The La Gaeta section is representative for the whole stripe east of Breglia. The thickness of the clastic succession decreases from ca. 120 meters at La Gaeta to 50 meters near Breglia. This decrease in thickness is probably not due to tectonic reasons since both the lower and the upper contact of the clastic succession are stratigraphic.

The clastic succession west of Breglia forms small, poorly exposed and strongly tectonized outcrops scattered along the Monte Grona line. The lithologies are the same as those described in the Gaeta section.

In the upper Val Rezzo not much can be seen now of the profiles described by LEHNER (1952): only some Servino-like blocks are found in the detritus.

Further to the west, outcrops of the clastic succession are found at Catelina (coordinates 724.70/102.87; Fig. 6) and north of the Denti della Vecchia. The former outcrops, although small, are interesting because they show quartzitic sandstones passing stratigraphically to the overlying Salvatore Dolomite. North of the Denti della Vecchia two parallel streamlets expose a tectonized succession mainly made up of Servino-like lithologies. Outcrop conditions are, however, much worse than described by LEHNER (1952).

Other outcrops, mainly of Verrucano-like conglomerates, are found in the Tesserete region however in a very unclear tectonic position (Fig. 6). At the base of the profile a few meters of volcanics have been found (LEHNER, 1952). In none of these outcrops a stratigraphic contact with the basement (here represented by Stabbiello Gneiss) can be demonstrated.

Salvatore Dolomite and Cunardo Formation
(Late Anisian-Early? Carnian)

Marine conditions are firmly established in the Late Anisian all across the presently exposed Southern Alps. Carbonate platforms with intervening basins characterize the paleogeography of the Lombardian basin (chapter 2.3.1).

The variety of Ladinian sedimentary environments, together with the burden of local traditions have pro-

duced an inflation of formational names. Following BERNOULLI (1964) I will use the term Salvatore Dolomite to indicate the entire carbonate complex of Late Anisian to Ladinian age between the clastic succession at the bottom and the Cunardo Formation at the top. The Cunardo Formation (ALLASINAZ, 1968) is made up by well bedded carbonates and passes upwards into the Raibl Beds.

The age of the base of the Salvatore Dolomite is Late Anisian (FARABEGOLI and DE ZANCHE, 1984). The ages of the top of the Salvatore Dolomite and of the Cunardo Formation are not known with certainty. The Cunardo Formation could be Early Carnian (Sheet Lugano of the Geologic Atlas of Switzerland; KÄLIN and TRÜMPY, 1977) or, on the base of a correlation with the uppermost Meride Limestone (BERNOULLI, 1964) Late Ladinian (GAETANI *et al.*, 1986; GIANOTTI and TANNOIA, 1987).

LEHNER (1952) considers the well-bedded bituminous carbonates (later named Cunardo Formation) to be part of the Raibl beds and therefore of Carnian age.

In the Lake Como/Lake Lugano region two main Late Anisian to Ladinian carbonate platforms are recognized: one in the Grigna mountains and one in the Lugano region.

The former platform (GAETANI *et al.*, 1986; GIANOTTI and TANNOIA, 1988) outcrops in the eastern part of the Grigna northern thrust sheet and is mainly made up by massive bioclastic packstones to grainstones (Esino Limestone) (Fig. 9). Well bedded, sometime bituminous limestones belonging to the Cunardo Formation are found towards the top. The middle and upper parts of the Esino platform pass laterally towards the west to well bedded, sometimes bituminous limestones with abundant slumping and platform derived sediments (Perledo-Varenna Limestone) (Figs. 8 and 9). Chert nodules are found towards the top. These limestones are indicative of a local depression with restricted circulation west of the Esino platform.

The Late Anisian-Ladinian paleogeography of the Lugano region is characterized by a carbonate platform in the north (Monte San Salvatore, Campione d'Italia. Fig. 9) which passes towards the south into the well bedded, mainly calcareous

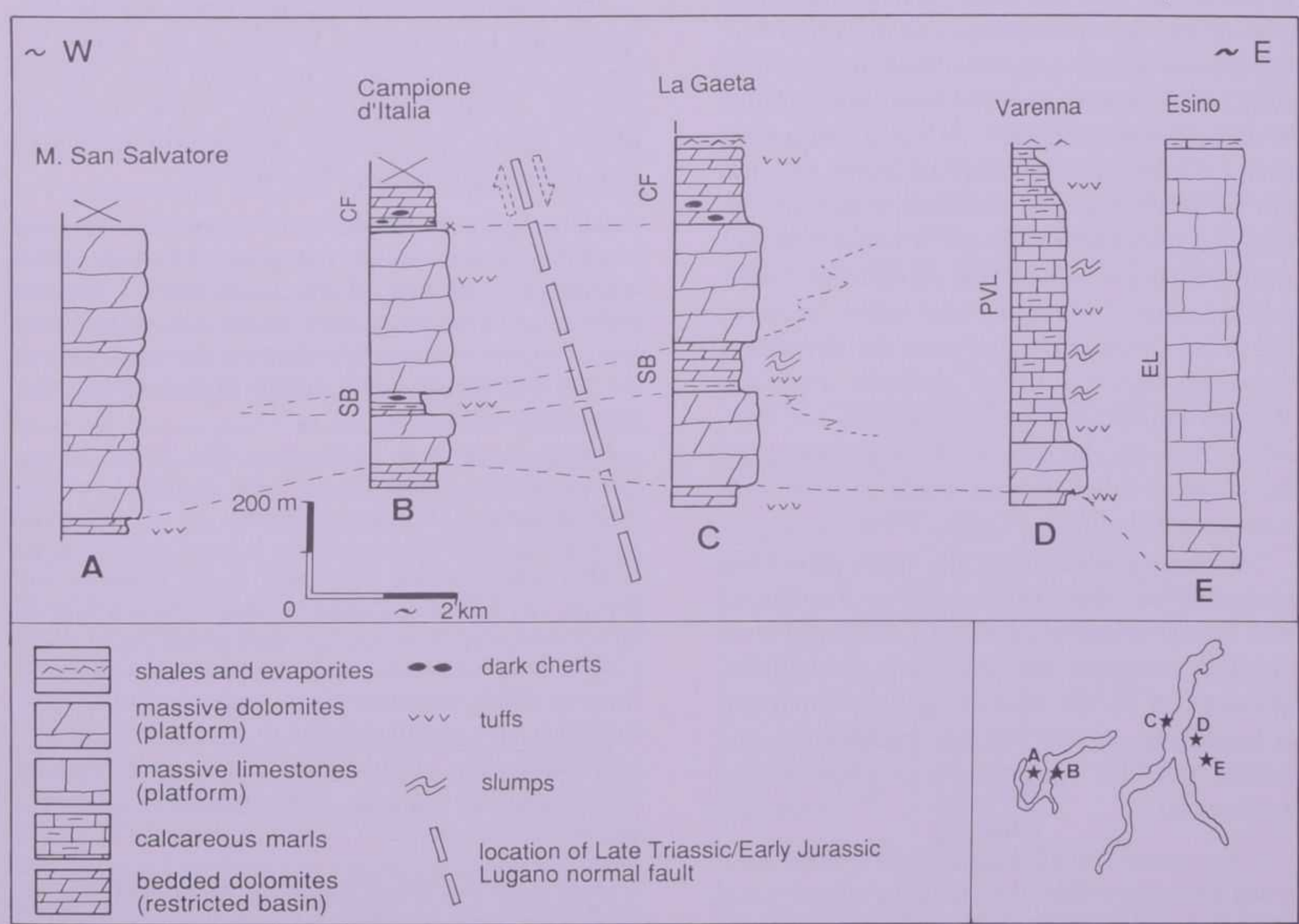

FIG. 9 - Stratigraphic columns of the Salvatore Dolomite in the Lake Como-Lake Lugano region. The La Gaeta, Varenna and Esino sections have been brought closer to the western ones in order to compensate the late Triassic to Liassic movement along the Lugano normal fault. Note similarity between the Gaeta and the Campione d'Italia sections pointing to the contiguity of the two areas. Inset in the lower right shows the *present day* position of the stratigraphic columns. Single stratigraphic sections are taken from GIANOTTI and TANNOIA (1988). SB = Besano Schists; CF = Cunardo Formation; PVL = Perledo-Varenna Limestone; EL = Esino Limestone.

succession of Monte San Giorgio (Grenzbitumenzone, S. Giorgio Dolomite and Meride Limestone). This is considered to be deposited in a shallow lagoon with restricted circulation (RIEBER, 1973). The Salvatore Dolomite of the Lugano region grades upwards into the well bedded bituminous limestones of the Cunardo Formation.

The Salvatore Dolomite and the Cunardo Formation in the area of study

The Salvatore Dolomite, as the older sediments, is found between Breglia and Lake Como and as tectonic slivers along the Monte Grona line. In the former area the dolomites form a ca. 500 m thick succession well visible along the La Gaeta section (Lake Como) (Fig. 6) (GIANOTTI and TANNOIA, 1988). The first 20-30 meters of carbonates overlying the clastics of the Bellano Formation, are well bedded, dark, peritidal dolomites. They rapidly grade into the massive part of the succession. This consists of light-coloured dolomites mainly made up by bioclastic wackestones to packstones with common dasyclad algae, gastropods and oncolites. Diagenesis is always very intense and former aragonitic skeletons are only outlined by preserved micrite rims. Stromatolitic horizons are also common. At about 200 meters from the base of the Salvatore Dolomite a ca. 100 m thick interval of well stratified, sometimes bituminous dolomites and shales is found. This horizon is characterized by the presence of resedimented material, slumping and tuffitic horizons and is tentatively correlated with the Grenzbitumenzone (Besano Beds) and the Perledo-Varenna Limestone (Fig. 9) (GIANOTTI and TANNOIA, 1988). On the base of this correlation, the age of the stratified horizon would be latest Anisian - earliest Ladinian (RIEBER, 1973).

The massive dolomites of the upper part of the succession are capped by several tens of meters of well bedded dolomites with well developed stromatolitic structures and diagenetic black cherts (outcrops on the old road along Lake Como and at coordinates 739.62/101.10) (see MALIVA and SIEVER, 1989 for a discussion of chertification mechanisms).

In thin section (Plate I, Figs. 2, 3 and 4) these interesting dolomites consist of an irregular alternation of sometimes graded peloidal layers and laminated horizons probably of algal origin deposited in a supratidal environment (SHINN, 1983). Black cherts grow across the sedimentary structures without destroying the preexisting texture. The presence of pyrite crystals points to reducing conditions during diagenesis.

Finely laminated, sometimes cross-laminated dolomicrites to dolomicrosparites follow which show no other sedimentary structures or fossils. The presence of detrital quartz and feldspars indicates incipient terrigenous contamination. This well-bedded part of the carbonatic succession is about 120 meters thick and corresponds to the Cunardo Formation (ALLASINAZ, 1968). The succession is, both as far as thicknesses and lithologies is concerned, very well comparable to that of Campione d'Italia (BERNOULLI, 1964) (Fig. 9). This is well compatible with my tectonic model according to which the paleogeographic position of the Middle Triassic succession presently found along Lake Como was originally adjacent to that of the Lugano region. These similarities are on the contrary hardly explainable with the paleogeographic picture proposed by GAETANI *et al.* (1986), GIANOTTI and TANNOIA (1987) and others according to whom the Monte San Salvatore and the Esino platforms are separated by the Monte Grona basement high.

The Salvatore Dolomite and the Cunardo Formation are found west of Breglia in the upper Val Rezzo, at Catelina and at the Denti della Vecchia (Fig. 6). All these outcrops are found along the Monte Grona line and are tectonically emplaced (see general remarks to this chapter).

In the upper Val Rezzo a few meters of stratified bituminous limestones are found which LEHNER (1952) considered to be part of the Raibl Beds. I consider them, in contrast, to be part of the Cunardo Formation. The base of these limestones is tectonic; towards the top they pass to a few meters of tectonized variegated marls (Raibl Beds).

Further to the west, at Catelina (Fig. 6), ca. 20 meters of massive dolomites underlie stratified, bituminous carbonates. The lower contact of the dolomites with Servino-like quartzarenites is stratigraphic. If the bedded carbonates are correlated to the Cunardo Formation, the Salvatore Dolomite would show a very reduced thickness. The bedded carbonates could, however, also be correlated with the Grenzbitumenzone. The top of the carbonates is not exposed.

Lithologies and thicknesses at the Denti della Vecchia are similar to the Catelina succession (LEHNER, 1952); however, the base of the dolomites is not visible.

A ca. 200 meters thick succession of Salvatore Dolomite with overlying Cunardo Formation is found in the Tesserete area (Fig. 6). The tectonic position of these outcrops with respect to the Lugano line and to the Monte Grona line is unclear. The dolomites are however likely to be part of the footwall of the Lugano normal fault.

Raibl Beds
(Carnian p.p.)

The Raibl Beds are characterized by rapid facies changes. Under the denomination Raibl Beds, all the lithologies are understood lying between the carbonates of the Cunardo Formation and the dolomites of the Dolomia Principale.

The Raibl Beds are generally very poor in fossils. The base of the formation is considered to be middle Carnian in the Lake Como region (GAETANI *et al.*, 1986). In the Lugano area it is also interpreted as middle Carnian (ASSERETO, 1973; JADOUL and ROSSI, 1982); alternatively GIANOTTI and TANNOIA (1988) proposed an Early Carnian age.

The age of the top of the Raibl Beds is not known and is conventionally attributed to the Carnian-Norian boundary.

The Ladinian carbonate platforms of most of the central-western Southern Alps died during the Carnian. This was due to the massive input of terrigenous sediments derived from the erosion of the basement highs and volcanic complexes of the "Southern mobile belt" (BRUSCA *et al.*, 1981). Carbonate sedimentation continued only in the areas less affected by terrigenous contamination: in the north and between the major clastic fans. GAETANI *et al.* (1986) recognize in the Carnian of central-western Lombardy two major clastic fans: a Val Brembana fan east of the Grigna mountains and a Lario fan in the Lake Como region (Lario being the old name for Lake Como). The well known Val Brembana fan is characterized by several hundred meters thick sandstones organized in two main cycles: the Val Sabbia Sandstone and the San Giovanni Bianco Formation. The two bodies are separated by shallow water carbonates deposited during an episode of reduced clastic input. The sandstones clasts typically consist of volcanics (GARZANTI and JADOUL, 1985; GARZANTI, 1985). The Lario fan is till now known only on the base of small outcrops at the western margin of the Grigna mountains. The sandstones of this fan differ from those of the Lario fan in having a high quartz content (GAETANI *et al.*, 1986).

The Raibl Beds of the Lugano region are made of sandstones to conglomerates, with quartzitic, volcanic and dolomitic clasts, and variegated marls (BERNOULLI, 1964). Further to the SW, in the Monte San Giorgio region, terrigenous input is limited; marls and dolomites are here typical lithologies (FRAUENFELDER, 1916). In both regions evaporites (gypsum) are present.

The Raibl Beds in the study area

The only complete succession of Raibl Beds in the study area is found between Breglia and Lake Como (Fig. 6). Outcrops are, however, quite poor so that only scattered observations can be made. A some 150-200 meters thick sequence of green to grey, sometimes red dolomicrites and marls (for instance at coordinates 738.8/100.47) overlies the dolomites of the Cunardo Formation (Fig. 10). The dolomicrites are irregularly bedded, sometimes bioturbated and show limited terrigenous contamination (quartz and muscovite). A some 10-20 meters thick quartzarenite is found (coordinates 739.07/99.88). I tentatively place it at the top of the dolomicrites.

The quartzarenite is grain-supported. Quartz is dominant among the clasts, but feldspars are also present; grains are well sorted, angular, typically around 0.1 mm large. The cement is calcitic. The succession is fairly well comparable with that found at Lierna, on the east side of Lake Como which is also characterized by abundant quartz clasts (GAETANI *et al.*, 1986). This would indicate an extension towards the west of the Lario fan larger than previously assumed. According to GAETANI *et al.* (1986), the quartz-rich sandstones should be coeval with the San Giovanni Bianco Formation of the Val Brembana fan.

A several tens of meters thick evaporitic body (Nobiallo gypsum, NEGRI and SPREAFICO, 1869; RENEVIER, 1879) is found at the top of the succession (Fig. 10). The evaporites become thinner towards the west where they are laterally substituted by a light-yellowish often brecciated microsparite containing rare quartz grains (for instance outcrops at coordinates 738.87/100.05).

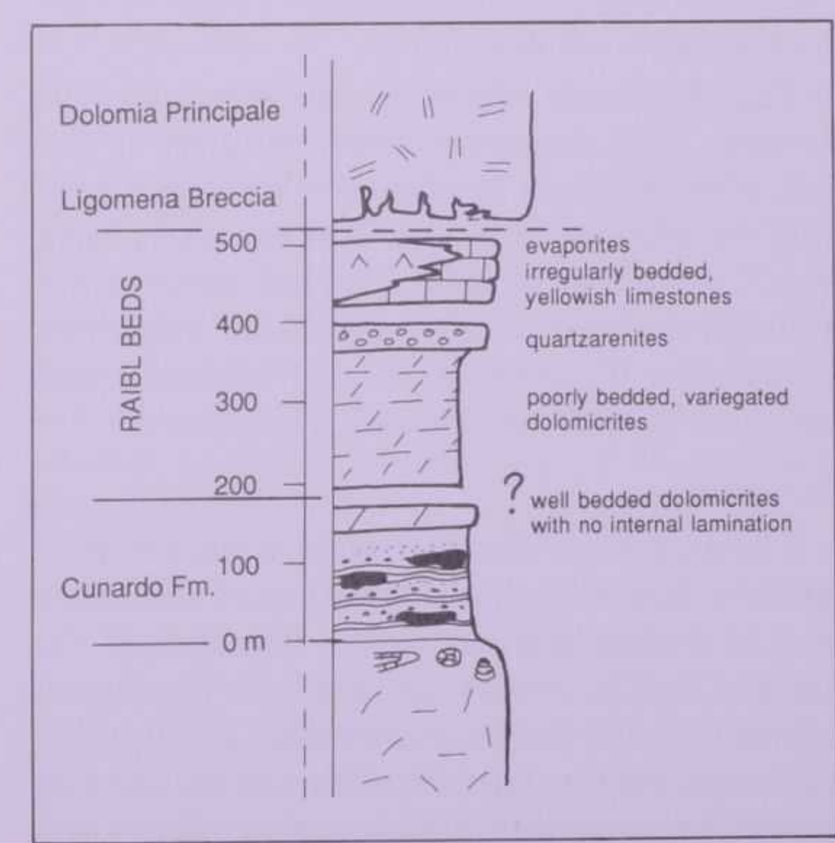

FIG. 10 - Tentative stratigraphic profile from the top of the Salvatore Dolomite to the base of the Dolomia Principale along the western shore of Lake Como.

Polygenic breccias are found along the Monte Grona line (Val Rezzo, upper Val Colla) which have been interpreted as Raibl Beds (LEHNER, 1952) (Fig. 6; Plate X, Fig. 3). The breccias are, however, of tectonic origin and formed during the extensional movements along the Monte Grona line. The breccias will be described and discussed in more detail in chapter 4.3.3.

A Carnian age has been attributed by GAETANI *et al.* (1986) and GIANOTTI and TANNOIA (1988) to an horizon of fine-grained breccias found beneath the Sasso di Cusino (Fig. 6; Plate X, Figs. 1 and 2) and to similar lithologies in Val Sanagra lying between the Manno Conglomerate and the Dolomia Principale. Similarly to the breccias of the upper Val Rezzo they have been interpreted as indicative of a transgression of the Dolomia Principale on the basement. However these horizons are clearly of tectonic origin (for a description and discussion of these breccias see the upper Val Rezzo and Piancabella profiles in chapter 4.3.3).

GIANOTTI and TANNOIA (1988) also propose a Carnian age for the sedimentary conglomerates found at the base of the Dolomia Principale east of Monte Grona (Fig. 23). The conglomerate is made up only of fairly-well rounded dolomitic clasts floating in an abundant dolomitic matrix. No paleontological evidence exists supporting the postulated Carnian age. The conglomerate is very similar to those of the Zorzino Formation: it is made up by platform-derived material and is therefore representative of deposition in a subsiding basin and not on a morphological high as proposed by GIANOTTI and TANNOIA (1988).

BISTRAM (1903) attributed the dark shales outcropping S of the Sassi della Porta (Fig. 6) to the Raibl Beds. As successively paleontologically demonstrated by SÜSSLI (1970) and on the base of unambiguous lithologic similarities the shales are now considered to be of Rhaetian age (Riva di Solto Shale).

Ligomena Breccia

The Ligomena Breccia (RENEVIER, 1879; LEHNER, 1952; GIANOTTI, 1984) is found at the base of the Dolomia Principale NW of Menaggio (Fig. 6). The breccia is some tens of meters thick and is found only along the contact between the Dolomia Principale and the Raibl evaporites (outcrops at the old quarries near Ligomena and above the gypsum quarries near Nobiallo).

Petrography

The Ligomena Breccia results from the progressive fracturing and dissolution of the lower part of the Dolomia Principale. The clasts of the breccia mainly consist of Dolomia Principale fragments; they are now, however, calcitic. The clasts are angular, typically fractured *in situ* and crossed by veins filled with a sparitic cement; no other kind of filling is observed. These composite clasts lie in a fine-grained reddish to yellowish matrix (Plate II, Fig. 1). Coarse sparitic cements are also common in different generations of veins. The cement is composed of up to mm-long crystals of columnar calcite (Plate II, Figs. 2, 3 and 6) with growth zonation and common fluid inclusions; some residual porosity is preserved. The reddish-yellowish sediment can be of two types: a) a very homogeneous silt made up of calcite grains typically 30-40 microns large with a diffuse reddish colouring (Plate II, Figs. 2 and 3) and b) a partially recrystallized colourless silt in which large dedolomitized dolomite rombohedra are preserved; the red colour is given by tiny iron oxides growing along crystallographic faces and is therefore different from type a (Plate II, Figs. 4 and 5). In neither kind of microsparite, fossils have been found. The various kinds of silt and cement do not show systematic chronological relationships: the red sediment can fill up cavities left over by incomplete cement growth (Plate II, Fig. 2) or act as substratum on which columnar calcite nucleates and grows (Plate II, Fig. 3). Dissolution episodes documented by eroded calcite crystals are often observed. On the base of these observation I assume that the growth of columnar calcite and the deposition of the silts were going on roughly at the same time. The breccia is presently mostly calcitic: the Dolomia Principale clasts and the dolomitic rombohedra have been dedolomitized.

Isotope data

Stable isotopes have been measured in order to gain some further information on the genesis of the Ligomena Breccia. Other measurements have

TABLE 2 - Isotopic data

Description of sample	$\delta^{18}O$	$\delta^{13}C$
Dolomia Principale	−5.28	2.47
Dolomia Principale	−3.71	2.68
Dolomia Principale	−6.86	2.74
Clasts	−10.36	−1.03
Clasts	−5.28	−2.29
red silt	−11.28	−3.77
red silt	−12.82	−1.78
microsparite with iron oxides	−13.40	−2.70
microsparite with iron oxides	−13.21	−2.89
coarse columnar cement	−9.00	−4.67
coarse columnar cement	−8.87	−4.79

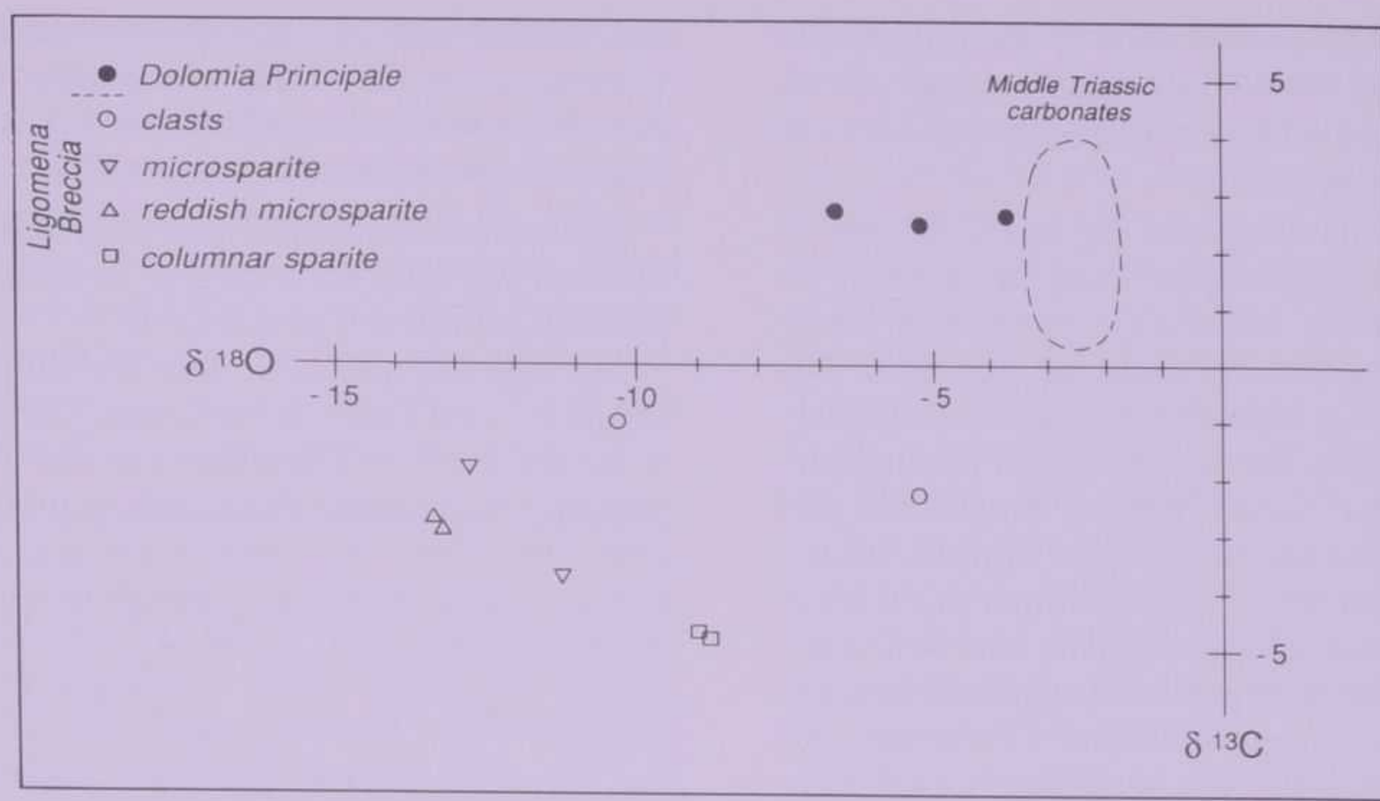

FIG. 11 - Isotope data from the Ligomena Breccia and the Dolomia Principale plotted on a $\delta^{13}C/\delta^{18}O$ scatter diagram. The dashed line outlines the field typical for Middle Triassic carbonates in the Southern Alps (after FRISIA-BRUNI *et al.*, 1989). Data are expressed relative to the PDB standard.

been carried out on Dolomia Principale samples collected tens to hundreds of meters away from the Ligomena Breccia.

The results are given in figure 11 and in table 2.

The samples were first roasted under vacuo at 400°C for 30 min and reacted with 100% H_3PO_4 at 50°C. The CO_2 gas evolved was analyzed with a VG-903 mass spectrometer. The data are expressed in the δ‰ notation relative to the international PDB standard.

The first clear feature resulting from figure 11 is that Ligomena Breccia and Dolomia Principale have different isotopic signatures. The Dolomia Principale samples fall not far away from the field of Middle Triassic marine carbonates as described by FRISIA-BRUNI *et al.* (1989) in the Lombardian Southern Alps and also by LO CICERO (1987) in the Late Triassic Panormide platform of Sicily. They are slightly shifted towards the field of deep burial diagenesis. I therefore assume that they substantially preserve their original, marine signature. The clasts, the silts and the cements of the Ligomena Breccia fall into a totally different field characterized by strongly negative $\delta^{18}O$ and $\delta^{13}C$ values. No major differences exist between the two kind of silts. The coarse columnar cement has also negative $\delta^{13}C$ values; $\delta^{18}O$ the signature is slightly less negative than that of the silt. The isotopic composition of the clasts in the breccia is slightly shifted towards the Dolomia Principale samples.

Interpretation

The isotopic data of the Ligomena Breccia are clearly incompatible with a marine origin (Fig. 11) of the red, the yellowish silts and of the sparite. The deposition of the cement and of the silt took place under meteoric water conditions. This is also compatible with the observed pervasive dedolomitization and the oxidation of ferrous iron in dolomite or ankerite. The absence of a trend connecting the various points, indicates further that the isotopic composition of the waters did not sistematically change during the deposition of the silt and the growth of the cement.

The age of formation of the Ligomena Breccia is more difficult to establish: the breccia could have formed during an emersion phase shortly after deposition or, alternatively, in relatively recent times. I favour the second hypothesis because of the following observations:

a) The isotopic composition of carbon and especially of oxygen in the calcite of silts and cement is extremely negative. This is hardly compatible with the much higher isotopic values expected for meteoric waters in the Triassic when the Southern Alps were at much lower latitudes (see discussion in FRISIA-BRUNI *et al.*, 1989; see also GOLDSTEIN, 1990).

b) The Ligomena Breccia is found only along the segment of Dolomia Principale which overlies the Raibl Beds evaporites. The breccia is not found west of the present occurrence of evaporites (Fig. 6). The presence of the evaporites apparently enhanced the formation of the Ligomena Breccia. Since the contact between Dolomia Principale and Raibl Beds is Alpine tectonic (chapter 3.2), the formation of the Ligomena Breccia should have occurred after Alpine shortening.

This hypothesis is reinforced by the observation that fracturing of the Dolomia Principale clasts, possibly related to Alpine thrusting, precedes the deposition of the red silt.

I consider therefore the Ligomena Breccia to have formed after Alpine orogeny, possibly in Quaternary times and under near-surface conditions. The base of the Dolomia Principale was fractured during Alpine thrusting enhancing circulation of fluids. The dissolution of the underlying evaporites caused further brecciation and further favoured the circulation of meteoric waters. The system can be defined as a interstratal karst (WRIGHT, 1984). The isotopic data are well compatible with the interpretation proposed. Isotopic compositions of oxygen similar to my data are not very common in calcite; however they have been found in present day caves in different parts of the world at latitudes comparable with the present day one of the Ligomena Breccia (*e.g.* SCHWARCZ, 1986).

My hypothesis clearly differs from that presented by GIANOTTI (1984). This author, mainly on the base of some supposed sedimentological similarities, draws a genetic parallelism between the Ligomena Breccia and the Macchia Vecchia Breccia of Arzo (WIEDENMAYER, 1963). This parallelism however cannot be accepted not because:

1) The Macchia Vecchia Breccia lies at the top of the Dolomia Principale and not at the base as the Ligomena Breccia.

2) Isotope signatures in clasts and matrix of Macchia Vecchia Breccia (WINTERER *et al.*, in press) fall well within the marine field of figure 11 and are therefore not comparable with the values derived from the Ligomena Breccia.

The Ligomena Breccia has therefore a meteoric dedolomitization origin and cannot be related to rifting as proposed by GIANOTTI (1984).

Dolomia Principale and Zorzino Formation (Norian)

With the cessation of terrigenous input following a transgression at the beginning of the Norian, the carbonate platforms could spread all over the region. It is during the Norian that the first tectonic movements related to rifting are recorded in the Lake Lugano-Lake Como region. Strongly subsiding domains are identified contrasting with domains of limited subsidence. Some of the normal faults had a morphological expression, others didn't. In the first case basins formed in which platform-derived material was redeposited interfingering with autochthonous carbonate sedimentation. In the second case, where no morphological differentiation was caused, the same facies are found on both sides of the fault. I will use the denomination Dolomia Principale for the carbonate platform sediments and the name Zorzino Formation for all the sediments deposited on the slopes and in the basins.

The term Dolomia Principale (equivalent to Hauptdolomit) has long been used to indicate the Norian platform sediments (LEPSIUS, 1878).

The denomination Zorzino Formation has been adopted for the first time by CASATI (1964) to indicate a succession of well bedded bituminous limestones overlying the Dolomia Principale in the Lake Iseo region. The same formational name has been used to indicate similar deposits in the Bergamasc Alps (M. Aralalta, JADOUL, 1985). I extend the formational name Zorzino Formation to the lithologies of similar age and facies found between Lake Como and Lake Lugano. The name Zorzino Formation replaces older denominations like "Plattenkalk" (BISTRAM, 1903) and Tn_2 (Sheet Chiavenna of the Geologic Map of Italy 1:100.000) and is equivalent to the "Aralalta Group" (JADOUL, 1985).

Biostratigraphic dating of the Dolomia Principale and of the Zorzino Formation is poor and limited to a few Norian macrofossils found in the Dolomia Principale. The two formations are both overlain by the Riva di Solto Shale so that they must be considered as at least partially coeval. The ages of bottom and top of the Dolomia Principale/Zorzino Formation complex are not exactly known. The general agreement is to attribute the base of the Dolomia Principale to the Carnian-Norian boundary and the top of the carbonate succession (Dolomia Principale or Zorzino Formation) to the Norian-Rhaetian boundary (ASSERETO and CASATI, 1965).

Two major domains with different subsidence rates can be recognized in the Lake Como-Lake Lugano region: the Lugano or Arbostora swell to the west and the Generoso basin to the east. These domains will maintain their subsidence pattern for the rest of the Triassic and for the Early Jurassic. Carbonate platform sedimentation goes on in the Lugano area for the whole of the Norian; subsidence is limited and the Dolomia Principale is only ca. 400 meters thick (FRAUENFELDER, 1916). The top of the Dolomia Principale along the eastern margin of the Arbostora high is affected by intensive Early Liassic tectonics (section 4.2.1) (WIEDENMAYER, 1963).

The Generoso basin is characterized by strong subsidence. The abrupt thickness contrast between Lugano swell and Generoso basin indicates that the boundary between the two domains was a normal fault, the Lugano fault, which therefore was active already during the Norian. The Lugano normal fault had at this time no morphological expression and facies on both sides of the fault are comparable. In the strongly subsiding Generoso basin, hangingwall of the Lugano normal

fault, small, tectonically controlled intraplatform basins developed during the Norian.

Sediments of Norian age east of Lake Como are found only further to the south, in the Corni di Canzo-Lecco region (western part of the Albenza plateau). Here, platform carbonates are found with thicknesses of 1200 meters (LICHTENSTEIGER, 1986). Thicknesses are roughly comparable with those of the eastern Generoso basin. These two domains were thus not separated by major faults.

Dolomia Principale and Zorzino Formation in the area of study

Sediments of the Dolomia Principale and Zorzino Formation are the most widespread lithologies in the study area. They are found along a some kilometers wide strip from lake Como to the area of Tesserete. The Norian sediments of the study area have been deposited in the strongly subsiding Generoso basin. Extensional faulting was going on not only along the Lugano fault but also within the Generoso basin itself causing the development of local morphological basins in which thick aprons of redeposited sediments were accumulated.

Dolomia Principale (platform deposits)

Platform sediments are very widespread in the study area. They are found in the west building the beautiful mountains between the Italian-Swiss border to the north and the lower Valsolda to the south. Platform sediments, although less extensive, are found also in the east, in the Monte Grona region, in the area west of Menaggio and, south of it, at the Sasso di S. Martino (Fig. 6).

In the recently investigated Sasso di S. Martino area (CIRILLI and TANNOIA, 1985; LUALDI and TANNOIA, 1985; GAETANI *et al.*, 1986) two main facies associations are recognized: a platform margin and a back-margin domain (Fig. 12).

The platform margin mainly consists of lenticular organic build-ups (bafflestone and bindstone) and of peloidal, bioclastic grainstones to packstones usually stabilized by incrusting algae and /or by very early diagenesis.

Behind the margin (internal platform of the mentioned authors) fairly well stratified carbonates are organized in shallowing upward cycles. The cycles typically have a lower part of subtidal bioclastic wackestone to packstone followed by a stromatolitic horizon with fenestrae and capped by a supratidal interval characterized by tepee structures (Fig. 12). Sedimentation is assumed to have taken place in a protected area, oscillating from shallow subtidal to supratidal. In the uppermost part of the succession, the internal platform progrades over the platform margin.

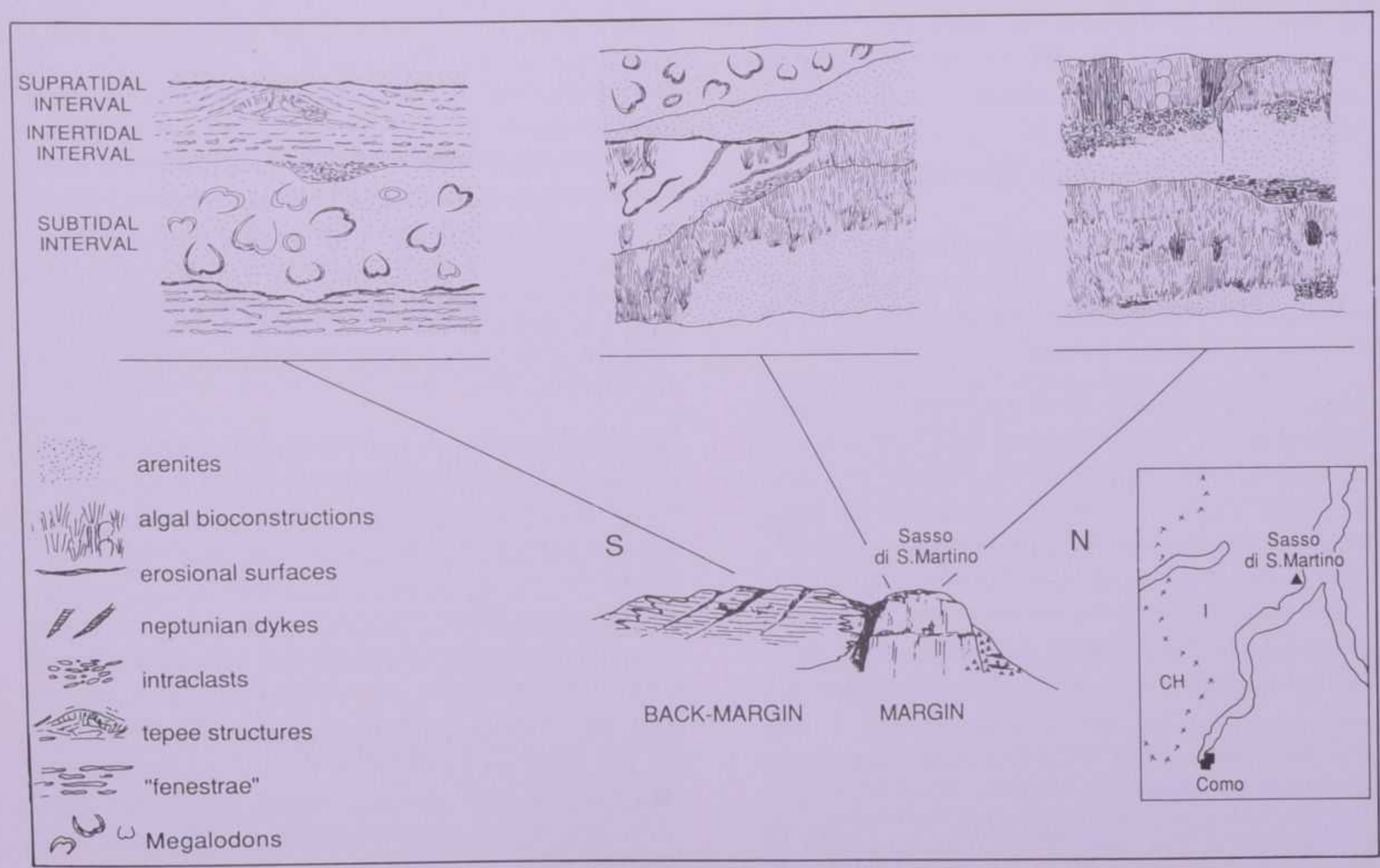

FIG. 12 - Facies relationships in the Dolomia Principale of the Sasso di S. Martino area. Insets in the upper part of the figure are outcrop sketches. From GAETANI *et al.* (1986) modified.

Mostly because of lack of data, it is difficult to compare the Sasso di San Martino to the vast platform complexes found in the rest of the study area. The most widespread facies to the west, are the thick-bedded peloidal, bioclastic grainstones to packstones interpreted by CIRILLI and TANNOIA (1985) as part of an internal platform. Differently from what suggested by CIRILLI and TANNOIA (1985) I found no direct relationship between the position of the marginal facies and that of syn-sedimentary extensional faults ([1]).

The thickness of the Dolomia Principale is, not surprisingly, highly variable. It reaches a maximum of about 1700 m in the west, while along the Val Sanagra it has a thickness of about 1200 m. The depocenter of the Generoso basin was therefore clearly located close to the Lugano fault.

Zorzino Formation (slope and basin deposits)

Slope and basinal deposits, here grouped together in the Zorzino Formation, are found mainly in the central part of the area of study (Fig. 23). The thickness of the sediments is highly variable being controlled by the local paleotectonic setting. Maximal values are ca. 1500 meters.

The overwhelming majority of the sediments is, independently of facies, dolomitic. Calcareous lithologies are sometimes found: they typically consist of thin beds of mud-supported conglomerates and of turbidites and do not include chaotic megabreccias. Most of the calcareous lithologies are found near the top of the Zorzino Formation or right beneath angular unconformities (Fig. 23).

Four groups of facies have been distinguished:
a) chaotic megabreccias;
b) stratified breccias and conglomerates;
c) doloarenites;
d) stratified, homogeneous dolomicrites (dolomicrosparites), and finely laminated dolomites.

Chaotic megabreccias. They typically form huge bodies located near the Norian extensional faults (Fig. 23). Good outcrops are common in the western part of the study area, for instance west of the upper Val Rezzo. The breccias (Plate III, Fig. 1) are matrix- to (rarely) clast-supported. Clasts range from cm- to m- sized and are mostly derived from the adjacent platform; some dark basinal clasts are also found. No bedding is observed. Deposition took place on a peri-platform talus (MC ILREATH and JAMES, 1984). Catastrophic events of breccia deposition at the foot of tectonic scarps were followed by more regular sedimentation of fine grained packstone to grainstone (Plate III, Fig. 2). Moving away from the fault, at distances of a few hundred meters, the breccias remain chaotic but tend to become better stratified (Fig. 13).

FIG. 13 - Tens of meters thick beds of megabreccias laterally pass into well bedded, mud-supported conglomerates moving away from the Norian M. Mugetto fault (to the W in figure). Northern side of the Cime di Bronzone. Sketch after photo.

Chaotic breccias also form bodies of lenticular shape, few hundreds of meters wide and few tens of meters thick intercalated in the better stratified part of the succession. While the top of the bodies is roughly planar, sediments at the base are slumped and chaotically deformed. Good examples are found in the southern flank of the Monte Pidaggia where they form bodies resistant to weathering (for instance E of Carlazzo at coordinates 733.47/101.45). I interpret these bodies as channels transporting sediments from the platform to the basin cut parallel to the transport direction; dimensions, relative distance and geometry of the channels are comparable to recent and fossil examples discussed by MULLINS and COOK (1986) for instance from the Devonian Prongs Creek Formation, Yukon Territory, Canada.

Stratified breccias. I include in this group breccias similar to those described above but organized in well defined strata. Good outcrops are found for instance along Val Sanagra and at the Cime di Bronzone.

The stratified breccias consist of mud-supported breccias to conglomerates with chaotically disposed clasts of variable dimensions but typically ranging from mm to cm (Plate III, Fig. 3). The breccias form beds with thicknesses typically varying from dm to meters, often associated with bundles of thinner-bedded turbidites. Not much erosion is visible at the base of the breccia beds at outcrop scale (Plate III, Figs. 4 and 5). The stratified breccias perfectly match with the

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[1] As a matter of fact this relationship is questionable also in the Sasso di S. Martino area itself. LUALDI and TANNOIA (1985), incorrectly considered the syn-sedimentary fault to strike WNW-ESE. This fault is however Alpine; the syn-sedimentary fault strikes NE-SW and therefore oblique to the E-W striking paleogeographic zones (Fig. 6).

description of mud-supported conglomerates, facies F of MULLINS and COOK (1986). A debris-flow mechanism is envisaged for the deposition of these sediments (see CREVELLO and SCHLAGER, 1980 for an actualistic example).

Turbiditic doloarenites to dolosiltites. This represents the most common sediment type present in the study area. Good outcrops are frequent in Val Sanagra, along the road from Carlazzo to Piano Porlezza, in Val Rezzo etc.

A variety of lithologies is here considered which have in common the grain size (<cm-sized) and typical sedimentary structures. Well developed bedding is ubiquitous. Included in this group are:

- cm- to dm-thick turbidites. They are generally graded (Plate IV, Fig. 1) and show often convolute laminations. Clasts are mm- to cm-sized and are both platform- and basin-derived. The base of the beds can be slightly erosive; amalgamation is sometimes observed.

- Thin-bedded turbidites: few cm-thick mud turbidites which show grading and convolute laminations. Complete Bouma intervals are sometimes observed. Clasts are subordinate to matrix and generally consist of mm- or smaller-sized, dark, reworked basinal sediments; platform-derived clasts are rare. Dark flakes of basinal sediments are sometimes observed at the top of the beds. No current or drag marks have been found. These microturbidites have often been deformed by slumping and syn-sedimentary decollements (Plate IV, Figs. 2 and 3) indicating an unstable depositional setting on a slope.

(Dolo)micrites. These are homogeneous beds of dolomicrite often recrystallized to (dolo)microsparite. They are thought to represent the basinal autochthonous sedimentation and/or fine-grained mud turbidites. Sediments are typically strongly bituminous.

The (dolo)micrites can be organized in cm- to dm-thick beds with no internal structure. They alternatively form mm-thick alternations of laminae of light-coloured dolomicrosparite and darker organic matter rich layers (Plate IV, Fig. 4). Lamination is irregularly disturbed by undulations of diagenetic origin. Burrows are conspicuously lacking.

The association of thin-layered dolomite and organic matter-rich layers could indicate deep water dolomitization in environments with low sulfate concentrations (reducing conditions) (KELTS and MCKENZIE, 1984). Conditions of this kind require a, at least temporarily, effective stratification of the water column. Such a paleooceanographic setting is also suggested by the presence of corrensite, a mineral indicating high salinities (DUNOYER DE SEGONZAC and BERNOULLI, 1976). The widespread occurrence of albite in the Zorzino Formation also suggests hypersaline conditions.

Depositional geometry. The depositional geometry of the Zorzino Formation is directly controlled by the syn-sedimentary normal faults (Fig. 23). As a general rule, a very coarse peri-platform talus of disorganized breccias was deposited adjacent to the fault scarps. Moving away from the faults, breccias and conglomerates become finer-grained and organized in roughly planar beds (Fig. 13) with channels along which the coarse, platform-derived sediments were transported into the deeper parts of the basin. Large amounts of mud-supported conglomerates and, quantitatively more important, turbiditic arenites were transported into the basins. Organic matter-rich, generally well laminated carbonates were deposited in the deepest part of the basins.

The absence of major angular unconformities in the basins shows that redeposition and tectonic movements along the normal faults, went on, on the whole, at comparable rates. A few local angular unconformities within the succession are found (for instance at Monti di Gnin in the middle Val Rezzo) (Fig. 23). The redeposited sediments, as well visible in Val Sanagra, form wedge-shaped aprons with thick proximal sequences of coarse-grained sediments and thin distal sequences of thinner turbidites and basinal carbonates.

Riva di Solto Shale, Zu Limestone, Conchodon Dolomite (Rhaetian)

The large scale tectonic pattern described for the Norian persists for most of the Rhaetian: in the west, the Lugano swell still shows very limited subsidence and sedimentation; in the central area, the Generoso basin undergoes very strong subsidence with high sedimentation rates, while in the east (Corni di Canzo-Lecco), subsidence has intermediate values. The transition from the Lugano swell to the Generoso basin is fault-controlled. The transition from the Generoso basin to the Corni di Canzo seems to be more continuous.

The Norian/Rhaetian boundary represents a major sedimentological change related to the input of large masses of fine-grained terrigenous sediments (chapter 2.3.1). These were obviously trapped mainly in the most subsiding areas, but also in the less subsiding regions marls and shales are found.

The Dolomia Principale of the Lugano swell is unconformably overlain by Rhaetian shallow

water limestones, dolomites and marls. The laterally and vertically very irregular Rhaetian succession (Tremona Series of KÄLIN and TRÜMPY, 1977) ranges in thickness between 0 and 70 meters and is therefore much thinner than the coeval succession in the Generoso basin.

The lower part of the Tremona Series is characterized by reworked Dolomia Principale clasts. The main part of the Tremona Series consists of shales, marls, oolitic limestones and dolomites considered to be coeval with the Zu Limestone (KÄLIN and TRÜMPY, 1977). A stratigraphic gap is found at the top. Stratigraphic gaps and angular unconformities are also found within the Rhaetian succession.

A somewhat similar succession is found in small tectonized outcrops immediately east of the Lugano line. This implies that the location at the surface of the Rhaetian Lugano normal fault was slightly eastward of the Early Liassic one.

Successions which can be referred to this domain are found in the Torrente Cassone valley NE of Lugano (Fig. 6) (LEHNER, 1952) and at Monte S. Agata along the Lugano line (BERNOULLI, 1964). The Torrente Cassone succession consists of dolomites, oolitic and bioclastic limestones, some dark marls, and breccias with Rhaetian components (glauconite in the matrix) at the top. The Monte S. Agata succession is quite similar (BERNOULLI, 1964).

The Rhaetian formations of the Generoso basin are extremely thick (up to 1300 m) and show the three-fold subdivision in Riva di Solto Shale, Zu Limestone and Conchodon Dolomite typical for the Lombardian basin (chapter 2.3.1) (GNACCOLINI, 1965a). The Riva di Solto Shale at the base mainly consists of organic-rich shales with subordinate marls and thin micritic intercalations. Calcareous horizons become more common and thicker upwards until the shales are only subordinate (Zu Limestone). While the facies remain substantially constant across the whole Generoso basin, thicknesses of the Zu Limestone and particularly of the Riva di Solto Shale are strongly variable and directly controlled by active normal faulting.

The Zu Limestone is overlain by the shallow water carbonates of the Conchodon Dolomite.

The time-stratigraphy of the Rhaetian formations is poorly known. The usually accepted convention is to identify the base of the Riva di Solto Shale with the Norian-Rhaetian boundary and the top with the end of the Rhaetian (GAETANI *et al.*, 1986).

Rhaetian sediments in the area of study

The Rhaetian sediments in the area of study are part of the Generoso basin.

The *Riva di Solto Shale* consists of dark shales and marls which can be very finely laminated. Some thin intercalations of limestones are found at different levels and tend to become more common upwards. The Riva di Solto Shale has been deposited in a moderately deep, poorly oxygenated environment. It occurs along a roughly E-W oriented stripe south of Val Menaggio and in the lower Valsolda. The thickness of the Riva di Solto Shale is directly controlled by syn-sedimentary faults (Fig. 6).

The Riva di Solto Shale is missing on the Sasso di S. Martino (Lake Como). A few hundreds of meters to the NW, after crossing a NE-SW striking Mesozoic normal fault, it reaches a thickness of about 400 m. This thickness remains quite constant towards the west in the well known Bene profile (ESCHER v.d. LINTH, 1853; GNACCOLINI, 1965a) and in Porlezza (445 m.). The Riva di Solto Shale is again missing on the Dolomia Principale of I Pizzoni west of Porlezza (Fig. 6). I interpret this as due to a syn-sedimentary normal fault now buried beneath the Quaternary sediments of the Val Menaggio. Further to the west the shales are found only in small NE-SW elongated grabens around Dasio (Valsolda). In the uppermost Valsolda and in the Denti della Vecchia region the Riva di Solto Shale is missing but the contact between Dolomia Principale and Zu Limestone is tectonic.

The Riva di Solto Shale is also found along the major, roughly E-W trending Alpine thrusts (Fig. 6). Along the northern thrust, the Riva di Solto Shale is found only S of the Sassi della Porta. Along the southern thrust the Riva di Solto Shale forms numerous outcrops from Val Rezzo in the east to Val di Duslin in the west.

The *Zu Limestone* typically consists of well-defined, grey-weathering and internally dark grey to black, dm- to m-thick beds. The limestone consists of homogeneous micrite to calcisiltite, oolitic grainstone and bioclastic packstone to grainstone. At different levels, meters-thick beds with corals in living position are found (GNACCOLINI, 1965a).

The Zu Limestone and the Riva di Solto Shale have been deposited on gently sloping muddy ramps between peritidal highs and basinal lows. Water depth was generally below wave-base but still reduced enough to allow reworking of sediment by storms (Plate IV, Fig. 5) and coral growth. The shales represent the background sedimentation which was interrupted by cyclic introduction of platform-derived calcareous mud, oolites, bioclasts etc. (MASETTI *et al.*, 1989) (chapter 2.3.1).

This interpretation differs from the generally accepted previous one (GNACCOLINI, 1965a) according to which the shales would be lagoon deposits and the limestone would represent ingressions of better oxygenated waters. However, environmental oscillations of

this regularity are not very realistic. Interpreting the limestones as allochthonous, it is also easy to explain the presence of platform-derived sediments in a succession mainly deposited in an anoxic environment.

The Zu Limestone is found in the study area only along a stripe from Lake Como in the east to the upper Valsolda in the west where it is involved in Alpine thrusting. The thickness of the Zu Limestone in the Generoso basin is variable, but on the whole more constant than that of the Riva di Solto Shale. This points to the diminishing activity of the syn-sedimentary faults within the Generoso basin during the Rhaetian.

The Zu Limestone is ca. 600 m thick west of the Sasso di S. Martino; this estimate is only indicative because of poor exposures and possible Alpine thrusting; it is however more realistic than the 340 m reported by GNACCOLINI (1965a). The thickness remains constant till the Val Menaggio (620 m according to GNACCOLINI, 1965a in the Bene profile). Further to the west, at I Pizzoni (Fig. 6), it drops to ca. 300 m and then reaches its maximal thickness in Valsolda: precise estimates are here difficult but we consider 700 m as a minimum. The Rhaetian depocenter therefore roughly coincides with the Norian one.

The *Conchodon Dolomite* (for the ethymological problems connected with this denomination see GNACCOLINI, 1965b; KÄLIN and TRÜMPY, 1977) is an horizon of mostly massive limestone to dolostone capping the Rhaetian sequence. The Conchodon Dolomite is generally made up of oolitic grainstones interbedded with bioclastic wackestones to packstones. The Conchodon Dolomite documents a re-establishment of shallow water conditions across the whole area.

The thickness of the Conchodon Dolomite varies regularly from the 80-85 m in the region east of Lake Lugano (GNACCOLINI, 1965a) to the about 150 m in the Valsolda. This increase, together with facies differences between the upper Rhaetian of the Generoso basin and that of the Lugano swell, indicates that activity along the Lugano normal fault was going on also during the latest Rhaetian. In contrast, the continuity of the Conchodon Dolomite horizon east of the Lugano normal fault, points to an extinction of the normal faults within the basin.

Moltrasio Limestone, Broccatello Limestone
(Early Liassic)

The already existing subsidence differences among the various domains (Lugano swell, Generoso basin, Corni di Canzo/Albenza plateau), became even more pronounced in the Early Liassic: while on the Lugano swell the Lower Liassic is less than 200 meters thick, the thickness of the coeval sediments in the Generoso basin is estimated at about 4000 m (BERNOULLI, 1964). A major sedimentological differentiation took place with the shallow water carbonates of the Lugano swell contrasting with the siliceous limestones deposited below the photic zone in the Generoso basin. Only with the Pliensbachian, also the Lugano swell was drowned (WIEDENMAYER, 1963; BERNOULLI, 1964; GAETANI, 1975; KÄLIN and TRÜMPY, 1977; BERNOULLI, 1980).

Early Liassic sediments of the Lugano swell consist of massive, yellowish to reddish bioclastic limestones (Broccatello Limestone) deposited under deeper neritic, perhaps subphotic conditions (BERNOULLI and GARRISON, in preparation). Thicknesses are variable because of active syn-sedimentary tectonics. Latest at the Sinemurian-Pliensbachian boundary the Lugano high was foundered beneath the photic zone and a few meters thick horizon of condensed pelagic limestones with cephalopodes (Besazio Limestone) was deposited (WIEDENMAYER, 1963). Persisting syn-sedimentary tectonic activity caused a complex pattern of veins, neptunian dykes, etc. The dykes are filled with polyphase breccias containing Broccatello and Besazio Limestones and clasts of Dolomia Principale and Rhaetian calcarenites (Macchia Vecchia) (WIEDENMAYER, 1963; BERNOULLI, 1980).

Blocks previously lying west of the Rhaetian Lugano normal fault (T. Cassone, S. Agata) were progressively incorporated into the basin: the basinal siliceous limestone of the Moltrasio Formation overlies the Broccatello Limestone and the Rhaetian of the Lugano swell. Drowning occurred during the middle/late Sinemurian. An east to west trend of younging ages for the shallow water /deep water transition is recorded (BERNOULLI, 1964).

The Early and Middle Liassic succession of the Generoso basin consists of the siliceous limestones of the Moltrasio Formation. The Moltrasio Formation is up to 4000 m thick (BERNOULLI, 1964) and, on the whole, very monotonous with no large-scale unconformities or lateral variations. Truncation surfaces, slumps, amalgamated beds and channels are, on the contrary, very common at the outcrop scale indicating an unstable depositional setting (BERNOULLI, 1964).

The typical lithology consists of dark-grey, grey-blue, well-bedded siliceous limestones. Petrographically they consist of a micrite very rich in sponge spiculae. SiO_2 is abundant both diffusely dispersed in the groundmass and forming the typical nodules and bands of diagenetic replacement cherts growing across sedimentary structures. Fine-grained conglomerates to arenites

with turbiditic features (commonly with T_a-T_c intervals) are frequent. Clasts are both skeletal and lithoclasts. Meter-sized blocks of swell-derived Rhaetian and Norian sediments are found close to the Lugano line (LEHNER, 1952; BERNOULLI, 1964). No crystalline rocks are observed among the clasts.

The base of the Moltrasio Formation is Hettangian (BISTRAM, 1903), the top Early Domerian (WIEDENMAYER, 1980).

The thickness of the Moltrasio Limestone decreases form the Generoso basin eastward. In the western part of the Albenza plateau the Moltrasio Limestone is ca. 1000 m thick (GAETANI, 1975). Locally, as in the Corni di Canzo area (Lecco), the Moltrasio Limestone is even totally missing (LICHTENSTEIGER, 1986).

The Moltrasio Limestone in the study area

Sediments of the Moltrasio Formation in the area of study outcrop south of Val Menaggio. Towards the west they cross Lake Lugano and form the mountains NE of Lugano. Only the lower part of the thick succession of Moltrasio Limestone forming the Generoso basin outcrops in the study area. The transition from the Conchodon Dolomite to the Moltrasio Formation in the study area is continuous.

The contact between the two formations can be observed along the path from the Monti di Nava to Monte Tremezzo (coordinates 737.60/95.85) (GNACCOLINI, 1965a). The Conchodon Dolomite, here calcareous, becomes gradually stratified and is overlain by a some 4 meters thick nodular limestone with very abundant bivalves; some 30 meters of stratified limestones with brachiopodes and gastropodes follow which eventually grade in the typical grey limestones of the Moltrasio Formation. No angular unconformity is observed. A very similar profile is found along the Bene profile (GNACCOLINI, 1965a). The same transition is exposed in the Valsolda along the pathway from Castello Valsolda to Alpe Bolgia. Both Conchodon Dolomite and the base of the Moltrasio Formation are here dolomitized.

3 ALPINE TECTONICS

3.1 ALPINE TECTONICS IN THE CENTRAL SOUTHERN ALPS

It is not more than a decade that the importance of Alpine shortening in the Southern Alps has been fully appreciated. Indeed, many major thrust surfaces had been recognized already at the end of the 19th century and at the beginning of the 20th: the thrusts in the Grigna region (PHILIPPI, 1987), the Orobic thrust (TRÜMPY, 1930), the Monte Generoso thrust (FRAUENFELDER, 1916) etc. However, these features were not integrated into a tectonic kinematic framework. The first authors who attempted a regional interpretation were those of the Dutch school (DE SITTER and DE SITTER-KOOMANS, 1949; DE JONG, 1967 and references therein). Their interpretation was based on gravity-driven displacements and a major implication of the basement was excluded. The gravitative interpretation remained dominant until the sixties.

During the late 70' and early 80' the first attempts were made to demostrate the relevance of crustal shortening (*e.g.* GAETANI and JADOUL, 1979; CASTELLARIN, 1979) and to give a general structural frame for the Southern Alps. A major contribution in this direction was given by the release of seismic profiles beneath the Po Plain (PIERI and GROPPI, 1981). The first attempt to estimate quantitatively the amount of shortening and give a coherent kinematic picture throughout the central Lombardian Alps is given by LAUBSCHER (1985).

The Southern Alps west of the Giudicarie line are now generally accepted to be a mainly south-verging fold- and thrust-belt (DOGLIONI and BOSELLINI, 1987). If this is the general, large-scale pattern, structures in detail show important along-strike variations. This is clearly due to the influence of mainly N-S trending structural lineaments inherited from the Mesozoic extension. Particularly relevant in this respect is the influence of the Mesozoic N-S striking normal faults which acted during compression as "transfer zones" separating domains with different geometries of shortening (LAUBSCHER, 1985) (Fig. 14). The influence of the Mesozoic configuration on Alpine tectonics is of central relevance at all scales and will be seen again and again in this work.

The problem of the age of deformation in the Southern Alps west of the Giudicarie line, particularly of its first stages, is still badly constrained. The only positive observations we have are:

- folds and thrusts are found in the Adamello region which are cut by the 40-30 Ma old Adamello pluton (DEL MORO *et al.*, 1983; BRACK, 1981) and must therefore be older;

- further to the south, the Chiasso-Stabio back-thrust bringing the Chiasso Formation on top of the Moltrasio Limestone is likely to be of Burdigalian or Tortonian age (BERNOULLI *et al.*, 1989) (Fig. 14);

- thrusts presently buried beneath the Po Plain sediments are of Tortonian age (PIERI and GROPPI, 1981).

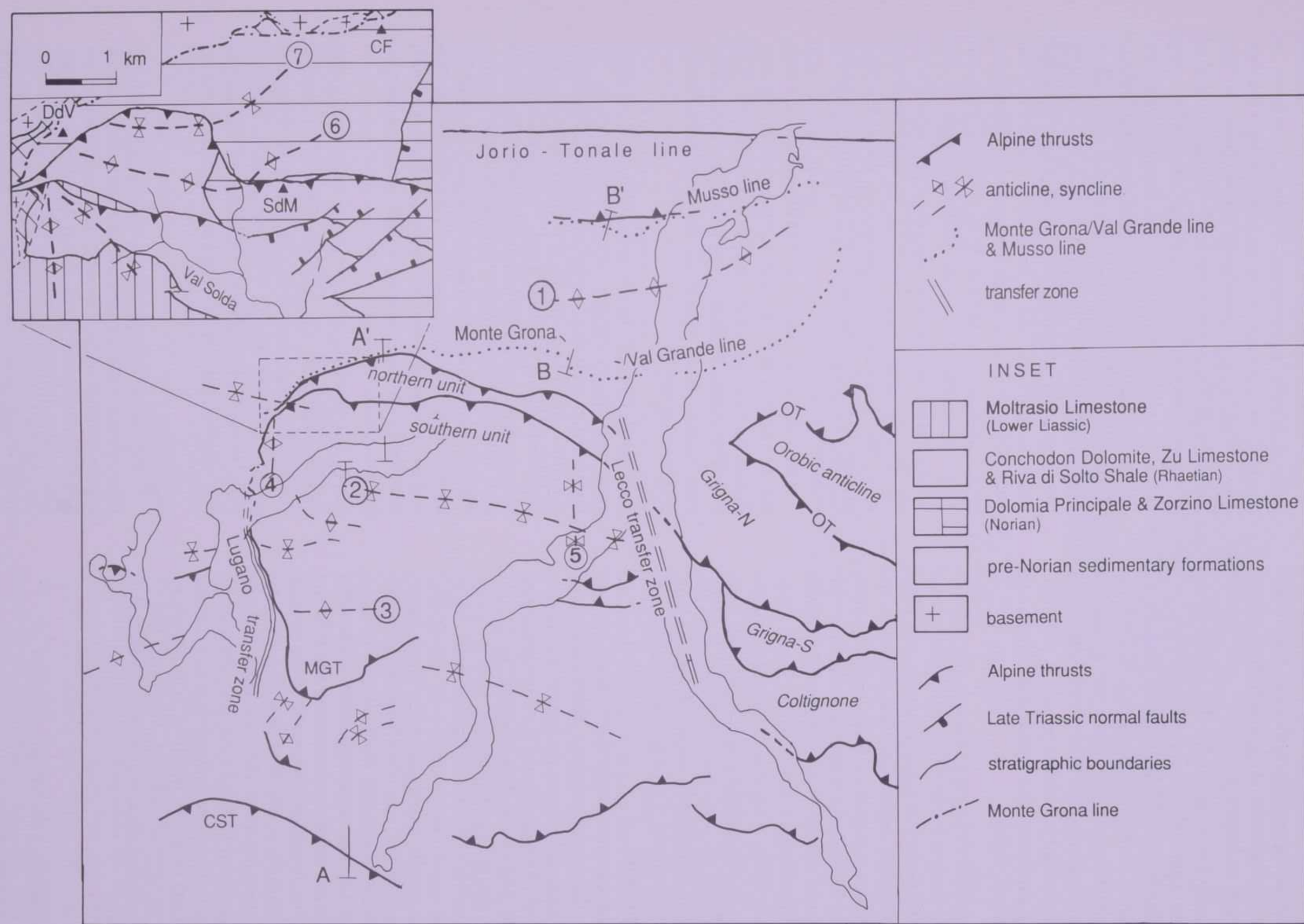

FIG. 14 - Map of Alpine tectonic elements in the Lake Como-Lake Lugano area. Inset shows in more detail the complex tectonic structures of the upper Valsolda. Modified after: BERNOULLI, 1964; LAUBSCHER, 1985; GIANOTTI and MONTRASIO, 1981; BIGIOGGERO *et al.*, 1981; GAETANI and GIANOTTI, 1981; FUMASOLI, 1974; CORNELIUS and FURLANI-CORNELIUS, 1930; BERNOULLI *et al.*, 1989. Capital letters indicate the trace of the profile shown in figure 21. Circled numbers refer to major folds mentioned in the text and are also indicated on the profile in figure 21. OT = Orobic thrust; MGT = Monte Generoso thrust; CST = Chiasso-Stabio thrust. Inset: CF = Cima Fiorina; DdV = Denti della Vecchia; SdM = Sass di Mont.

While a general younging of the structures towards the south is indeed observed, it is often impossible to date the single structures given the existence of out-of-sequence thrusts, reactivated thrusts etc. Along-strike structural variations are moreover very widespread.

In this study I concentrate on the Alpine structures of the Lake Como-Lake Lugano region. I will first discuss them in the regional frame and then present a profile from the Insubric line to the Po Plain. The profile runs roughly parallel to the trace of the reflection seismic profiles which have been recorded in 1988 and which are presently being processed (for preliminary result see BERNOULLI *et al.*, 1990).

The study area is located between the tectonically poorly known Lugano-Varese region in the west and the Grigna mountains in the east (Fig. 14). In the Lugano region only few structures related to Alpine shortening have been till now recognized. Particularly important are the M. Tamaro thrust in the north (REINHARD, 1964) and the Chiasso-Stabio thrust in the south (BERNOULLI *et al.*, 1989). A further thrust, of unknown importance, is suggested by the Raibl Beds outcropping below the Salvatore Dolomite E of M. San Salvatore (see Sheet Lugano of the Geologic Atlas of Switzerland). However, no complete reconstruction exists.

The Grigna mountains in the east, are characterized by three south-vergent thrust sheets of Middle Triassic carbonates: decollement horizons are the Early Anisian Carniola di Bovegno and the Carnian Raibl Beds. The basement is also involved in shortening. LAUBSCHER (1985) envisaged the existence of three main basement thrust sheets, the upper two separated by the Orobic thrust. The middle basement thrust sheet forms the Orobic anticline and disappears beneath the surface west of the triple junction east of Lake Como (Fig. 14); the upper basement sheet with its sedimentary cover is continuous towards the west with the crystalline basement of the studied area. Further to the south (S of Nobiallo) the geologic continuity between the two sides of Lake Como is precluded by the Lecco line which acted as a transfer zone separating the south-vergent Grigna thrusts from the north-vergent thrusts of the northern Generoso region. The Lecco line is, as the Lugano line, indeed a reactivated Mesozoic normal fault [1].

3.2 ALPINE TECTONICS IN THE AREA OF STUDY

3.2.1 *The general setting*

The northern part of the study area is occupied by the basement. The foliation and the ductile folds are clearly of Variscan age.

The basement underwent brittle deformation during Alpine shortening. *Alpine structures* in the basement of the study area are poorly known. The Variscan foliation is sub-horizontal in the region S of Tesserete (Lugano) (Fig. 2). North of Tesserete, the foliation abruptly becomes steep and dips to the S to SE describing a synform. Since the sediments of the adjacent Valsolda are also folded with a similar geometry, I interpret this change as due to Alpine folding (Fig. 14). Outcrop-scale structures like brittle folds and faults indicate for the region N of Lugano a NE-SW, NNE-SSW oriented, subhorizontal compressional axis with a WNW-ESE striking intermediate axis. The Variscan foliation east of the Swiss-Italian border still dips on the whole to the SE. Brittle, large-scale folding is however common. The basement disappears in the subsurface towards the south beneath the roughly S-dipping sediments.

The sediments north of Val Menaggio are subdivided by two WNW-ESE striking, SSW-dipping thrust surfaces into two major thrust sheets (Fig. 15). The thrust surfaces are generally marked by thin remnants of Rhaetian Riva di Solto Shale (Figs. 6 and 16).

Although most of the Riva di Solto Shale outcrops have been known already since BISTRAM (1903), the juxtaposition with the overlying Dolomia Principale has either not been interpreted or considered to be due to normal faulting (LEHNER, 1952). This is kinematically impossible since the Dolomia Principale overlies the Riva di Solto Shale with a bedding-parallel contact. The thrusts were only implicitly recognized as such by BISTRAM (1903) and, partially, by LEHNER (1952). This is quite surprising since the thickness of the Dolomia Principale, with no thrusts, would be of several kilometers!

Transport seems to have been roughly upwards-directed in present day coordinates. Once the Alpine steepening is removed, transport was towards the north. This vergence is deduced from ramp-folds found at the base of the southern tectonic unit. A N-vergent anticline is found

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[1] The vertical displacement across the Mesozoic Lecco normal fault is much lower than that across the Lugano normal fault. Actually, the two sides of the lake show no drastically differing Mesozoic evolution. The striking difference between the geology of the two sides of the Lecco arm of Lake Como, which was observed already at the end of the 19th century (PHILIPPI, 1895), is mainly a consequence of different geometries of Alpine shortening.

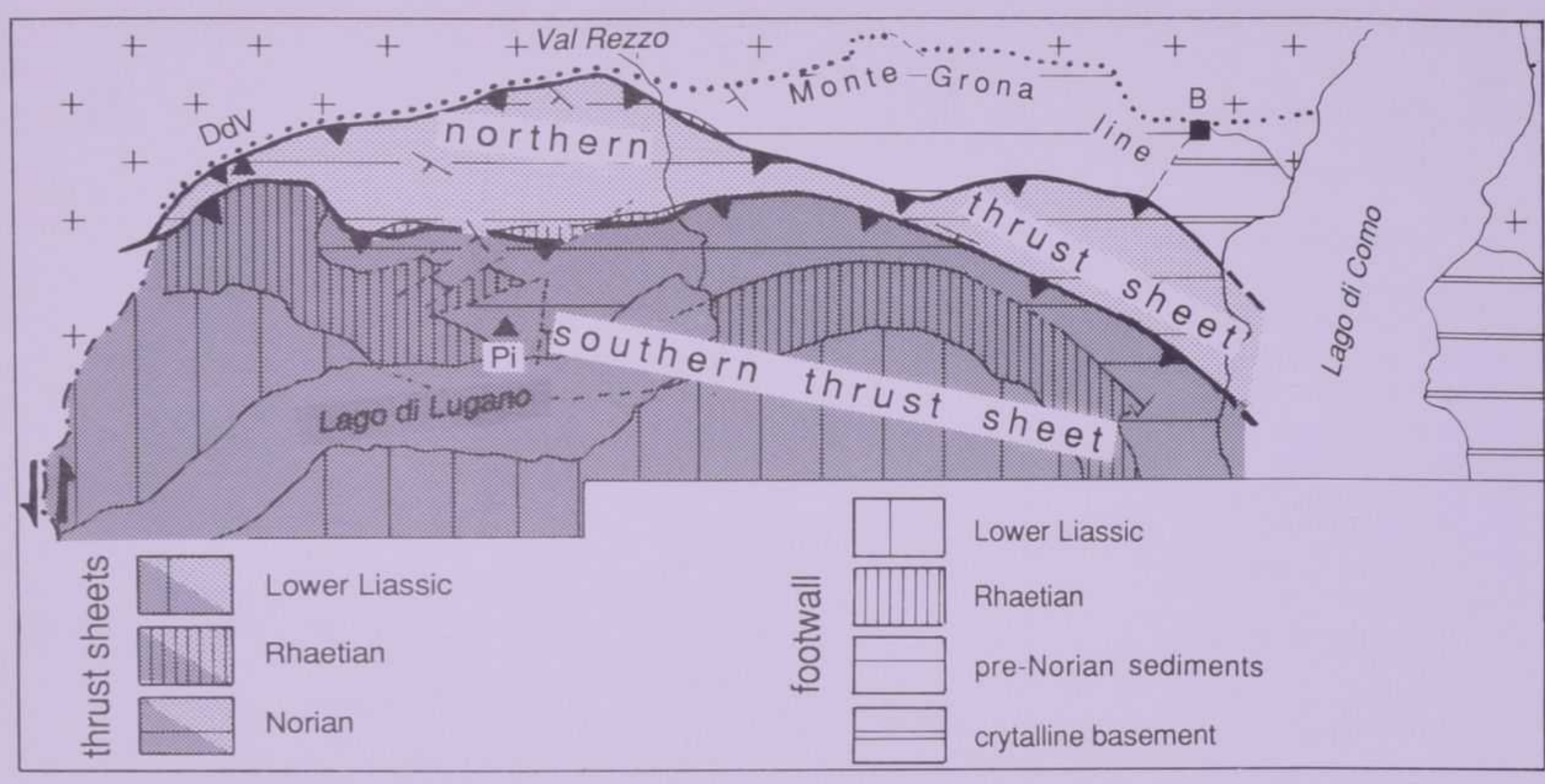

FIG. 15 - Northward-directed Alpine thrusting. The two thrust sheets are thrusted on a composite footwall made up of basement rocks in the west and basement with sediments east of Val Rezzo. The Monte Grona line east of Val Rezzo shows no significant Alpine reactivation. DdV = Denti della Vecchia; B = Breglia; Pi = I Pizzoni.

N of Porlezza (profile d in Fig. 16 and Fig. 17) which can be followed towards the west up to the Sass di Mont (profile b in Fig. 16 and Fig. 18). A geometrically very similar anticline is found at Pian di Scagn (upper Valsolda). Small scale thrusts and folds in the Sasso di S. Martino area (A. SOMMARUGA, 1990 personal communication) also indicate a shortening in a NNE-SSW direction. The thrust sheets thus form a north-vergent duplex system (DAHLSTROM, 1970).

As a result of N-S directed shortening, the outcrops of the tectonic units are presently elongated in a roughly E-W direction (Fig. 15). This means that palinspastic control in the N-S paleodirection, i.e. parallel to trasport, is poor. Moreover, given the present day E-W elongation of the units, the presently visible shape of the thrust sheets is not only a result of shortening but is also substantially controlled by the inherited, mainly N-S trending structural grain.

The *footwall of the northern thrust sheet* east of Val Rezzo (Figs. 6 and 15) is made up of Late Permian to Triassic sediments and crystalline rocks. The upper detachment horizon is bedding-parallel with respect to the footwall. West of Val Rezzo, the sediments wedge out so that the northern thrust sheet comes to lie directly on the basement. The upper detachment of the footwall coincides here with the Monte Grona line (Fig. 15). The formations of the footwall are continuous with the similar rocks east of Lake Como, north of the Grigna Group (Figs. 6 and 14).

The *northern thrust sheet* is made up of Dolomia Principale and Zorzino Formation with thin but relatively continuous remains of Riva di Solto Shale. The *southern thrust sheet* shows a complete succession through the Rhaetian and the Early Liassic overlying the Norian carbonates and forms the lower part of the Moltrasio Limestone outcropping in the Monte Generoso area.

Dips of bedding become more gentle south of Val Menaggio (Fig. 16) and the Late Triassic sediments disappear beneath the Moltrasio Limestone of the Monte Galbiga-Monte Generoso area.

3.2.2 *Thrust-surfaces and thrust-sheets*

The roof thrusts of the duplexes have, in the presently E-W elongated sections, a "flat" geometry and run more or less along the base of the Riva di Solto Shale and are therefore subparallel to bedding in the underlying sediments (Fig. 19). The sole thrusts have a more complex geometry: in the eastern part of the thrust sheets they run either along the Dolomia Principale-Raibl Beds contact or between Dolomia Principale and Zorzino Formation. Along these segments the bottom thrusts are bedding-parallel. In the western part of the tectonic units the sole thrusts climb upsection towards the west. This explains why the base of the thrust-sheets becomes younger towards the west (Figs. 6 and 19). This geometry is not a consequence of an Alpine lateral ramp as demonstrated by the absence of corresponding

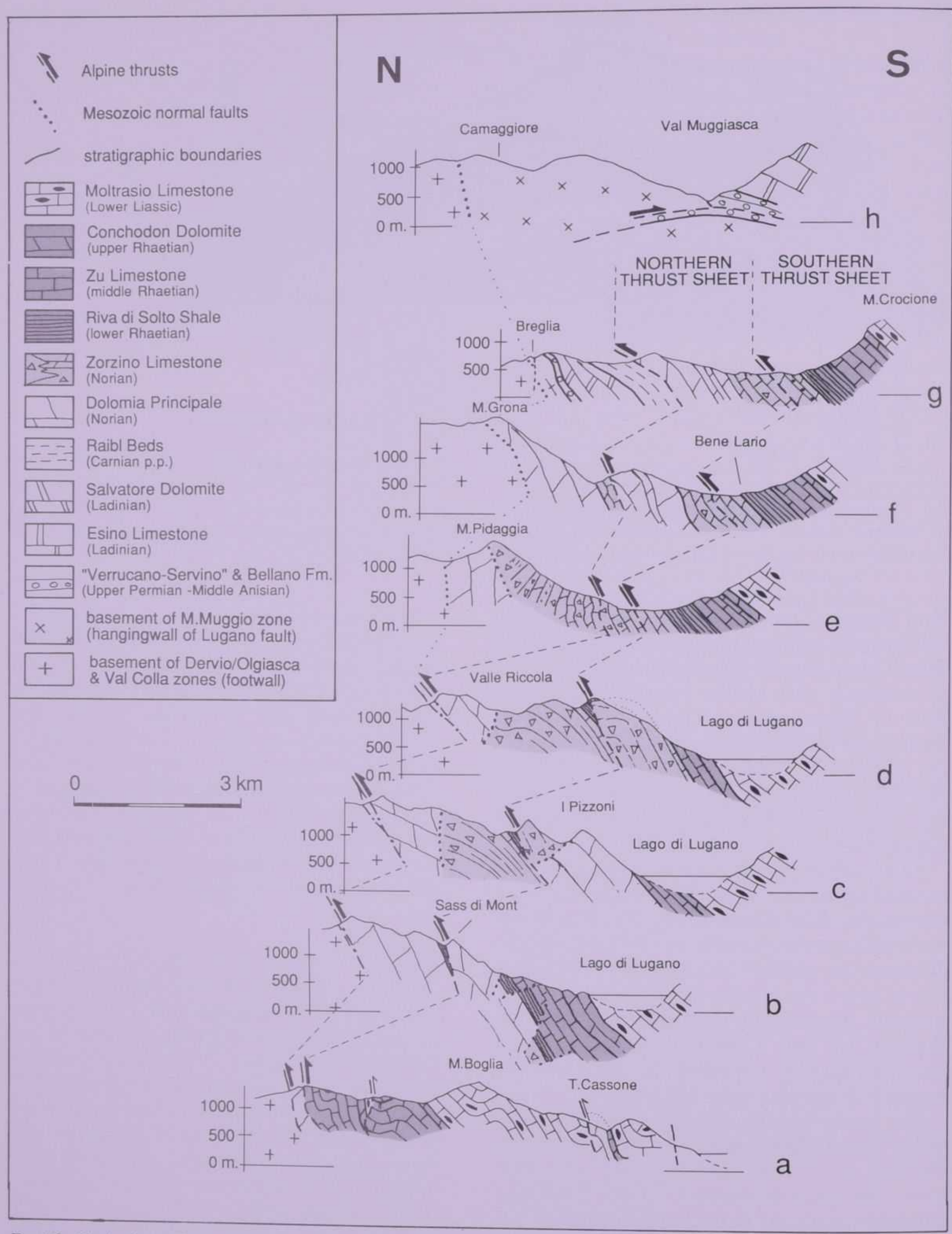

FIG. 16 - Geologic profiles across the study area. Location of profiles is given in figure 6.

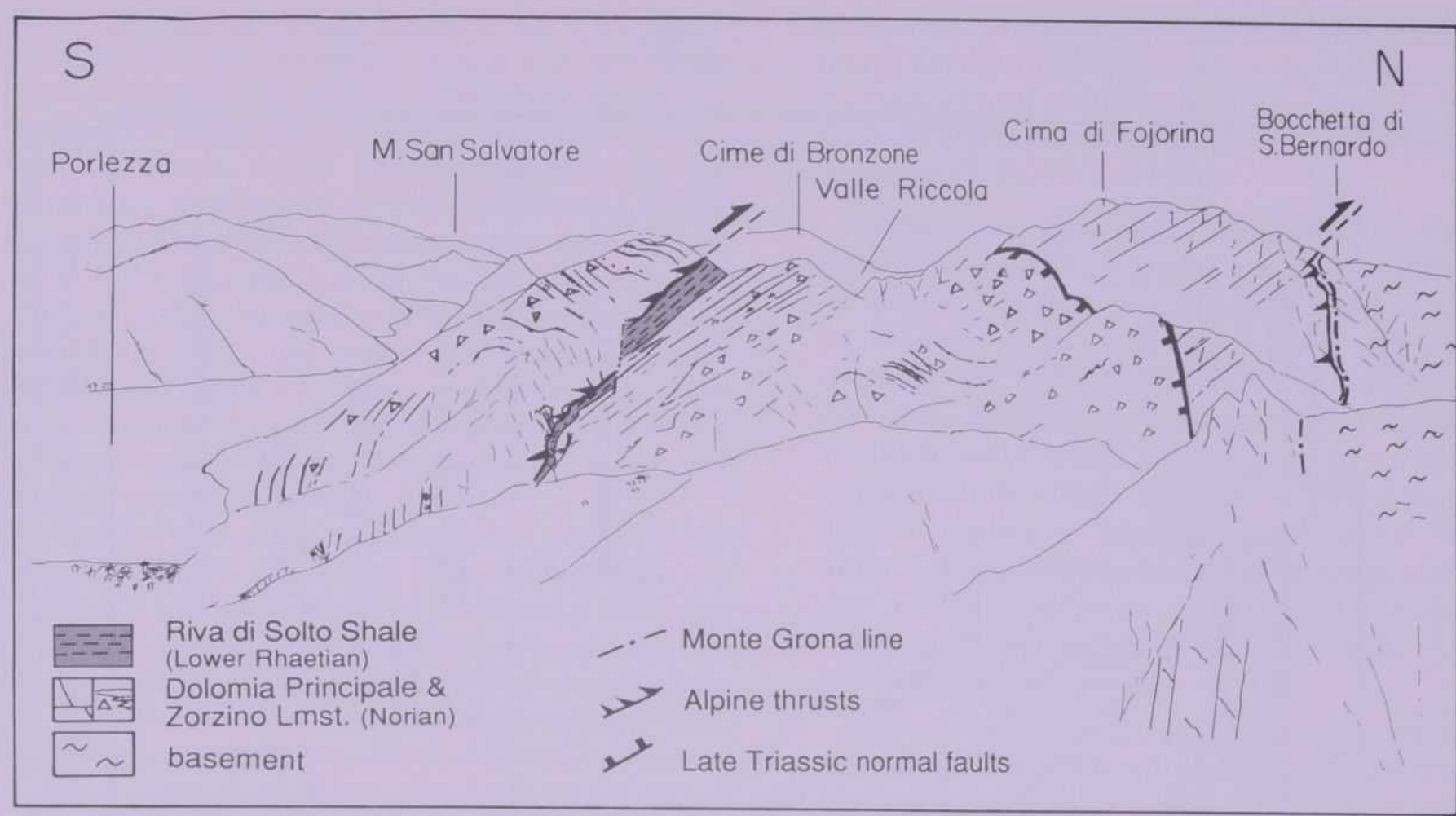

FIG. 17 - Relationships between the southern and the northern thrust sheets. Note north-vergent ramp fold in the southern unit; the M. Mugetto synsedimentary fault in the northern tectonic unit is also visible. Sketch from photograph taken from M. Pidaggia.

sediments in the footwall. The discordant bases of the tectonic units were shaped before Alpine shortening and coincided with the hangingwall of the Lugano normal fault. This takes care of the absence of Early and Middle Triassic sediments along the Monte Grona line (stratigraphic cut-off related to normal faulting) and of non-parallelism between Monte Grona line and bedding (Figs. 3 and 19). The angles between bedding and Monte Grona line are thus cut-off angles of the Mesozoic Lugano fault. As a matter of fact, a very similar geometry is shown by the rocks of the footwall of the Alpine thrust system. These rocks are discordantly cut by the Monte Grona line which represents the northern steepened segment of the Lugano line with no substantial reactivation in Alpine times.

The *footwall* of the thrust sheets is formed by basement rocks west of Val Rezzo. E of the valley, the upper decollement leaves the basement/sediment contact (i.e. the Monte Grona line) and runs on top of the Norian sediments with thin remnants of Rhaetian Riva di Solto Shale. Further to the east a Mesozoic fault juxtaposes the Dolomia Principale of Monte Grona with the Lower Triassic to Carnian sediments of the Nobiallo region. The upper detachment runs here above the Raibl Beds (Fig. 6). East of Val Rezzo the Monte Grona line shows no evidence for significant Alpine reactivation.

The *northern thrust sheet* is found from the Denti della Vecchia in the west to Lake Como in the east. It is not found on the eastern side of Lake Como being limited by the Lecco transfer zone (Fig. 14). The unit is made up essentially of Dolomia Principale and Zorzino Formation. Only thin remnants of the overlying Riva di Solto Shale are preserved more or less continuously from Val Rezzo in the east to Valsolda in the west. The base of the unit is roughly bedding-parallel from Lake Como to Val Sanagra. West of this valley, the base of the unit climbs westwards upsection until it reaches the base of the Zorzino Formation. The base of the unit remains then bedding-parallel until Val

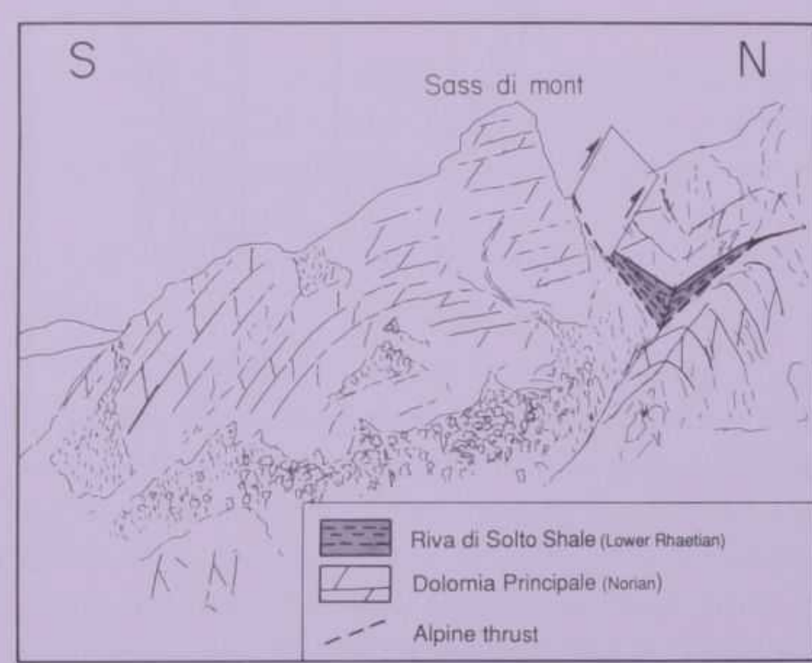

FIG. 18 - Thrusting of the southern thrust sheet onto the northern one in the Sass di Mont area. Sketch from photograph.

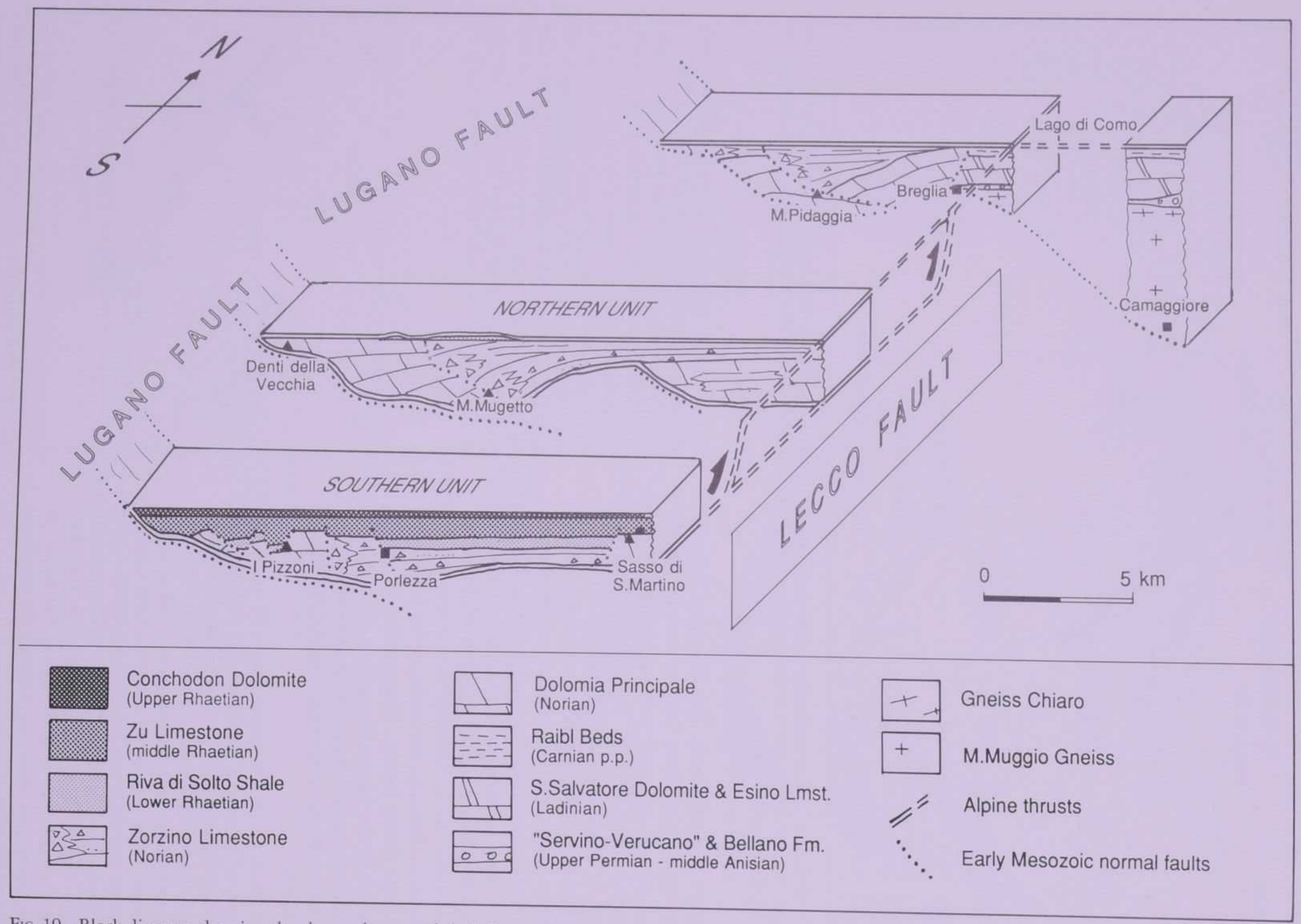

FIG. 19 - Block diagram showing the thrust sheets and their footwall palinspastically restored to their pre-Alpine situation. Only the preserved record, as visible in the field has been shown. From the southern unit the Moltrasio Limestone has been removed for clarity. Note bedding-parallel roof detachment surfaces; the sole thrust can be bedding-parallel (in the east) and oblique (in the west). In the latter case it utilizes the pre-existing Lugano normal fault.

Rezzo and returns oblique further to the west (Figs. 6 and 19). The bottom thrust merges with the Monte Grona line west of Val Rezzo. The Monte Grona line west of Val Rezzo must therefore have been reactivated as an Alpine thrust. This is compatible with the small-scale imbrications of Verrucano-Servino, Gneiss Chiaro and Manno Conglomerate found in the uppermost Val Rezzo and west of the Swiss-Italian border (LEHNER, 1952). No similar imbricates are found along the Monte Grona line east of Val Rezzo. The upper decollement horizon is always bedding-parallel. As for the footwall of the thrust system, angular relationships between upper and lower thrust surfaces are such that the unit wedges out towards the west.

Only the base of *the southern unit* is exposed in the area of study. It consists of Dolomia Principale and Zorzino Formation overlain by the Rhaetian and Lower Liassic formations. The latter is continuous with the Moltrasio Limestone of the Monte Generoso. The base of the unit runs parallel to bedding from Lake Como till Porlezza; further westwards it becomes oblique to bedding climbing up-section towards the west (Figs. 6 and 19).

3.2.3 *Folds*

Since the most widespread lithologies are the massive carbonates of the Dolomia Principale, folds are not very common in the study area. Folds are limited to the well-bedded facies of the Zorzino Formation, to the Zu Limestone and to the Moltrasio Limestone.

At a meter- to hundred meters-scale two main groups of folds can be genetically and geometrically distinguished: fault-related folds and buckle folds.

Fault-related folds. These folds are always found above tectonic surfaces (thrusts or normal faults). They form when the irregularly-shaped base of a tectonic body is brought on a geometrically different surface. Typical examples are ramp-folds (fault-bend folds of SUPPE, 1983) in compressional domains and "extensional ramps" in extensional domains (DAHLSTROM, 1970). Thrust-related ramp folds in the area of study have been illustrated in chapter 3.2.1. Comparable folds are obtained when the irregularly-shaped hangingwall of a normal fault is successively thrusted on the bedding-parallel top of the footwall. Folds of this kind are mainly found in the western part of the study area and are geometrically characterized by steep fold axes.

A nice example of an extensional ramps is given by the Dolomia Principale of the northern thrust sheet in the region between Bocchetta di S. Bernardo (Swiss-Italian border) in the east and the Denti della Vecchia in the west (folds number 6 and 7 of Fig. 14). The top of the Dolomia Principale swings from a E-W direction in the east to NW-SE and back again to E-W in the west. All this while the base of the thrust sheet along the Monte Grona line shows no similar changes and keeps a WSW-ENE direction.

Buckle folds. These folds are common in the Zu Limestone of the Valsolda and in the Moltrasio Limestone. Their general shape is asymmetric: the folds usually mimic at a smaller scale the shape of the regional synclines and anticlines. Anticlines generally have a relatively flat northern limb and a steep, S-dipping to overturned, southern limb. They sometimes evolve into thrusts. Folding is typically accommodated by flexural slip between bedding: well developed calcitic flexural slip striae are nearly ubiquitous. Folds axes generally strike NW-SE to E-W and rarely show strong plunges.

Fold axes in the *Zu Limestone* of the upper Valsolda generally show a NW-SE trend (Fig. 14). The axes are however undulated by later N-S striking folds.

Folds in the Moltrasio Limestone are very common. They show an interesting pattern and have therefore been looked at in more detail. Fold axes generally strike NW-SE in the western part, *i.e.* close to the Lugano line and roughly E-W in the central part of the basin (Fig. 20). Relevant is the observation that while fold axes close to the Lugano line have intermediate plunges to the NW, fold axes in the central part of the basinal succession are flat. A correlation seems to exist between steepness of the fold axes and the distance from the Lugano line, that is from the Lugano normal fault. Since in the same distance an increase in thickness of the Moltrasio Limestone must occur, I interpret the described pattern as being controlled by the geometry of the Moltrasio Limestone body. More specifically the westward plunge of fold axes is a geometric consequence of the stronger uplift experienced during Alpine shortening by the central, *i.e.* thicker part of the sedimentary body compared to the western, thinner part (Fig. 20).

The study area is then interested by large-scale, N-S striking open folds (folds number 4 and 5 in Fig. 14). Their geodynamic origin is not known.

3.3 A PROFILE FROM THE INSUBRIC LINE TO THE PO PLAIN

The observations made in the study area are here integrated with data from the literature and with personal observations in order to construct a geologic profile from the Insubric line to the Po Plain. The profile runs in a NNE-SSW direction along the Lago di Como transect in the north and continues in a N-S direction, shifted by some 10 km further to the west (Fig. 14).

The most prominent feature in the northern part of the profile is a major, E-W trending anti-

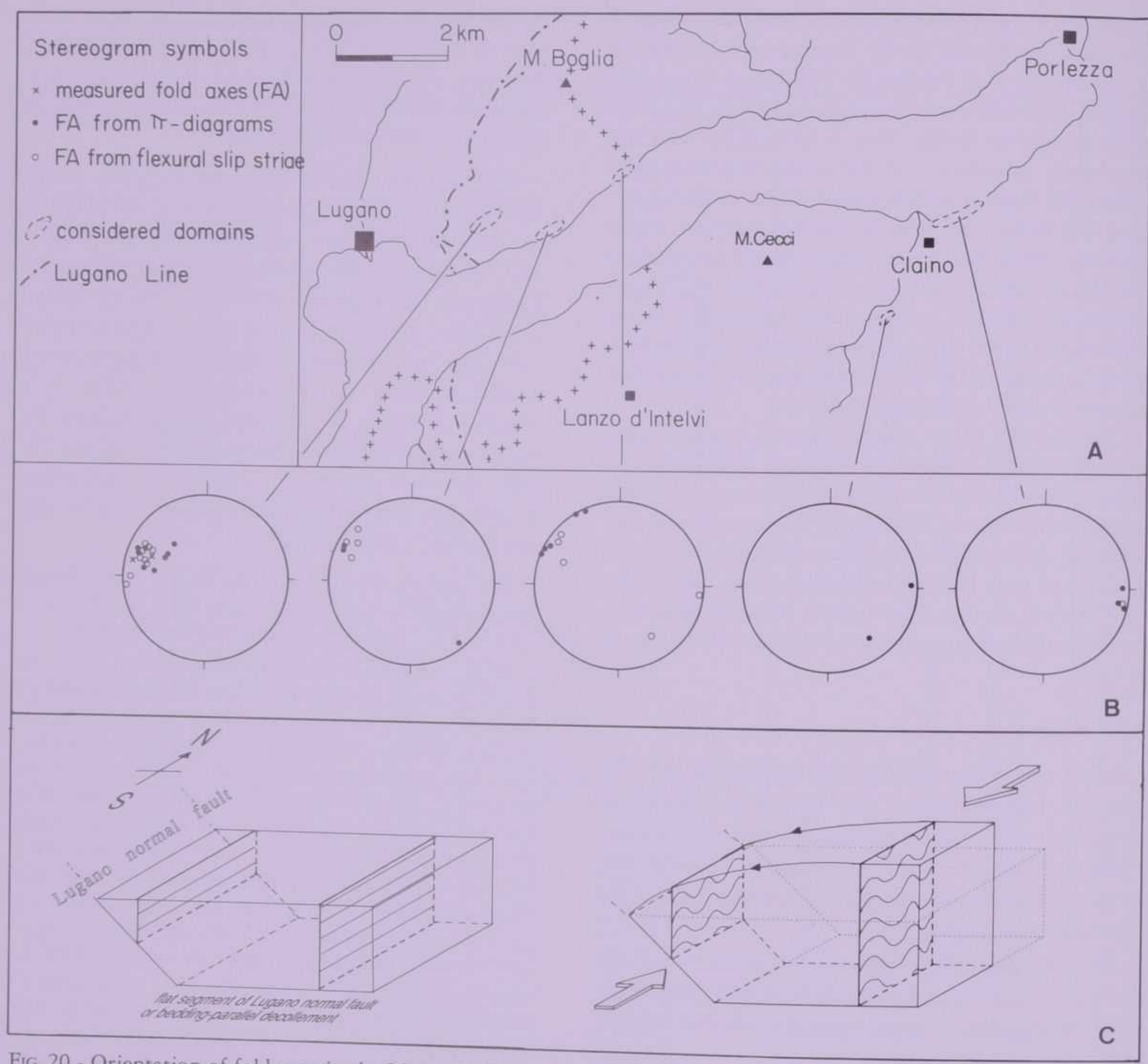

FIG. 20 - Orientation of fold axes in the Moltrasio Limestone in different localities of the study area (A and B). Note westward plunges in the west and flat to gently east-dipping axes in the east. (C) block-diagram illustrating the proposed explanation: plunges towards the syn-sedimentary fault are acquired once shortening affects a body with laterally decreasing thickness.

clinal structure located N of the Monte Grona/Val Grande line (n. 1 in Figs. 14 and 21). The anticline, which will be referred to as Val Varrone anticline, has been postulated by EL TAHLAWI (1965) on the base of observations made east of Lake Como. Its prolongation west of the lake is less clear because the northern limb of the fold is cut by a normal fault of unknown importance. The core of the anticline is formed by the Dervio-Olgiasca zone and by the eastern prolongation of the Val Colla zone. The anticline folds the sediments and thrusts of the study area and must therefore be late Alpine. The anticline must also fold pre-Alpine structures like the northern part of the Lugano normal fault, *i.e.* the Monte Grona/Val Grande fault zone. I propose that the prosecution of the Monte Grona/Val Grande fault zone north of the anticline should be identified with the fault zone roughly coinciding with the Musso line and its eastward continuation north of Piona (Fig. 14). Following this interpretation the Gravedona zone, similarly to the Monte Muggio zone, would have been, during the Mesozoic, part of the hangingwall of the Lugano normal fault. I have already underlined the similarities between the Gravedona zone, north of the Musso mylonitic zone and the Monte Muggio zone, south of the Monte Grona/Val Grande fault zone. Both represent shallower crustal levels with their sedimentary cover (Table 1).

Preliminary investigations in the *Piona area* have allowed for the recognition of fault rocks comparable

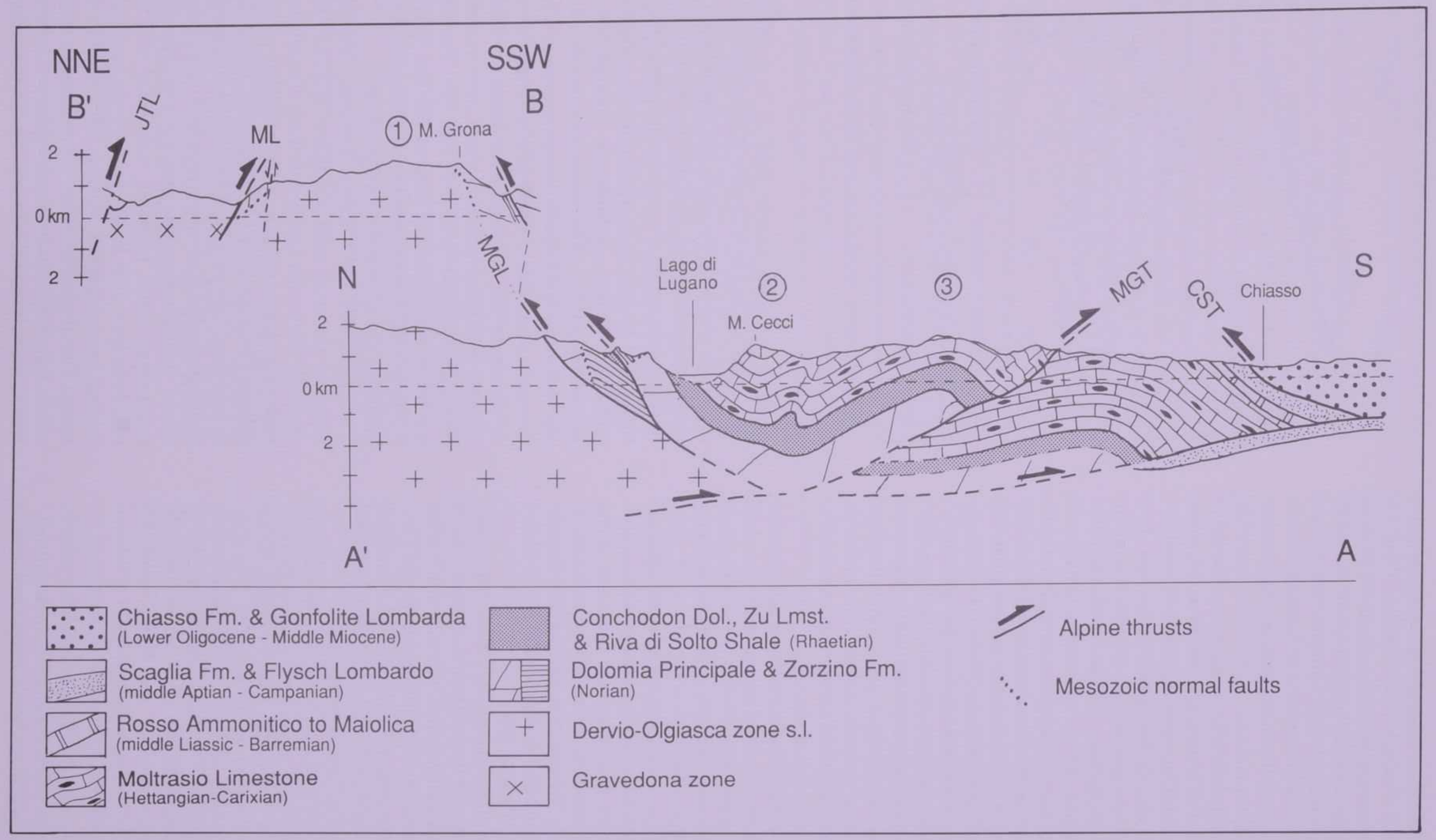

FIG. 21 - Geologic profile from the Insubric line to the Po Plain. Northern part of the profile from FUMASOLI (1974) and REPOSSI (1904) modified. Southern part from BERNOULLI *et al.* (1989). Circled numbers and capital letters refer to structures in figure 14 and mentioned in next. Profile trace is given also in figure 14.

to those found in the Camaggiore area (chapter 4.3.3). In particular, in Montecchio (north of Piona) (Fig. 14) mylonites and ultramylonites are found which strike E-W. The age of the Piona mylonites is constrained by the Triassic radiometric ages obtained from the syn-mylonitic pegmatites (chapter 2.2.1) (HANSON *et al.*, 1966). At least some of these dykes are parallel to the mylonitic foliation and seem to be syn-mylonitic (B. ILDEFONSE, 1989, personal communication). The age of the mylonites would therefore be similar to that postulated for the Monte Grona/Val Grande fault zone.

The outcrops in the *Musso area* are not very clear, so that only a tentative reconstruction will be given. The general setting (Fig. 22) is characterized by steeply dipping schists with an intercalation of Paleozoic marbles discordantly overlain by non-metamorphic dolomites. The latters have been paleontologically attributed to the Norian and therefore correlated with the Dolomia Principale (REPOSSI, 1904). The Dolomia Principale is generally thought to be flat lying and directly in contact with the basement without the interposition of Late Permian to Middle Triassic sediments.

Preliminary investigations (partly together with G. SCHÖNBORN) have produced new important informations. The following points are relevant for the understanding of the geology of Musso:

a) Bedding in the Dolomia Principale is clearly discordant with respect to the basement-sediment contact. While bedding dips 20°-40° to the west (occasionally to the south), the base of the Dolomia Principale clearly dips to the east;

b) Slivers of Gneiss Chiaro and of Manno Conglomerate can be observed along the basement/sediment contact. The Gneiss Chiaro has been found west of the Dolomia Principale; the Manno Conglomerate lies at the base of the Dolomia Principale and has been observed and paleontologically documented by LEPORI (1961) in a hydroelectric pipe.

In spite of several uncertainties, there is a number of geologic features of the Musso area which closely recall the situation along the southern limb of the Val Varrone anticline like in Val Rezzo (see chapter 4.3.3). These elements support the interpretation of the Musso Dolomite and the Gravedona zone being part of the hangingwall of the Lugano normal fault.

GIANOTTI and TANNOIA (1988) have recently proposed a transgressive nature for the base of the Dolomia Principale of Musso in order to explain the absence of the lower part of the succession. This interpretation is strongly influenced by their interpretation of the Monte Grona/Val Grande line. Both because of the new findings presented above and because of the general interpretation of the Monte Grona/Val Grande line, I do not agree with the model of GIANOTTI and TANNOIA (1988).

Proceeding with the discussion of the profile in figure 21, two possibilities exist in order to explain the origin of the Val Varrone anticline: it could be a simple buckle fold or it could be a ramp fold related to a deeper S-directed thrust. I prefer the second alternative because it better explains the substantial shortening accommodated by the sediments on the southern limb of the anticline and because it is kinematically compatible with the Grigna profile as interpreted by LAUBSCHER (1985). The basement north of the Monte Grona/Val Grande line is continuous with the uppermost thrust sheet of the Grigna profile as shown by the continuity of Early to Middle Triassic sediments across Lake Como north of Menaggio and near Varenna (Fig. 14). Therefore the thrust running at the base of the Val Varrone anticline would correspond to the westward prolongation of LAUBSCHER's Orobic thrust.

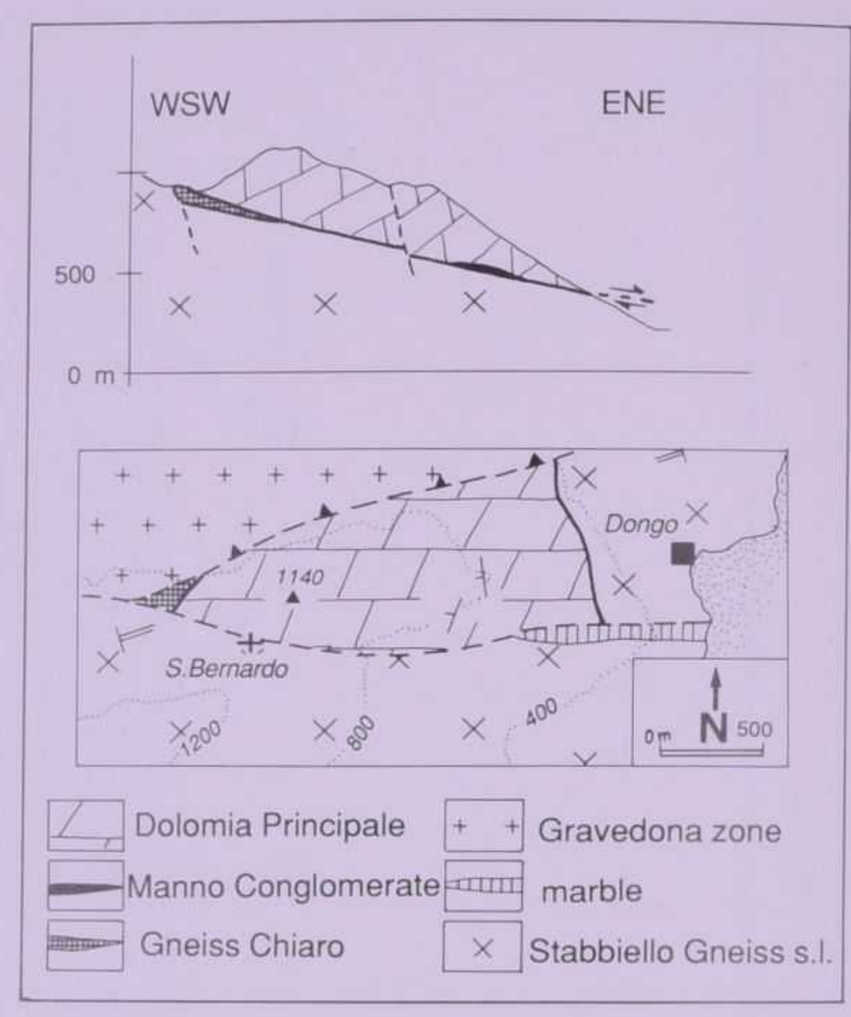

FIG. 22 - Geologic profile and map of the Musso area.

South of the Val Varrone anticline, the sediments of the study area are found (Fig. 6). They generally dip to the south and form the northern limb of a major, roughly E-W trending syncline (Monte Cecci syncline of BERNOULLI, 1964) (fold n. 2 in Figs. 14 and 21). The syncline formed in front of the Val Varrone ramp-fold. The Orobic thrust should run at the base of the preserved sedimentary units, probably consisting of Dolomia Principale (the lower part of the succession having been removed during Mesozoic extension). The thrust branches in an upper thrust which forms the south-vergent Monte Generoso thrust well described already by FRAUENFELDER

(1916) and a lower one kinematically related to the Po Plain thrust system (PIERI and GROPPI, 1981).

The Mesozoic succession is limited to the south by a back-thrust, the Chiasso-Stabio thrust (BERNOULLI *et al.*, 1989) (Figs. 14 and 21). The Moltrasio Limestone together with the younger Mesozoic units forms an indenter between the base of the Oligocene Chiasso Formation and the older formations (see JONES, 1982 for a somewhat similar geometry in the Alberta foothills of Canada).

The presence of a major thrust underlying the M. Generoso area indicates that the Monte Generoso sediments are detached from the basement and form a large pop-up structure (BUTLER, 1982).

It is very difficult to say something about the westward prolongation of the described structures. The basement of the Lugano region seems not to be separated by major thrusts from that of the Val Colla area (Fig. 14); this would imply that the Orobic thrust remains at depth and is not kinematically linked to the Lugano line. This implies also that the Lugano line is allochthonous, as the sediments east of it and the whole Lugano region.

In Alpine times the Lugano line acted as a transfer zone between two zones with differing superficial shortening geometry and thrusts east of the line do not necessarily have a direct prolongation west of it. It is therefore conceivable that different segments of the Lugano line experienced different senses of movement depending on the amount of shortening accommodated by the different thrust and fold systems on the two sides of the line (BERNOULLI, 1964). In the northernmost segment of the Lugano line, where the eastern sector was shortened with N-vergent thrusts, Alpine reactivation is likely to have been sinistral. Further to the south, with the sediments east of the Lugano line forming the south-vergent Monte Generoso thrust, movements are likely to be dextral (as shown by small-scale structures in Valle di Lembro near Rovio).

4 MESOZOIC EXTENSION: THE MONTE GENEROSO BASIN AND THE MONTE GRONA/VAL GRANDE FAULT ZONE. CRUSTAL EXTENSION IN THE WESTERN LOMBARDIAN BASIN

4.1 INTRODUCTION

The sedimentary and stratigraphic record (chapter 2) clearly shows that the area east of the Lugano line was in early Mesozoic times part of a strongly subsiding region, the Monte Generoso basin. The subsidence of the basin is directly related to extensional faulting, itself related to the Mesozoic rifting of the Southern Alps. Evidence from sedimentary record shows that the evolution of the Generoso basin was mainly controlled by an east-dipping crustal normal fault: the Lugano fault.

As a consequence of Alpine shortening (chapter 3) the northern segment of the Lugano normal fault was brought into a steep position and is now, after erosion, exposed at the surface. This steepened segment of the Lugano normal fault has been identified with the Monte Grona/Val Grande fault zone (Fig. 3). The exposure at the earth's surface of a segment of a crustal normal fault is obviously a rare occasion to look in detail how deformation occurs along such a fault. Since the Lugano normal fault originally dipped to the east and since it has been folded around a roughly E-W trending anticline, deeper paleo-structural levels of the fault are exposed at the surface moving from west to east in present day coordinates. Therefore the chance is given to look at deformation along the Lugano normal fault at different crustal levels.

Facies distribution and thickness changes in Norian to Rhaetian formations east of the Lugano line, further indicate that normal faulting was going on also inside the Monte Generoso basin itself. These normal faults are presently visible in the different sedimentary units.

The faults observed in the sedimentary units will be described separately from the fault rocks of the Monte Grona/Val Grande fault zone (chapters 4.2 and 4.3 respectively); the sedimentary units are in fact detached from their substratum and the two data sets are therefore not directly comparable. The two data sets will be then integrated to describe the geometry and evolution of the Monte Generoso basin (chapter 4.4).

The Monte Generoso basin will then be compared with the Monte Nudo basin lying west of Lugano. Considering the two basins, an upper crustal profile can be constructed running from the Canavese in the west to the Albenza plateau in the east (Fig. 1). On the base of the profile, a stretching factor for the western Lombardian basin will be estimated (chapter 4.5). The tectonically-derived stretching factor will be used to constrain, together with the data derived from the sedimentary record, the subsidence history of the basin (chapter 5).

4.2 EXTENSIONAL STRUCTURES IN THE SEDIMENTARY UNITS

In this chapter I shall describe the extension-related tectonic structures as visible in the sedi-

mentary units. I shall first give a description of the single structures and then remove the effects of Alpine shortening in order to discuss the paleogeographic distribution of the faults.

4.2.1 *Extensional structures in the area of study*

The Lugano normal fault

The Lugano normal fault was active from the Norian to the Late Liassic (chapter 2.3.1). It accommodated during the Norian a vertical displacement of at least 1300 meters but had no morphological expression at the surface: the Dolomia Principale on the two sides of the fault is very similar. A certain difference existed during the Rhaetian between the Lugano swell with very shallow water conditions and the Generoso basin where deposition was going on mostly below wave-base. The transition between the two domains, however, was smooth as suggested by the absence of redeposited material east of the normal fault and by the limited water-depth difference. The situation drastically changed in the Early Liassic with the foundering of the Generoso basin while the Lugano swell remained under shallower water conditions. At the same time the fault-controlled basin margin migrated westward progressively incorporating into the basin new fault-bounded segments which were previously parts of the Lugano swell [(1)].

While strong subsidence was going on in the Generoso basin, the Lugano swell was affected by intensive sub-marine syn-sedimentary tectonics. Mainly NNW-SSE striking faults, small grabens and smaller-scaled fractures cut the Norian to Early Liassic formations. Veins and fractures were filled with marine sediments and submarine carbonate cement and the rocks were repeatedly fractured forming a complex breccia known as Macchia Vecchia Breccia (WIEDENMAYER, 1963; BERNOULLI, 1980).

The vertical displacement accommodated during the Rhaetian to Late Liassic as deduced from comparison of sedimentary columns not corrected for compaction amounts to ca. 5000 meters.

Extensional faults east of the Lugano line

The sediments east of the Lugano line are subdivided by two thrust surfaces into two major thrust sheets presently elongated in a E-W direction (Fig. 6 and chapter 3.2). Extensional features will therefore be treated separately for the two units and for their footwall (Fig. 15). The footwall of the thrust sheets and the northern tectonic unit show only pre-Rhaetian formations so that the faults can be described only for this time interval. The southern tectonic unit consists also of younger formations allowing a reconstruction of younger movements. Since all sediments have been steepened during Alpine orogeny, the present day map (Fig. 6) can be looked at as a partially distorted paleo-vertical section (Fig. 3).

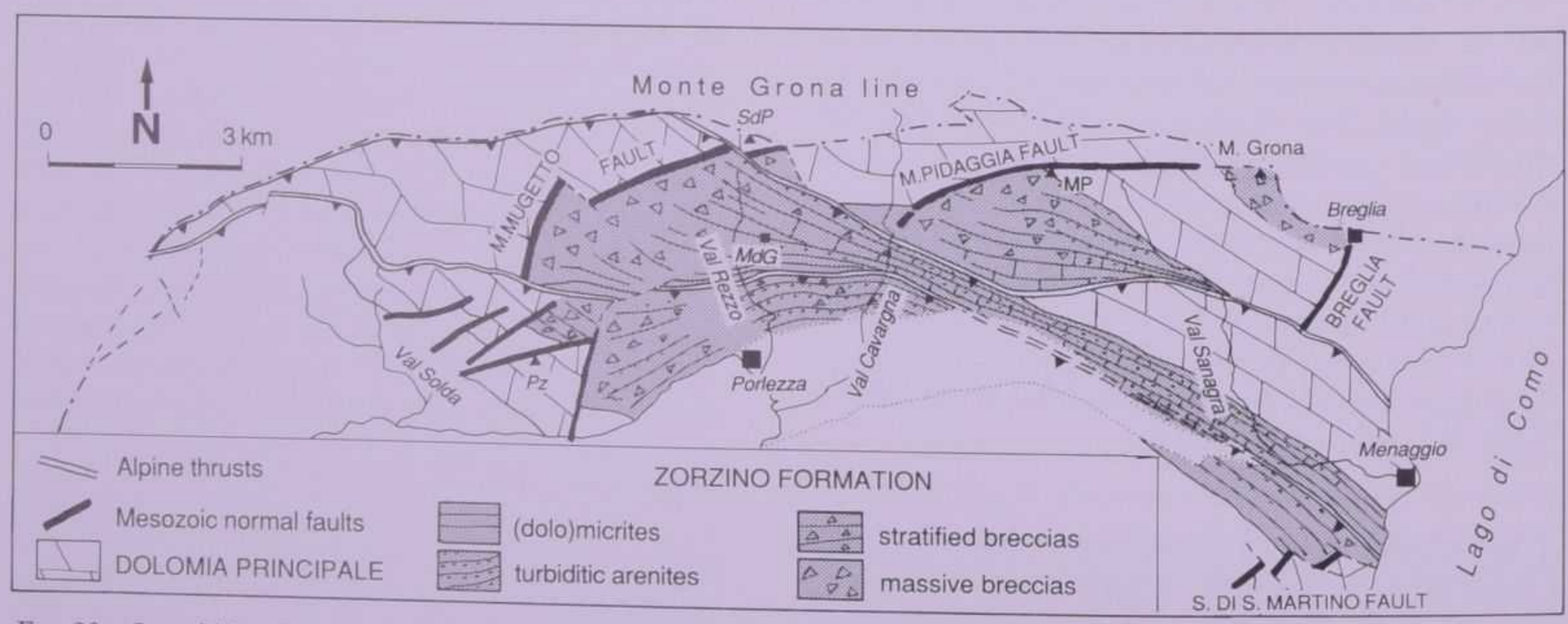

FIG. 23 - Late Triassic syn-sedimentary normal faults and facies distribution of the Zorzino Formation. SdP = Sassi della Porta; MdG = Monti di Gnin; Pz = I Pizzoni; MP = Monte Pidaggia.

Three major normal faults are found in the *footwall of the imbricates*. From west to east they are the Sassi della Porta fault, the Monte Pidaggia fault and the Breglia fault (Fig. 23). At the west-

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(1) A similar geometry of progressive incorporation of parts of the footwall into the hangingwall has been described across the Garda escarpment (CASTELLARIN and PICOTTI, in press).

ern termination of the sediments a presently WSW-ENE striking fault separates the platform dolomites of the Sassi della Porta in the west from massive carbonate breccias (Zorzino Formation) in the east. The thickness of the Zorzino Formation is here limited to a few hundred meters; however, the Dolomia Principale shows strongly differing thicknesses east and west of the fault so that the vertical displacement must have been of the order of several hundred meters.

Further to the east a major fault, the Monte Pidaggia fault, limits to the NNW the redeposited sediments of the Zorzino Formation outcropping on the southern flank of Monte Pidaggia (Fig. 23).

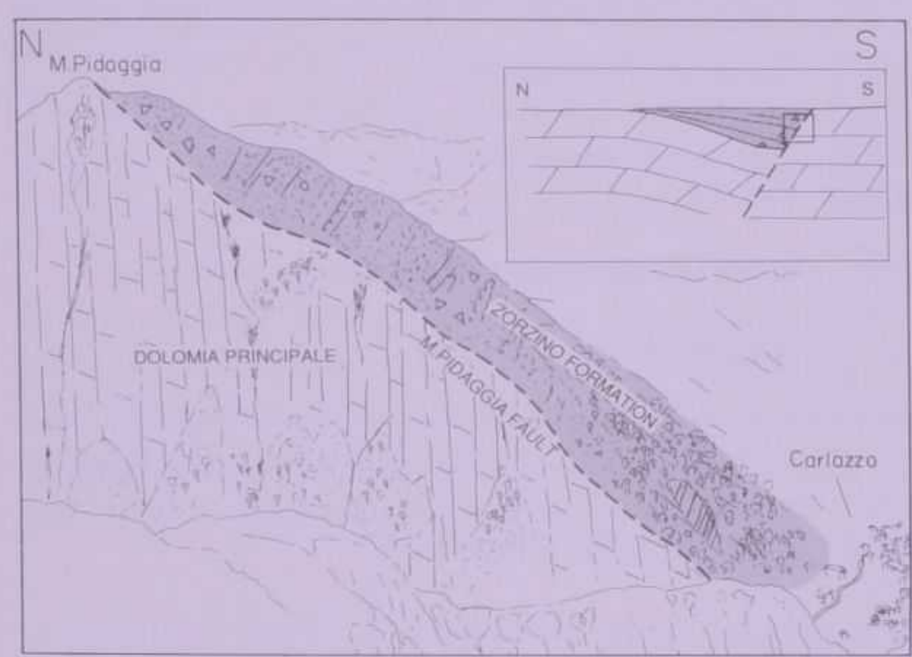

FIG. 24 - Sketch from photograph of the western flank of Monte Pidaggia. Note the presently S-dipping Late Triassic Monte Pidaggia normal fault separating the sub-vertical Dolomia Principale from the N-dipping breccias of the Zorzino Formation. Inset: schematic representation of the fault during the Norian after the Alpine steepening is removed. The square shows the area covered by the sketch.

The fault is beautifully exposed on the western side of Monte Pidaggia (Fig. 24). The relationship between bedding and fault is such that the fault descends downsection from west to east. From the Val Cavargna to the summit of Monte Pidaggia the fault marks the boundary between the sub-vertical, E-W striking Dolomia Principale and the breccias and arenites of the Zorzino Formation. East of the summit, that is at deeper paleostructural levels, the fault lies inside the Dolomia Principale separating two blocks with different orientation: the NW-SE striking hangingwall in the south and the E-W striking footwall in the north. Facies distribution and depositional geometry of the sediments deposited in the Monte Pidaggia basin are controlled by the evolution of the fault (Fig. 23). No significant angular unconformity is observed in the hangingwall either between the Dolomia Principale and the Zorzino Formation or in the Zorzino Formation itself. A fanlike arrangement of bedding in the Zorzino Formation is observed.

The footwall of the fault suffered only limited deformation: only very close to the fault some dm-long, irregularly oriented syn-sedimentary dykes are found filled with dolomitic bioclastic-oolitic grainstone similar to that of the Dolomia Principale. Up to few m-thick horizons of tectonic breccias are found along the fault: the breccia consists of well-rounded, mm-to cm-sized dolomitic clasts in an abundant cataclastic matrix (presently a dolomicrosparite). Similar breccias have been described from Mesozoic extensional faults in the Austroalpine domain (FROITZHEIM, 1989). I agree with FROITZHEIM in considering the abundant cataclastic matrix as characteristic for Mesozoic extensional faults in contrast with breccias from Alpine thrusts which show large proportions of cement.

Further to the east, in the region south of Breglia, a poorly exposed, NE-SW striking fault separates the Dolomia Principale of Monte Grona in the west from the Early and Middle Triassic succession of the Nobiallo region to the east (Figs. 6 and 23). The fault represents a puzzling feature and is generally considered to be somehow related to the Lecco fault and to its Alpine reactivation (LAUBSCHER, 1985; PEROTTI, 1987). However, since the fault is cut to the south by an Alpine thrust I propose a Mesozoic age for the fault. The Breglia normal fault would thus be a Mesozoic NW-dipping normal fault lowering the M. Grona block and, possibly, represents the northern prolongation of the Mesozoic Lecco fault.

A Mesozoic age for the fault also explains the NW-ward thinning body of sedimentary breccias and conglomerates found at the base of Monte Grona (chapter 2.3.2) (Fig. 23). These breccias were deposited in a basin formed very early in the Norian which was filled with platform-derived material coming from the Dolomia Principale formerly overlying the Raibl Beds of the Nobiallo region. The basin was then filled and sutured by the progradation of the Dolomia Principale platform (for a similar case in the Lake Garda region see PICOTTI and PINI, 1988). The Dolomia Principale overlying the Raibl Beds east of the fault could have been removed during Alpine thrusting.

In the *northern thrust sheet*, a major, now NE-SW striking normal fault separates the thick platform succession of the upper Valsolda region in the west from the huge body of redeposited sediments found all across the middle Val Rezzo (Fig. 23). This fault is referred to as Monte Mugetto fault.

The Monte Mugetto fault (Fig. 17) separates along its whole length platform carbonates in the west from slope and basin deposits in the east. The fault is cut by Alpine thrusts both at the bottom (presently north)

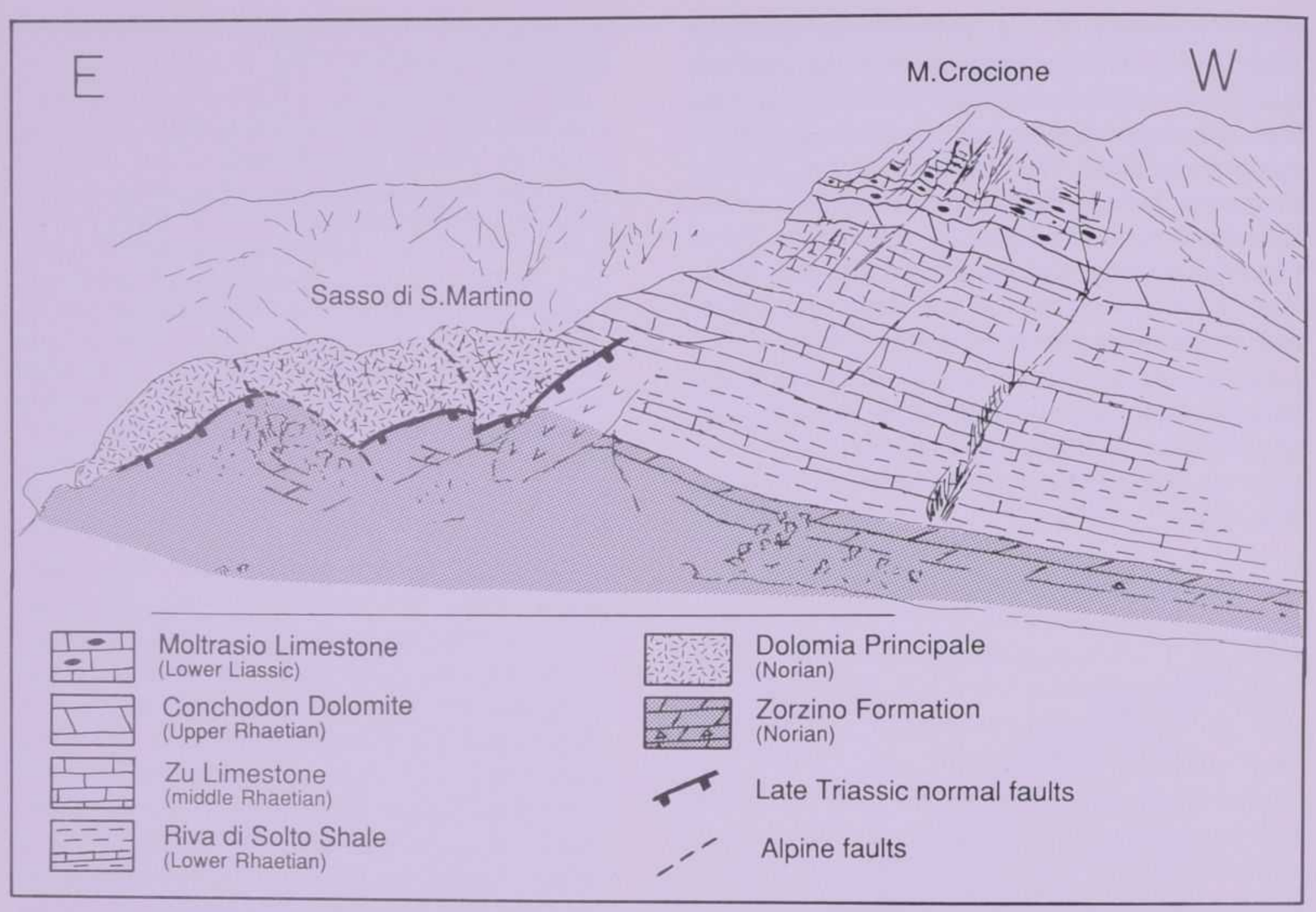

FIG. 25 - Sketch from photograph of the Sasso di S. Martino - Monte Crocione area. The Norian syn-sedimentary fault separates the massive Dolomia Principale of the Sasso di S. Martino from the Zorzino Formation. Note that the upper part of the Zu Limestone and the Conchodon Dolomite are not displaced by the fault.

and at the top (south in present day coordinates). The fault must have had a strong morphologic expression with the fault scarp furnishing large amounts of breccias (Fig. 23). The Monte Mugetto fault directly controlled facies distribution and depositional geometries of the sediments deposited in the basin. The bedding in the redeposited sediments is not parallel to that of the Dolomia Principale in the footwall and forms a roll-over anticline. Angular unconformities of local importance are found in the redeposited succession (for instance in the Monti di Gnin area, Fig. 23).

Similarly to what seen at the Monte Pidaggia fault, deformation in the footwall of the Monte Mugetto fault is limited to some irregularly oriented extensional fissures filled with dolomitic sediment.

The existence of a further normal fault is also suggested by the peculiar geometry of the base of the northern thrust sheet east of Val Sanagra (Figs. 6 and 19). The base of the thrust sheet is here oblique with respect to bedding, climbing upsection towards the west. This geometry could be a consequence of the reactivation of a Mesozoic normal fault.

Syn-sedimentary extensional faults in the *southern thrust sheet* are found in Valsolda and north of the Sasso di S. Martino.

In Valsolda a set of now roughly NE-SW striking normal faults is observed (Figs. 6 and 23). The faults are only less than a km-spaced and have vertical displacements of some hundred meters. The Riva di Solto Shale was deposited only in the grabens and is not found on the horsts (e.g. on I Pizzoni and on the Costa dei Ciappei). The base of the normal faults is cut by the basal thrust of the tectonic unit. The faults do not displace the upper Zu Limestone and the Conchodon Dolomite.

The close-spacing and the localization near the Lugano normal fault of the Valsolda faults could indicate that they are blind normal faults formed during the movement of the hangingwall of a listric normal fault (KELLER, in press). With this configuration, the upper part of the hangingwall is subjected to an extension larger than that of the lower part. Faults therefore form which die out downwards.

Another Triassic normal fault can be well documented in the Sasso di S. Martino area in the eastern part of the southern tectonic unit (Figs. 23 and 25). The fault separates the fine-grained carbonates of the Zorzino Formation to the NW from the platform dolomites of the Sasso di S. Martino area to the SE. The Sasso di S. Martino fault, in spite of its significant vertical displace-

ment (around 1 km) had a very limited morphological expression as suggested by the limited amount of coarse-grained platform derived sediments found in the adjacent basin (chapter 2.3.2). Similarly to the Valsolda faults, the Sasso di S. Martino fault dies out during the middle Rhaetian and does not affect the late Rhaetian Conchodon Dolomite.

4.2.2 *Normal faulting in the sedimentary units: paleogeography and paleotectonic evolution*

To reconstruct the paleogeographic and paleotectonic evolution of extension faulting, the effects of Alpine thrusting and folding must be removed. The thrusting is compensated knowing that the transport direction was towards the north (chapter 3.2). The distance between the thrust sheets in a N-S paleodirection is unknown. The effects of folding are filtered out through geometrical rotations bringing the sedimentary units into their depositional position.

This is done rotating the sediments around a subhorizontal E-W striking axis in a anticlockwise direction looking eastwards. This because the steepening of the sediments is related to folding with E-W striking axes (chapter 3.3). The beds are rotated by an amount large enough to let them loose any northward or southward dip. Since bedding generally strikes WNW-ESE the strata often keep a westward dip after rotation. This is corrected in order to bring the top of the Dolomia Principale (youngest horizon preserved in all tectonic units) to horizontal. The same rotations are then applied to the Mesozoic extensional faults.

By these operations, three E-W striking vertical profiles through the footwall of the duplexes and through the two thrust sheets themselves can be constructed (Fig. 19); a paleotectonic map showing the distribution of faults in the Generoso basin at the end of the Norian is also obtained (Fig. 26).

From the sections of figure 19 it is apparent that substantial extension occurred during the Late Triassic not only across the Lugano fault, but also within its hangingwall, *i.e.* the Generoso basin. Since the sedimentary units are detached from the original substratum, I can only speculate how this extension was accommodated at deeper levels. In some cases (Monte Pidaggia fault, Monte Mugetto fault) bedding in the footwall and in the hangingwall are not parallel implying a flattening of the fault at depth. Indeed the Raibl Beds may have acted as a decollement horizon for a number of faults but it is not known how this

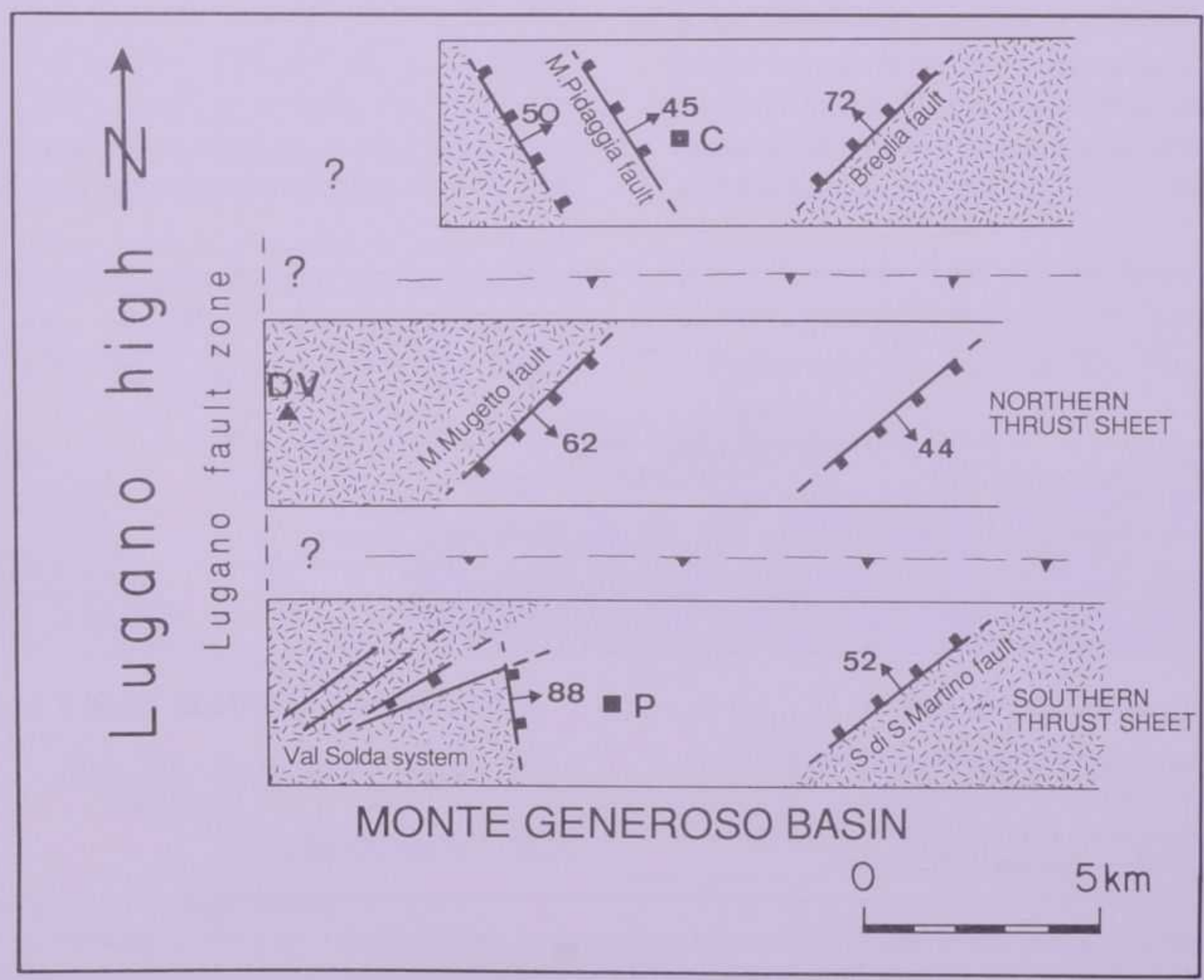

FIG. 26 - Paleotectonic map for the late Norian times after removal of Alpine steepening and thrusting (see text for modalities of reconstruction). Scale is indicative only for the E-W direction. The position of the future Alpine thrusts is also given. Random dashes = Dolomia Principale, white = Zorzino Formation. Figures indicate dip of syn-sedimentary normal faults. C = Carlazzo; DV = Denti della Vecchia; P = Porlezza.

extension was transferred to deeper crustal levels.

The paleotectonic map (Fig. 26) shows the distribution of normal faults in the area. Dip values of the fault surfaces resulting from the geometrical restoration described above are also given. The general map-view pattern is quite irregular. NE-SW strikes seem, however, to be predominant among the faults. This trend could be indicative of sinistral transtensional regime at the regional scale. At the surface the faults dip more than 45°: they are not low-angle faults (or very low-angle as appropriately emphasized by JACKSON, 1987).

Some of the Late Triassic faults caused the formation of morphological basins. These are concentrated in a central zone and were probably interconnected resulting in a roughly N-S oriented trough. The basins opened towards the north and were therefore not limited to the present outcropping area.

Most of the normal faults within the Generoso basin died out before the end of the Rhaetian. Where the upper termination is visible, these faults do not cut the upper part of the Zu Limestone and the Conchodon Dolomite (Figs. 6 and 19). The often observed lateral continuity of the Conchodon Dolomite is usually interpreted as due to a quiescence phase separating two rifting pulses (JADOUL, 1985; GAETANI *et al.*, 1986). I see in the study area no reason to confirm this interpretation. The differences in facies and thicknesses of middle Rhaetian to Lower Liassic sediments on the two sides of the Lugano fault require persisting normal faulting. The extinction of the Triassic faults within the Generoso basin is rather due to a concentration of strain along the Lugano normal fault during ongoing rifting.

4.3 DEFORMATION ALONG THE LUGANO NORMAL FAULT AS DERIVED FROM THE FAULT ROCKS OF THE MONTE GRONA/VAL GRANDE FAULT ZONE

4.3.1 *Introduction and main results*

A stripe of fault rocks is found along the Monte Grona/Val Grande fault zone varying in thickness from few tens of meters in the Val Colla - Val Rezzo region up to several hundred meters east of Val Sanagra (Fig. 27). The fault rocks have been studied with a double aim: a) to check the compatibility with our general model and b) to constrain the geometry of the fault before Alpine shortening. I will first describe the fault rocks outcropping along the fault zone (chapter 4.3.2) and then present a series of deformation profiles across the fault zone from Lake Como to Val Colla (chapter 4.3.3).

Fault rocks are found both north and south of the Monte Grona/Val Grande line, that is both in the footwall and in the hangingwall respectively of the Lugano normal fault (Fig. 27). The fault zone *N of the Monte Grona/Val Grande line* (footwall of the Lugano normal fault) is characterized by a variation of fault rocks both moving from west to east as well as from north to south. Only brittle deformation overprints the Variscan

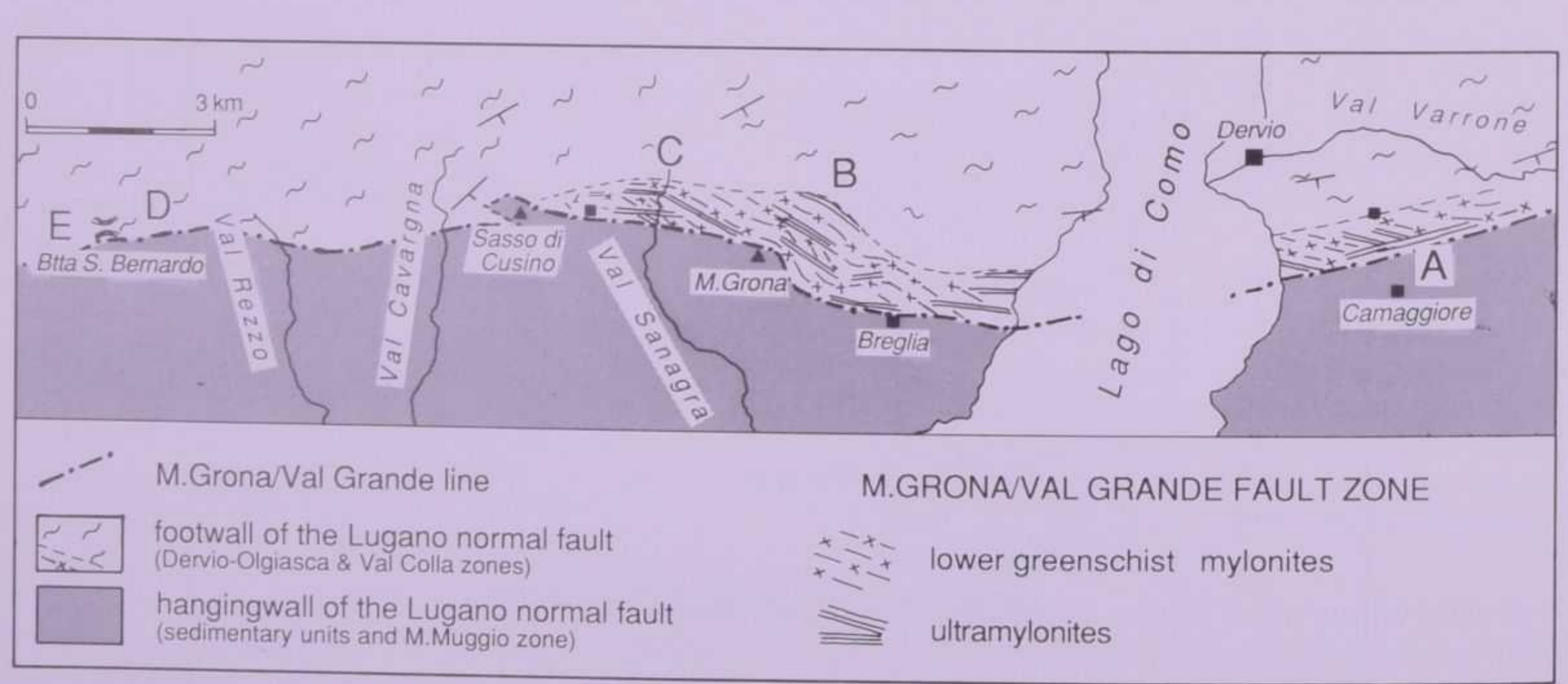

FIG. 27 - The Monte Grona/Val Grande fault zone; distribution of various fault rocks. Note geometric relationships between the Monte Grona/Val Grande line, expression of the last tectonic movements, and the foliation in the fault rocks: the line is parallel to the ultramylonitic foliation in Val Sanagra and from Breglia eastward. Only in the Monte Grona area the Monte Grona line is parallel to foliation of lower greenschist mylonites and oblique with respect to the ultramylonites (see text). Profiles across the fault zone described in text: A = Camaggiore; B = Monte Grona ridge; C = Val Sanagra; D = upper Val Rezzo; E = Piancabella.

rocks W of Val Cavargna (Fig. 27). Mylonites which become thicker and better developed towards the east are found from Val Sanagra eastwards. These mylonites formed under lower greenschist metamorphic conditions. This pattern is obviously compatible with the Monte Grona/Val Grande fault zone cutting across deeper crustal levels moving from west to east in present day coordinates. Horizons of ultramylonites are found in the lower greenschist mylonites which become more significant approaching the Monte Grona/Val Grande line. Cataclastic overprint is ubiquitous immediately north of the line.

Only brittle deformation is observed *south of the Monte Grona/Val Grande line* (hangingwall of the Lugano normal fault).

Deformation profiles across the Monte Grona/Val Grande fault zone are thus strongly asymmetric: mylonites are found north of the line and become progressively overprinted by lower temperature deformation down to cataclasis further to the south. South of the line only brittle deformation is observed. This pattern is typical for crustal normal faults with a cold hangingwall being juxtaposed to a warmer footwall (Fig. 28). The juxtaposition causes the footwall to cool.

Similar deformation profiles have been recognized in the last years from the metamorphic core complexes of the Basin and Range Province (Lister and Davis, 1989; and references therein) and from other regions like the western Swiss Alps (Simplon line, Mancktelow, 1985). In all these cases deformation has been related to major crustal normal faults.

The asymmetry between the fault rocks north of the Monte Grona/Val Grande line which were deformed under elevated temperatures (lower greenschist) and those south of the line which underwent only cold, brittle deformation goes hand in hand with the asymmetry in the radiometric ages briefly presented in chapter 2.2 (see also chapter 4.3.5): the northern block (footwall) with Late Permian to Early Jurassic ages and the southern block (hangingwall) with the Variscan paragenesis and radiometric inprint being preserved.

The kinematics of movements across the Monte Grona/Val Grande fault zone is also well constrained. Stretching lineations associated to the S-SW dipping mylonites and ultramylonites typically trend E-W to WNW-ESE (Fig. 29 a and b). Kinematic indicators (shear bands, quartz preferred orientations, mica fishes; review in Simpson and Schmid, 1983) indicate a movement of the southern block towards the east in present day coordinates. Both stretching lineations and kinematic indicators are consistent for the mylonites, for the ultramylonites and, in the few discernible cases, for the cataclasites. This implies that in spite of changing deformational mechanisms, the kinematic conditions remained substantially the same during deformation. Restoring the Monte Grona/Val Grande fault zone to its pre-Alpine position, the stretching lineations assume a dip slip geometry with eastward dips and the sinistral movement becomes a top-to-the east sense of movement (Fig. 3). These are the geometric features of the Lugano normal fault and demonstrate the plausibility of the correlation between the Monte Grona/Val Grande fault zone and the Lugano normal fault.

4.3.2 *The fault rocks of the Monte Grona/Val Grande fault zone: metamorphic features*

Fault-rocks are rocks whose dominant micro-

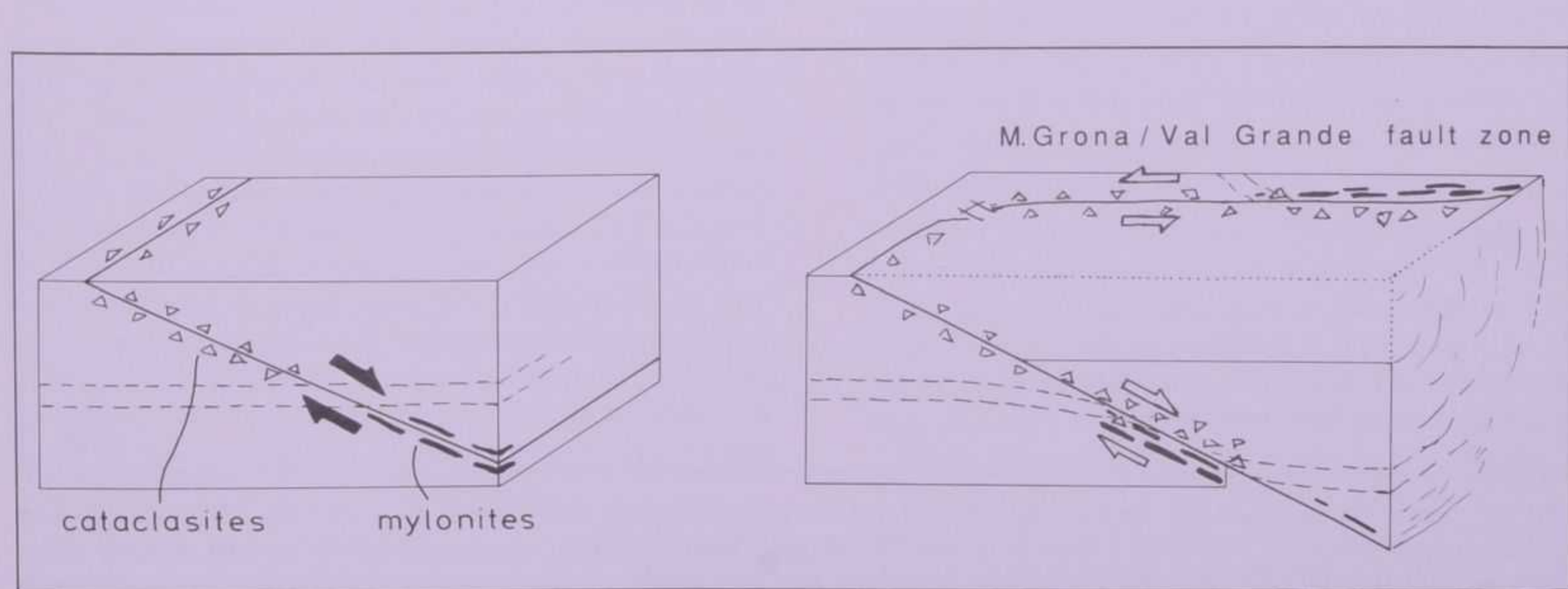

Fig. 28 - Cartoon illustrating the distribution of fault rocks along a crustal normal fault like the Monte Grona/Val Grande fault zone. The cataclasites formed on the hangingwall in the upper part of the fault are transported downwards and juxtaposed to the mylonites of the footwall. The juxtaposition of the two blocks causes a distorsion of the isotherms in the footwall: dynamic recrystallization is replaced by cataclasis.

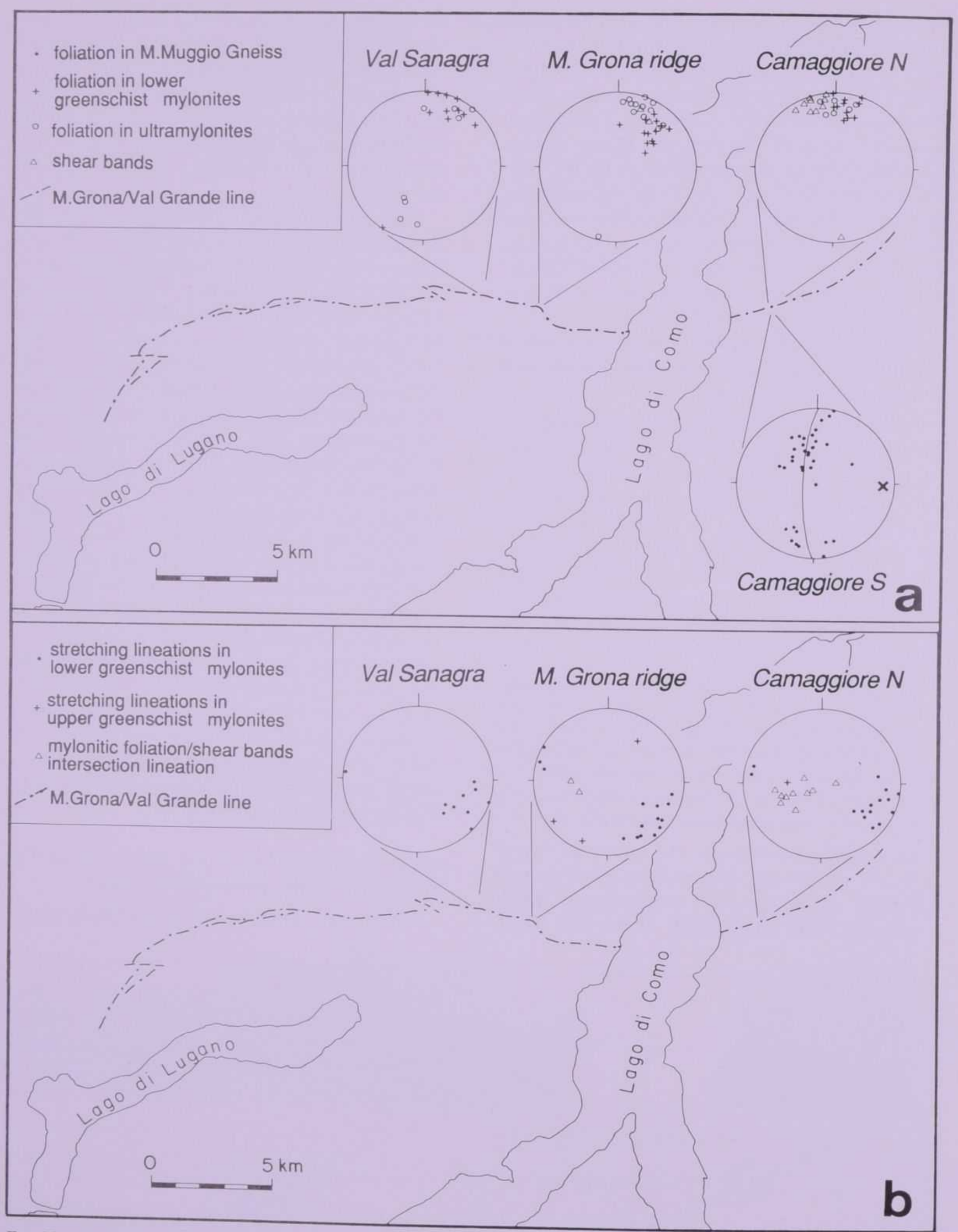

FIG. 29 - Summary of structural data along the Monte Grona/Val Grande fault zone. a = attitude of mylonitic and ultramylonitic foliations (Variscan foliation in the M. Muggio zone). Note in Camaggiore N and M. Grona ridge sections how the field of ultramylonites poles overlaps the two fields of lower greenschist mylonites and shear bands. b = stretching lineations and shear band/foliation intersections. Note how the (unfortunately scarce) stretching lineations in Variscan upper greenschist mylonites deviate from the general trend.

structure and texture ([1]) formed during deformation within a fault zone. Different types of fault-rocks occur along the Monte Grona/Val Grande fault zone recording a prolonged deformational history under changing physical conditions. In this work they are grouped into mylonites, ultramylonites and cataclastic rocks on the base of microstructure and texture. These groups include rocks with similar characteristica and are mainly thought as mapping units. The description I give does not pretend to give exhaustive information about the processes involved in the genesis of the rocks. For this purpose I will refer to the abundant literature.

Low-temperature mylonites

The mylonites belonging to this group formed at temperatures below the biotite isograde. Biotite is absent or unstable, being typically replaced by chlorite. Quartz microstructure indicates dynamic recrystallization by polygonization (rotation recrystallization of Drury and Urai, 1990).

Low-T mylonites make up most of the fault-rocks north of the eastern part of the Monte Grona/Val Grande line (Fig. 27). These rocks best allow for structural and metamorphic investigations.

Lower greenschist mylonites consist of well foliated gneisses with common cm- to dm-thick quartz veins (Plate V, Fig. 1). Up to cm-large feldspathic augens are widespread (Plate V, Fig. 2). The paragenesis is made up by quartz, K-feldspar, plagioclase and chlorite. Garnet is common and usually partly replaced by chlorite. Relic biotite crystals are sometimes found partly replaced by chlorite (usually penninite). Zircon, epidote and apatite are found as accessories.

Quartz grains always show a variable degree of dynamic recrystallization. The effective recrystallization mechanism is subgrain-rotation leading to polygonization (Plate V, Figs. 4 to 6) (Poirer and Nicolas, 1975; Urai *et al.*, 1985; Drury and Urai, 1990). The newly formed strain-free grains typically have dimensions of 20-30 microns. Polygonization is often not complete so that core-mantle structures are observed with variable proportions of new grains vs. parental grains. Parental grains are generally stretched and elongated parallel to the foliation. Boundaries of new grains are either parallel to or make an acute angle with the foliation (Plate V, Fig. 5). A good crystallographic preferred orientation is typically developed.

Feldspars are the second most important phase. They are made up of large, up to 1 cm, crystals. They can be poikiloblastic with inclusions of biotite, garnet etc. Plagioclase shows exsolution (Plate VI, Fig. 1) and (strain-controlled?) myrmekitic structures (Plate VI, Fig. 2). K-feldspar is often microcline twinned (Plate VI, Fig. 8). Neither plagioclase nor K-feldspars show sign of dynamic recrystallization. They are brittely deformed (Plate V, Fig. 3). The plagioclase can be strongly sericitized and is often replaced by albite + chlorite + epidote aggregates.

Muscovite is abundant both as clasts and as small crystals in the groundmass. Biotite is present only as a relic and is generally replaced by chlorite.

Chlorite, both as penninite and as clinochlore, is common. It can form large crystals replacing biotite. It can also be found in smaller and less regular crystals growing along the shear bands and in shadow zones of garnets, feldspars etc. Chlorite is therefore interpreted to be synkinematic.

The mylonites are often crossed by long and well defined shear zones made up of fine-grained chlorite, sericite and possibly epidote. These horizons develop starting from large plagioclases showing advanced alteration (Plate VI, Figs. 4 and 5). The horizons are parallel to the main foliation or to the shear bands and represent the initial stage of thicker ultramylonitic horizons (see below).

Quartz microstructures similar to those described above have been reported from several fault zones (Sibson, 1977; Simpson, 1985; Mancktelow, 1985; Hyndman and Myers, 1988; etc). All authors agree in interpreting them as due to deformation occurring under lower greenschist conditions (Voll, 1976). On the base of the quartz microstructure and of the paragenesis, I assume that the low-temperature mylonites formed at around 300°C temperature.

Ultramylonites

The rocks grouped here are finely laminated, mostly very fine grained, grey to dark schists (Plate VII, Fig. 1). They are referred to as ultramylonites on the base of their high matrix vs. clast content and because of their laminated aspect (Sibson, 1977; White *et al.*, 1982).

Under the microscope the ultramylonites consist of a fine-grained, almost opaque matrix with clasts and thin, sporadic, mylonitic quartz-ribbons (Plate VII, Figs. 2 and 4). The groundmass is mainly made up of up to 10-20 microns large feldspar and quartz crystals and by needle-shaped minerals, probably zoisite and chlorite (Plate VII, Fig. 5). Sericite is also abundant. The groundmass typically shows a good crystallographic preferred orientation. SEM images in the backscatter mode show the microstructure of the rock with the needle-like minerals elongated parallel to the foliation (Plate VII, Fig. 3).

The clasts consist of sub-rounded, often poikiloblastic feldspars and of muscovite. No biotite or garnet is

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[1] The term microstructure refers to the shape of the mineral grains and their geometric relationships with matrix etc; texture refers to the orientation of the optical elements of the mineral.

found; these phases showed already in the lower greenschist mylonites signs of chemical instability. Feldspars closely recall the clasts previously described in the mylonites showing similar features. They are however definitely smaller (typically about 0.1 mm). Plagioclases are generally replaced by chlorite, sericite, epidote and albite (Plate VII, Fig. 6).

Discontinuous, tenths of mm thick, mylonitic quartz bands are found parallel to the foliation (Plate VII, Fig. 2). Their microstructures and textures are identical to those found in the mylonites. With progressive ultramylonitization, the mylonitic quartz bands become shorter and more irregular.

Lower greenschist mylonites and ultramylonites are only end-members of a continuous succession. The mylonites are often crossed by very thin shear zones along which chlorite, epidote and sericite are found (Plate VI, Figs. 4 and 5). The shear zones become thicker and better developed towards the Monte Grona/Val Grande line and finally form the ultramylonitic horizons. The localization of the small-scale shear zones is often controlled by aggregates of albite, epidote, sericite and chlorite. These aggregates are thought to be derived from the break-down of plagioclase of intermediate composition under low temperature (lower greenschist and cooler) and in the presence of fluids (STÜNITZ, 1989). These aggregates are easily deformed. Proceeding deformation in the presence of abundant fluids causes further break-down of plagioclase which leads to a decreasing size of plagioclase crystals (Plate VI, Fig. 6). At the same time the proportion of matrix vs. clasts increases (Plate VII, Figs. 2 and 4), and the phases which were unstable in the lower greenschist mylonites like biotite and garnet, completely disappear. At the end of this process, a typical ultramylonite is obtained. The interpretation is that changes in physical conditions during deformation caused the progressive disactivation of dynamic recrystallization of quartz as the effective deformational mechanism and its replacement by the deformation of the chl+ser+epidote, plagioclase-derived aggregates. In fact, STÜNITZ (1989) who exhaustively investigated this topic, also describes cases where plagioclase-derived aggregates are more ductile than quartz-controlled dynamic recrystallization. I interpret this transition from quartz-controlled to plagioclase-aggregate-controlled deformation to be a consequence of a syn-tectonic decrease in temperatures.

Cataclasites

Cataclasites are rocks in which "*crystal structure remains undistorted, but grains or groups of grains become cracked and the fragments may exhibit frictional sliding with respect to one-another. The process necessarily involves dilatancy and is therefore pressure sensitive*" (RUTTER, 1986).

Cataclasites are found all along the Monte Grona/Val Grande fault zone. West of Val Cavargna they are the only fault rocks present north of the Monte Grona/Val Grande line. East of the valley they are associated with the lower greenschist mylonites and ultramylonites and tend to be concentrated close to the Monte Grona/Val Grande line. The clasts of the cataclasites along this segment of the fault zone are always derived from mylonites and/or ultramylonites. The mylonitic and ultramylonitic foliation is progressively disrupted (Plate VII, Fig. 1) and clasts mainly consisting of shreds of mylonitic quartz ribbons end up floating in a fine-grained, isotropic, dark-coloured matrix. The matrix is mostly identical with that of the ultramylonites, but pseudotachylite-derived material might also be important.

South of the Monte Grona/Val Grande line cataclasites are the only fault rocks found. West of Breglia (Fig. 27) their clasts are almost exclusively derived from the sedimentary formations. The cataclasites of the southern block between Breglia and Lake Como are made up mainly of Gneiss Chiari preserving their Variscan micro-structure. Cataclasites east of Lake Como are made up of Monte Muggio Gneiss but are not very thick in spite of the widespread fracturing.

Pseudotachylites

Pseudotachylites are relatively common along the fault zone. They have been found in the crystalline rocks immediately south of the Monte Grona/Val Grande line. They are also found intruding mylonites and ultramylonites immediately north of the line (Plate VII, Fig. 2). Pseudotachylites were probably more common than presently observed and have been successively overprinted by mylonitization and/or cataclasis. In fact fine-grained pseudotachylite-derived material could form part of the fine-grained matrix of ultramylonites and cataclasites.

Other fault rocks associated with, but genetically not belonging to the Monte Grona/Val Grande fault zone

High-T mylonites. In various localites (Camaggiore, Monte Grona, Val Sanagra, Val Rezzo, etc) mylonites are found which show stable biotite. They thus formed under upper greenschist conditions. These mylonites are described here because they are found associated with the fault zone, but

they were not formed during Mesozoic deformation along the Monte Grona/Val Grande line. In fact, their high-grade paragenesis and their kinematic features (stretching lineations and sense of shear) are not compatible with the general deformational picture of the Monte Grona/Val Grande fault zone.

The high-T mylonites form well-banded gneisses (Plate VIII, Fig. 3). The paragenesis typically consists of quartz, feldspar, muscovite, biotite, garnet, zoisite and accessories like zircon, apatite etc. In thin section the foliation is seen to be controlled by the alignment of small mica flakes and by elongated quartz crystals. In quartz-rich lithologies, quartz grains are typically sub-rectangular with the short sides perpendicular to the micas marking the foliation. Grains are usually 0.3 mm long and 0.05 mm high (Plate VIII, Fig. 4). The dimensions are, however, variable and are partly controlled by the abundance of other minerals, like micas. Quartz grains show a fairly good crystallographic preferred orientation indicating that the rock is a mylonite. The microstructure of the rock is partially annealed (especially in quartz-rich layers). The plagioclase can be post-kinematically partially substituted by clinozoisite+sericite±albite. Some quartz and feldspar grains are larger and overgrow aligned mica flakes (Plate VIII, Fig. 5). These crystal could be related to fast grain boundary migration (with no influence by the second phase mineral) (Drury and Urai, 1990) or to static recrystallization after the cessation of deformation.

"*Quartzite*". In some localities (Camaggiore, Val Sanagra), enigmatic bodies of light-coloured, massive rocks, are found which lie always immediately north of the Monte Grona/Val Grande line.

The rock is crossed by numerous veins (Plate VIII, Fig. 6) and a breccia-like aspect can result. In thin section the rock consists of a fine-grained aggregate of not always recognizable, light-coloured minerals (quartz and feldspars) and widespread tiny crystals of chlorite, sericite and epidote. Phantoms of large feldspars are sometimes recognized under parallel nichols. None of the otherwise common phases like muscovite, biotite or garnet, is observed. The texture and microstructure of the rock are completely isotropic. The rock is typically cross-cut by numerous, undeformed vein systems: the veins are filled with quartz, chlorite and more rarely and only west of Lake Como, by calcite.

El Tahlawi (1965) mapped this rock as a quartzite. This attribution is obviously incorrect. I assume that these rocks are derived from an almost complete fluid-controlled alteration of an originally coarse-grained quartzo-feldspathic rock. The problem is that no such rocks are found in the basement close to the Monte Grona/Val Grande fault zone. The only comparable rocks are the Triassic pegmatites of the Piona area which are made up of up to cm-sized quartz, plagioclase and K-feldspar with subordinate muscovite. Other minerals like biotite, garnet etc. are missing (El Tahlawi, 1965 and references therein). I suggest that similar pegmatites were intruded in the Monte Grona/Val Grande fault zone during the last stages of deformation and were then completely altered by abundant fluid circulation. This correlation represents a further similarity between the Monte Grona/Val Grande and the Musso/Piona fault zones (chapter 3.3).

4.3.3 *Deformation profiles across the Monte Grona/Val Grande fault zone: metamorphism and structures*

Camaggiore section

The Camaggiore area, east of Lake Como (Fig. 27), shows the best outcrops of the fault zone. Rocks of the Variscan basement are found until ca. 900 m north of the Val Grande line (Fig. 30). The transition to the fault rocks of the Monte Grona/Val Grande fault zone is not exposed. Lower greenschist mylonites prevail in the northern part of the fault zone (Fig. 31). The mylonitic foliation generally strikes WNW-ESE (Fig. 32 A). Stretching lineations on lower greenschist mylonites gently dip to the ESE (Fig. 33 A). The mylonites typically show sinistral shear bands i.e. a movement of the southern block towards the east in present day coordinates. During mylonitization small intrafolial folds formed with a geometry compatible with that of the general movement. Variscan upper greenschist mylonites are found at different levels of the mylonitic succession. Horizons of ultramylonites become more frequent and better developed near the Val Grande line. The ultramylonites can be parallel to the mylonitic foliation or parallel to the shear bands (Figs. 31 and 32 C). The ultramylonites again show gently ESE dipping stretching lineations and sinistral shear bands. The general picture is therefore that of a southern block moving to the east in present day coordinates. Cataclasites are very widespread immediately north of the Val Grande line. They often consist of mylonitic clasts floating in a fine-grained dark matrix made up of ultramylonites and, possibly, pseudotachylites.

The Val Grande line, expression of the last movements along the fault zone, is oblique with respect to the lower greenschist foliation and has the same attitude as the shear bands.

The Monte Muggio Gneiss south of the Val Grande line is pervasively fractured for at least 2 km south of the line. However, only relatively thin horizons of cataclastic breccias are formed. The brittle deformation overprints directly the

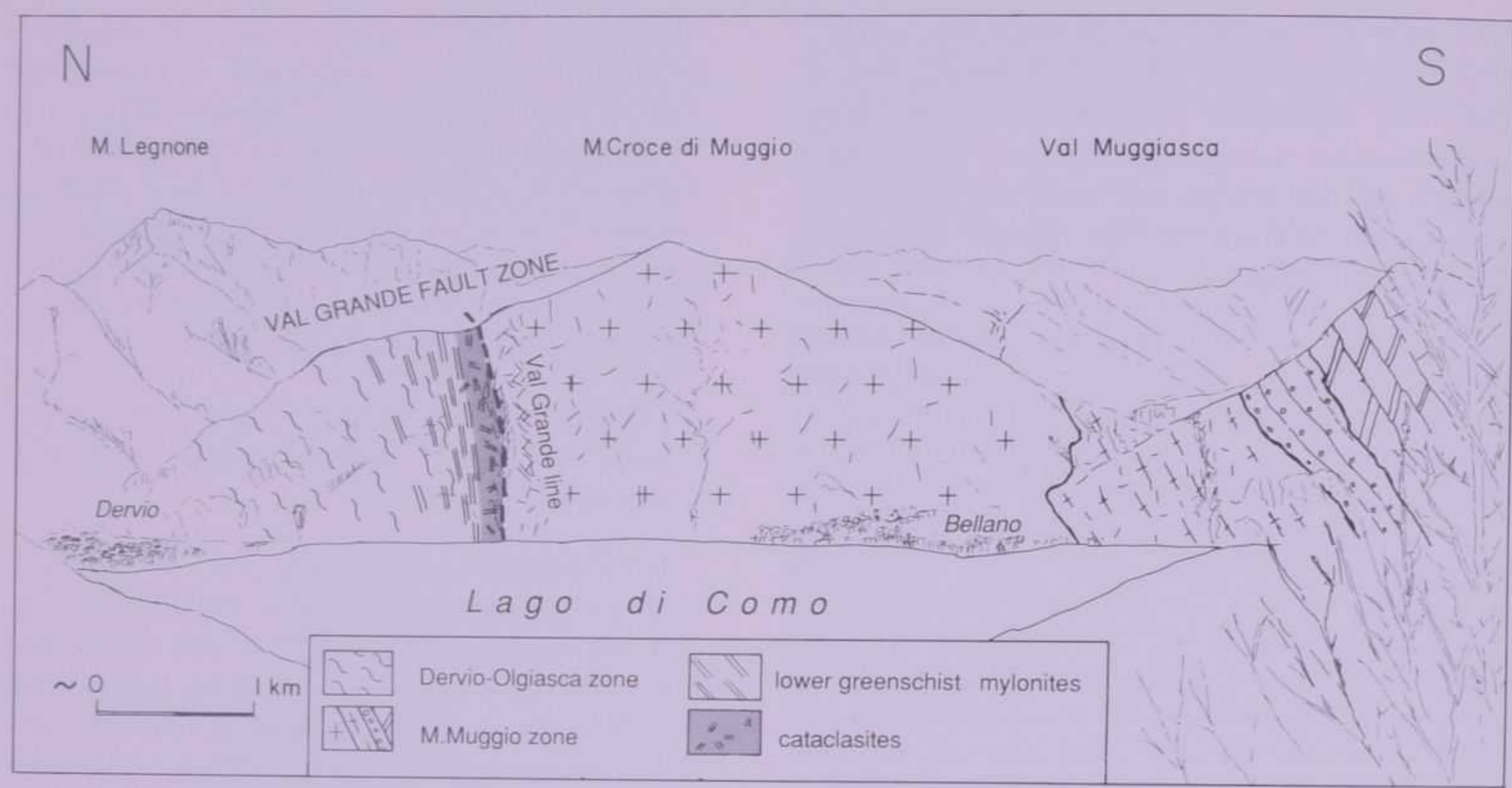

FIG. 30 - Panoramic sketch of the Alpine steepened formations and structures underlying the M. Generoso basin along the eastern shore of Lake Como. The footwall of the Mesozoic Lugano normal fault is represented here by the Dervio-Olgiasca zone (left) which becomes mylonitized and then fractured nearing the fault (Val Grande line). To the right of the fault, the hangingwall of the Lugano normal fault is made up of M. Muggio Gneiss and Gneiss Chiaro which is stratigraphically overlain by lower Permian to Middle Triassic sediments.

Variscan microstructure and practically no mylonitization is observed.

The deformation profile is on the whole strongly asymmetric with high temperature deformation present only in the northern block and a syn-tectonic thermal gradient indicating decreasing temperatures towards the south.

Lithologies

The rocks north of the Val Grande fault zone consist of the staurolite-bearing two-micas gneisses of the southern Dervio-Olgiasca zone and are found in the

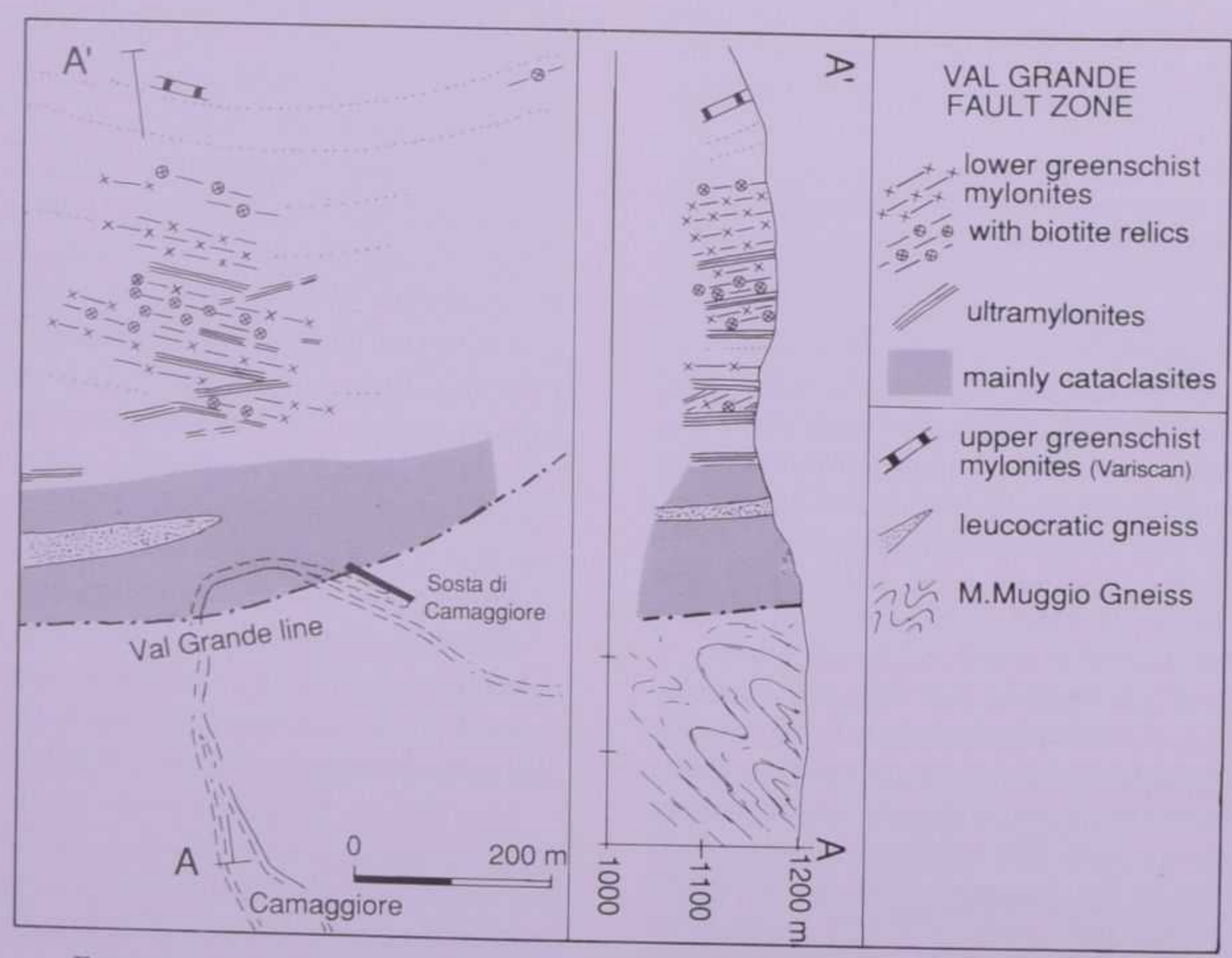

FIG. 31 - Lithologies in the Camaggiore section across the Val Grande fault zone.

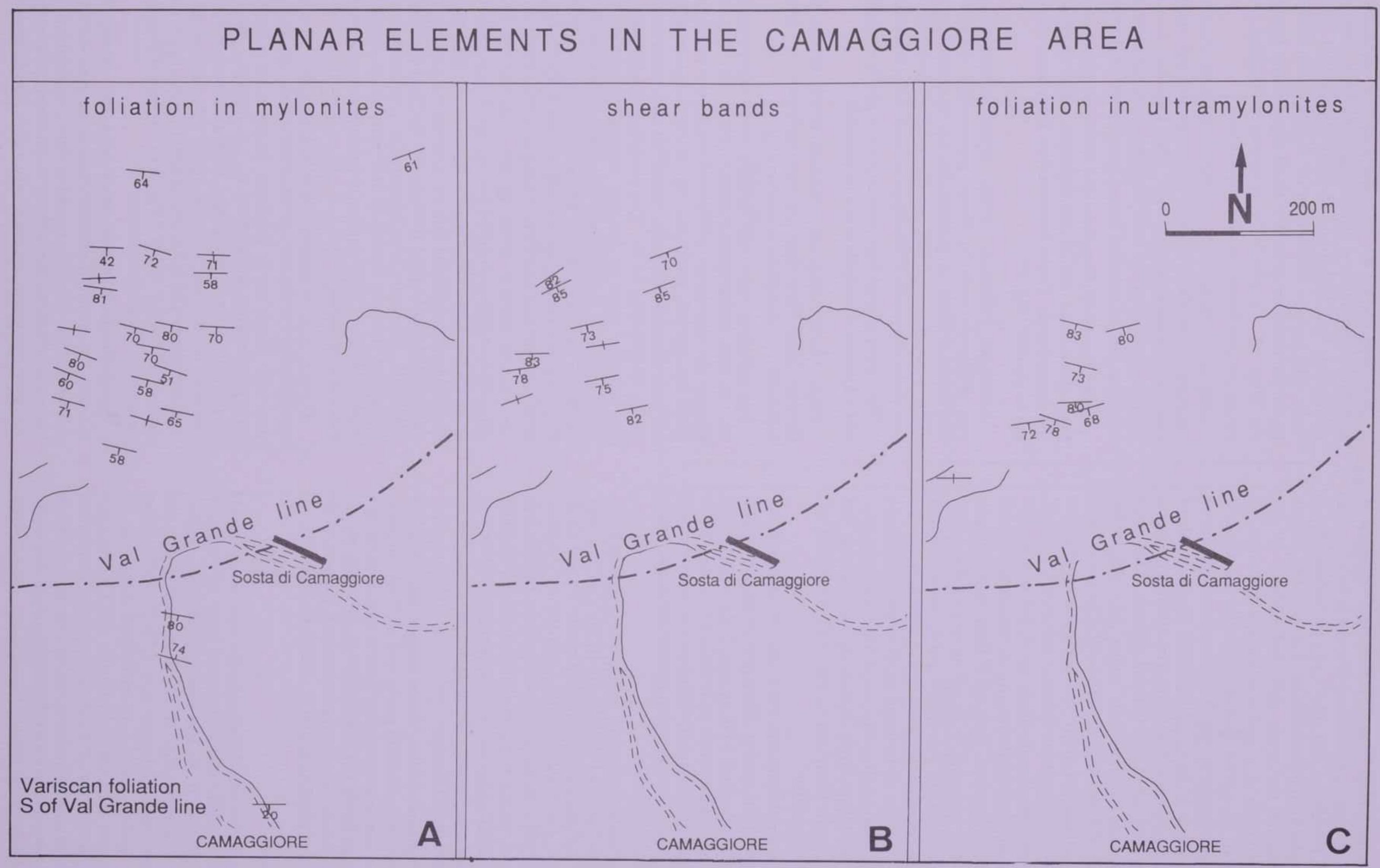

FIG. 32 - Attitude of mylonitic (A), ultramylonitic (C) foliation and of shear bands (B) in the Camaggiore area.

area around Dervio (Fig. 27). Moving southwards, a horizon of upper greenschist mylonites is found in the Alpe di Pratolungo area which can be followed for a few hundred meters to the east (Fig. 31). The transitions towards the north with the gneisses of the Dervio-Olgiasca zone and that towards the south to the lower greenschist mylonites of the Val Grande fault zone are not exposed. The first mylonites related to the fault zone are found some 500 meters north of the Val Grande line. They form a ca. 100 m thick band and are made up of well foliated gneisses. Mica- and feldspars-rich layers show an "augen" fabric due to the presence of large feldspars. The same layers show well developed sinistral shear bands containing chlorite. Foliation-parallel quartz veins are pervasively dynamically recrystallized with nicely developed core-mantle structures. Some thin horizons of ultramylonites are sporadically observed. A ca. 50 m thick band of upper greenschist mylonites is found further to the south: the mylonites show widespread, partly unstable biotite and relic ribbons of sub-rectangular quartz grains. Beginning from ca. 250 meters north of the Val Grande line the ultramylonites become thicker and more frequent: the morphologic depression north of the Val Grande line is partly due to the easily weathered ultramylonitic schists.

Up to dm-thick quartz veins are often found along this part of the section. They can be parallel to the mylonitic foliation or oblique to it. Intrusion occurred before, during and after mylonitization: some of the foliation-parallel dykes show dynamically recrystallized quartz grains, in others quartz grains show only limited deformation bands.

Cataclasis disrupting the mylonitic and ultramylonitic foliation becomes very strong in the southern part of the morphological depression. Clasts of mylonites and ultramylonites end up floating in a dark cataclastic matrix. The leucocratic rocks mapped by El Tahlawi (1965) as quartzites are found here (description and discussion in section 4.3.2). They form a E-W elongated, some meters-thick tabular body running from Camaggiore to Lake Como.

The gneisses of the Monte Muggio zone south of the Val Grande line still preserve their Variscan paragenesis and microstructure. Quartz grains show deformation bands and some, very limited dynamic recrystallization. Late circulation of fluids caused the alteration of plagioclase, garnet, staurolite and biotite.

Structures

Planar elements. Foliation in the mylonites generally strikes WNW-ESE with rather steep dips towards the south (this is also the attitude of the upper greenschist mylonites) (Fig. 32 A). From west to east the foliation seems to rotate into a WSW-ENE strike. Shear bands in mylonites strike WSW-ENE (Fig. 32 B). Angular relationships with the mylonitic foliation are representative of a sinistral sense of shear. Well developed ultramylonitic horizons can be parallel to the mylonitic foliation (WNW-ESE) or parallel to the shear bands (WSW-ENE) (Fig. 32 C).

Stretching lineations. Mylonites of the Camaggiore section commonly show stretching lineations. They are particularly well developed in foliation-parallel quartz veins. The general trend is quite regular with WNW-ESE directions and gentle E-ward dips clearly prevailing (Fig. 33 A). Similarly trending stretching lineations are also found on ultramylonites.

Only in one case could a clear stretching lineation associated with the Variscan upper greenschist mylonites be observed (Fig. 33 A). The lineation has an intermediate dip to the west and thus strongly deviates from the general trend of lower greenschist mylonites.

Other lineations found on the foliation plane result from the intersection of the foliation plane itself and the shear bands. These intersection lineations steeply dip to the SW (Fig. 29).

Kinematic indicators. Shear sense indicators are very consistent both in lower greenschist mylonites and ultramylonites: in present day coordinates they indicate a movement of the southern block towards the east along the gently ESE-dipping stretching lineation.

The upper greenschist mylonites are partly annealed and a kinematic analysis was not possible.

Folds. Ductile folds are common north of the Val Grande line. They can be syn-mylonitic or post-mylonitic.

Syn-mylonitic folds are usually well expressed by cm- to dm- thick quartz veins (Plate IX, Fig. 1). Hinges are commonly preserved while the limbs are often sheared out. Fold axes lie on the foliation plane but have variable dips both to the SE and, more rarely, to the W-SW (Fig. 33 B). The position of fold axes on the foliation plane is roughly controlled by the distance from the Val Grande line: syn-mylonitic folds nearer to the line tend to show parallelism between fold axes and stretching lineation. The folds are typically asymmetrical and have a vergence consistent with the general sinistral sense of shear across the fault zone. Quartz grains in folded veins are dynamically recrystallized (Plate IX, Fig. 2). Since axial planes are parallel to the mylonitic foliation and quartz grains elongated parallel to the foliation, the folds must be syn-mylonitic. However, crystallographic preferred orientations of the two limbs of the fold show opposite senses of shear. This implies a preservation of an active stage of folding.

Ductile, post-mylonitic folds are dm- to m-sized folds folding the mylonitic foliation (Plate IX, Fig. 3) and must therefore be younger. Folds have generally long planar limbs and tend to be symmetrical. No axial plane foliation is developed. Fold axes have variable orientations but seem to lie on a sub-vertical E-W striking plane. The folds are generally associated with semi-brittle shear zones (Fig. 34) to which they are genetically related (see below).

Some ductile folds are observed south of the Val Grande line: since no retrograde mylonitization affected these rocks, the folds must be Variscan. Brittle folds are quite common: they are cm-to m-sized, very irregular and sometime develop an axial plane fracture

FIG. 33 - Stretching lineations (A) and fold axes (B) in the Camaggiore section.

cleavage. Fold axes strike often E-W with gentle plunges to the E (Fig. 33 B) and are therefore compatible with N-S Alpine shortening. The vergence of the folds is towards the south between the Val Grande line and Camaggiore.

Semi-brittle shear zones. Meters-sized semi-brittle shear zones are found in the lower greenschist mylonites north of the Val Grande line. The mylonitic foliation is folded into the shear zones which are themselves marked by a thin cataclastic gouge (Fig. 34 and Plate IX, Fig. 4). The shear zones are typically steep. Conjugated systems are often observed. NW-SE, WNW-ESE striking zones are dextral while NE-SW, ENE-WSW ones are sinistral. The sinistral faults are generally better developed than the dextral ones. Stress axes reconstructed from conjugate systems show a horizontal, N-S directed compression and a horizontal, E-W directed extension in present day coordinates.

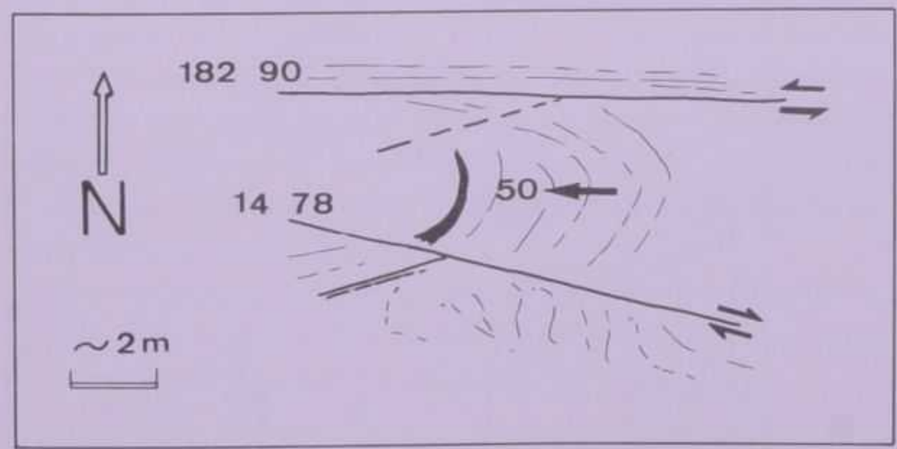

FIG. 34 - Outcrop sketch of semi-brittle shear zones with associated post-mylonitic ductile folds (fold axis indicate by the thick arrow). Note acute angle bisected by extensional axes.

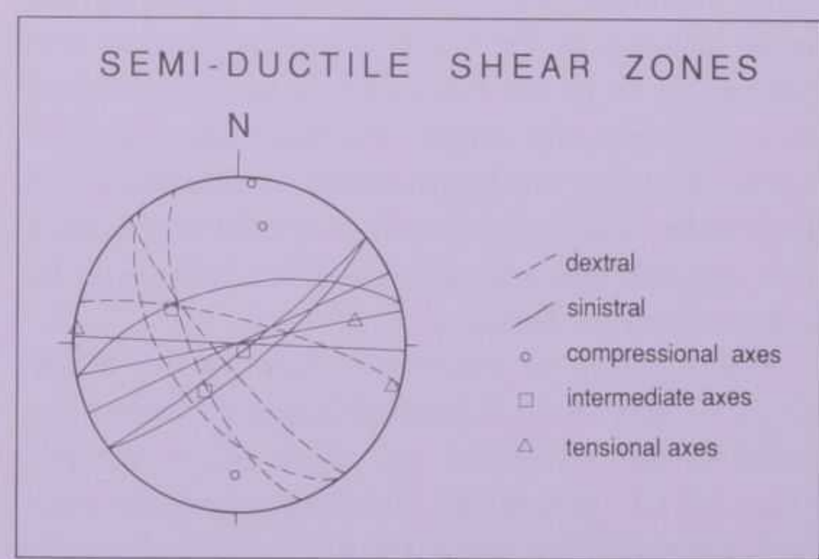

FIG. 35 - Summary of semi-ductile shear zones and of stress axes derived from conjugate sets of semi-ductile shear zones.

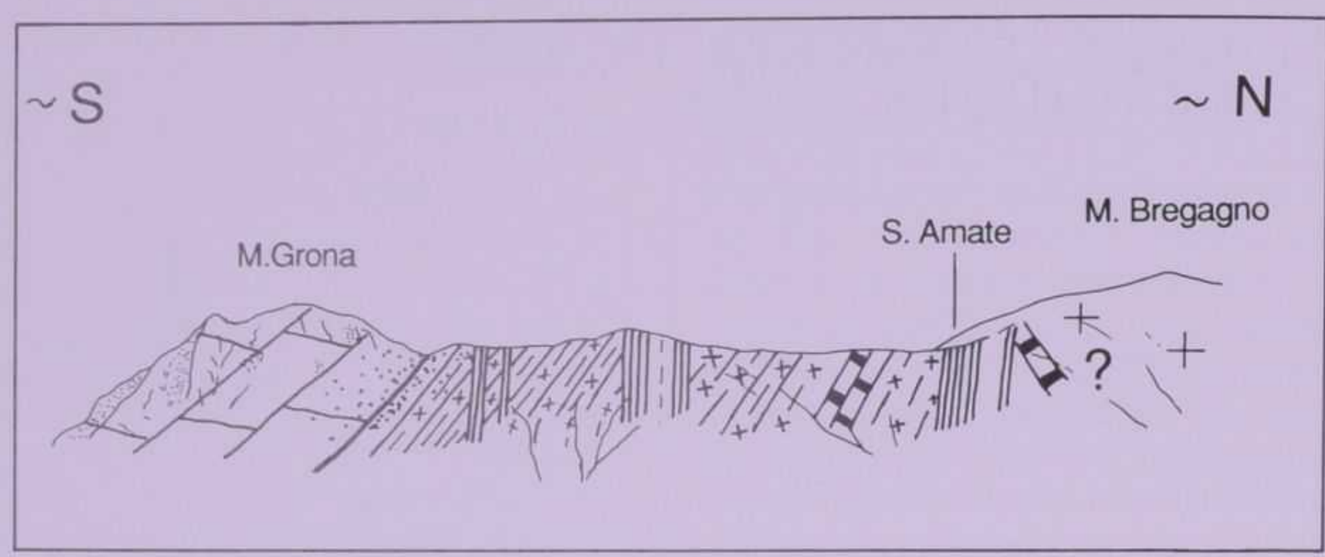

FIG. 36 - Sketch from photograph of the ridge between Monte Grona and Monte Bregagno. Fault rock legend is the same as in Fig. 37.

Conjugate systems show compressional axes bisecting the obtuse angle (Fig. 35). This geometry is typical for ductile shear zones (RAMSAY and HUBER, 1987) and is generally explained with a rotation of the shear zones towards the XY plane (X>Y>Z) possibly associated with a pure shear component of deformation. The rotation of the shear zones requires the removal of material from the two dihedra with decreasing angle: this could be accomplished by the ductile, post-mylonitic folds described above (Fig. 34). Because of the ductility of these shear zones (requiring high temperatures) I assume that they formed before Alpine steepening. Rotating back the shear zones, the compressional axes become sub-vertical and the extensional ones trend E-W. This geometry is obviously not compatible with Alpine N-S shortening but is well in accord with the E-W directed extension which occurred during the Mesozoic and which formed the Monte Grona/Val Grande fault zone.

Monte Grona ridge

A fairly well exposed profile across the Monte Grona/Val Grande fault zone is found NNE of Monte Grona along the ridge leading to Monte Bregagno (Figs. 27 and 36). The fault zone along this profile is ca. 1000 meters thick. Lower greenschist mylonites and ultramylonites make up the large majority of the fault-rocks. Cataclasites are, differently from the Camaggiore section, limited to a relatively thin stripe. Variscan upper greenschist mylonites are found north of and inside the fault zone. The dolomites S of the Monte Grona line are only brittely deformed but with only limited cataclastic breccias.

The foliation of mylonites dips to the S-SW (Figs. 37 and 38 A). Ultramylonites are either parallel to the mylonites or, more often, steeper (Fig. 38 C). Stretching lineations are common both on mylonites and ultramylonites and generally strike NW-SE (Fig. 38 D). Kinematic indicators show a movement of the southern block towards the SE. Fault rocks and kinematic features are thus similar to those observed along the Camaggiore profile.

Differently from Camaggiore, the Monte Grona line is parallel to the mylonitic foliation and oblique with respect to the ultramylonites.

Lithologies

Stabbiello Gneiss with no retrograde deformation is found till ca. 1100 meters north of the Monte Grona line (Fig. 37). South of the Stabbiello Gneiss (Figs. 36 and 37), some 150 m north of the S. Amate chapel, a horizon of Variscan upper greenschist quartzo-

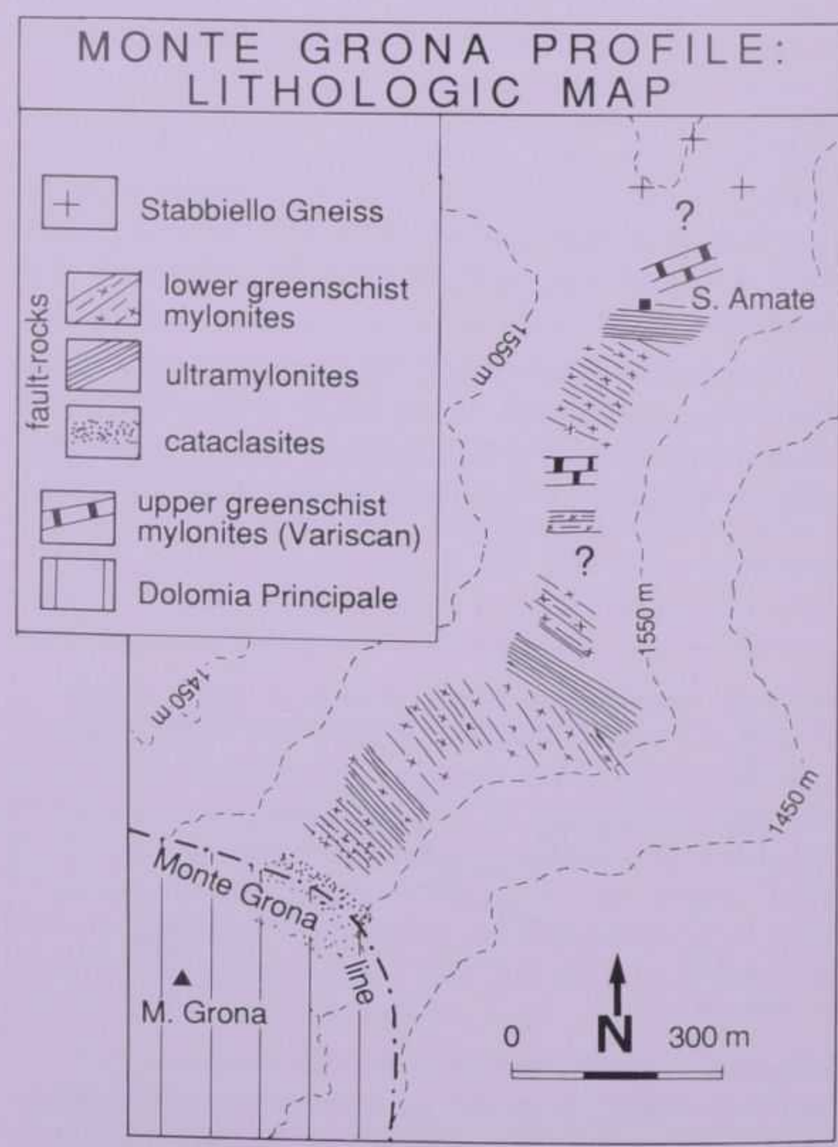

FIG. 37 - Fault rocks along the Monte Grona ridge profile.

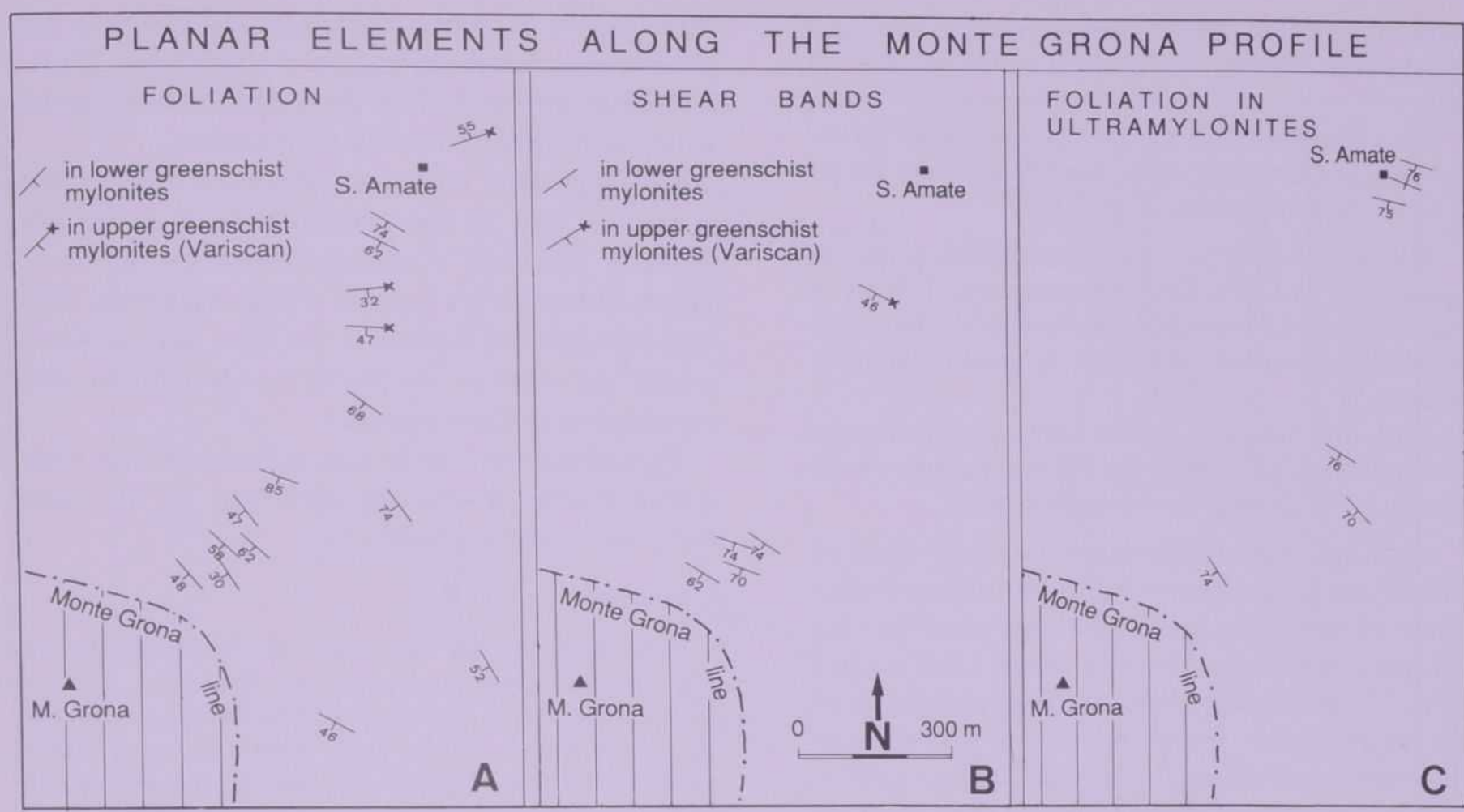

FIG. 38 A, B, C - Planar elements along the Monte Grona ridge section.

mylonites is found. They preserve their high temperature assemblage and microstructure. Neither the transition towards the Stabbiello Gneiss in the north, nor that towards the Monte Grona fault zone in the south are exposed. The first fault-rocks related to the Monte Grona fault zone are found at the S. Amate chapel and consist of nicely developed ultramylonites. A 150 meters thick stripe of Variscan upper greenschist mylonites with limited lower greenschist overprint follows to the south. They macroscopically consist of well foliated gneisses with feldspathic augens. In thin section they show stable biotite along the foliation and in clast shadow zones; large feldspars overgrow the older mylonitic foliation. Sericitized relics of andalusite are also found. Garnet is often replaced by chlorite and plagioclase is deeply sericitized.

Lower greenschist mylonites are again met towards the south. A thick band of ultramylonites forms the SE-NW trending ridge descending to Lake Como. Typical mylonites and ultramylonites make up the rest of the profile. Brittle fracturing becomes apparent beginning from ca. 100 meters north of the Monte Grona line. However, the thickness of the brittely deformed rocks is not comparable to that along the Camaggiore profile. Exposures immediately north of the Monte Grona line are quite poor but the maximal thickness of the breccias is ca. 20 meters.

The dolomites south of the Monte Grona line show brittle deformation for a thickness of ca. 300 m, however with only limited cataclastic breccias.

Structures

Planar elements. The attitude of foliation of lower greenschist mylonites is relatively constant with intermediate dips to the SW (Fig. 38 A). Shear bands are observed in these mylonites mainly immediately north of the Monte Grona line. Shear bands dip to the SSW and are steeper than the mylonitic foliation (Fig. 38 B). The ultramylonites, which are less common than in Camaggiore, generally have the same orientation as the shear bands.

Foliation in upper greenschist mylonites is discordantly cut by the ultramylonites of the S. Amate chapel and is sub-parallel to the lower greenschist foliation in the central part of the profile.

Stretching lineations. Stretching lineations on the mylonites of the Monte Grona profile typically trend

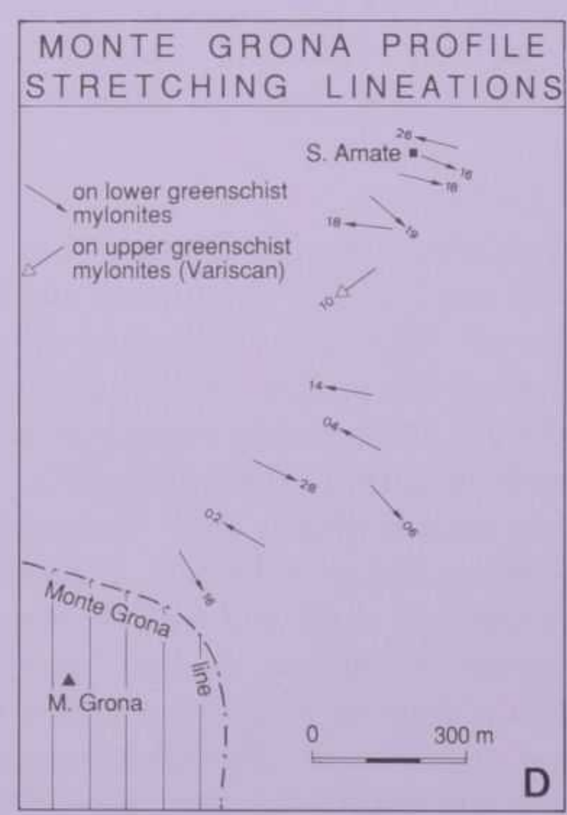

FIG. 38 D - Stretching lineations along the Monte Grona ridge section.

NW-SE to WNW-ESE. Plunges are always gentle and can be both towards the NW and to the SE (Fig. 38 D).

Stretching lineations in the Variscan upper greenschist mylonites gently dip to the SW and have therefore an aberrant position (Fig. 38 D).

Kinematic indicators. All shear criteria in lower greenschist mylonites and ultramylonites indicate a movement of the southern block towards the ESE with a variable downward component in present day coordinates.

Kinematic indicators on the few upper greenschist mylonites indicate a top-to-the SW sense of movement along the SW-dipping stretching lineations.

Outcrops in the thick-wooded eastern flank of the Monte Grona-Monte Bregagno ridge are very poor. However, in several isolated outcrops (for instance east of Breglia and along Lake Como) (Fig. 27) lithologies similar to those of the Monte Grona profile are found. Kinematic indicators in mylonites and ultramylonites are always indicative of a movement of the southern block towards the E-SE. Pseudotachylites are relatively common.

Brittle deformation and cataclastic breccias south of the Monte Grona line, *i.e.* in the hangingwall of the Lugano fault, are particularly well developed at the base of the Gneiss Chiari east of Breglia. Meso- and micro-structures (fracture patterns, small-scale folds, tension fractures, etc.) in the cataclastic stripe at the base of the Gneiss Chiaro (Plate IX, Fig. 5) show a movement of the southern block towards the E-ESE, *i.e.* the same as recorded by the mylonites of the footwall. This indicates that cataclasites, mylonites and ultramylonites all formed under similar kinematic conditions.

The Monte Grona line becomes parallel to the shear bands east of Breglia (Fig. 27) and has therefore the same setting as in Camaggiore.

Val Sanagra

The Val Sanagra profile is exposed along the valley road north of the outcrops of the Manno Conglomerate (Fig. 27). Exposure conditions along most of the profile are rather poor. The fault rocks are lithologically similar to those encountered in the already described profiles. Lower greenschist mylonites are quite common. They often preserve relics of biotite. Ultramylonitic stripes, sometimes thick and well developed, are found parallel or oblique to the typical lower greenschist mylonites. As in the profiles further to the east, remnants of upper greenschist mylonites are found within the fault zone. They are always characterized by the presence of biotite.

Compared with the previously described profiles, biotite is more common indicating a less pronounced lower greenschist overprint. Biotite is absent in the better developed lower greenschist mylonites and in ultramylonites.

The mylonitic foliation generally steeply dips to the SW and, in the central part of the profile, to the S. The E-W trending Monte Grona line is clearly oblique with respect to the mylonitic foliation immediately north of the line; it is however parallel to some of the shear bands (similar relationships as in Camaggiore).

Typical for the Val Sanagra profile is the widespread brittle fracturing observed at all scales along most of the profile.

Lithologies

The southernmost outcrops of Stabbiello Gneiss with no retrograde overprint are found ca. 1000 m north of the Monte Grona line. The first rocks belonging to the Monte Grona fault zone are found 700 m north of the line (Fig. 39 A) and consist of strongly folded and altered lower greenschist mylonites. They are limited to the south by a wide band of nicely developed ultramylonites (immediately north of the road junction to Alpe Logone) which are intruded by small veins filled with calcite, chlorite and iron oxides. No further outcrops are seen until the small quarry south of the junction. The strongly weathered (rusty) rocks of the quarry mainly consist of strongly folded and fractured biotite-bearing mylonites; chlorite and sericite are common. Further to the south brittle deformation tends to obliterate older features. Some 200 m north of the Val Grande line, fractured and folded lower greenschist mylonites are found. Small biotite crystals are still preserved. Ultramylonitic bands are common both at meso- and micro-scale; they tend to become more common moving southwards. Well developed ultramylonites are found some 70 meters north of the line. Further to the south the fault rocks consist of pervasively fractured lower greenschist mylonites with common ultramylonites.

The southernmost metamorphic rocks consist of a fault-bounded massive greenish-coloured gneiss forming a thin horizon along the Valle del Carbone (Fig. 39 A). The gneiss consists of very fine-grained feldspars, quartz, sericite, epidote and chlorite. Veins are very common; they are not deformed (Plate VIII, Fig. 6) and are filled with quartz, blue chlorite (penninite) and calcite successively. The rock strongly recalls the leucocratic gneiss outcropping in the Camaggiore area. Both bodies lie immediately north of the Monte Grona/Val Grande line.

Manno Conglomerate together with tectonic breccias with dolomite clasts outcrop in the Valle del Carbone. South of the Valle del Carbone the Dolomia Principale is intensively brecciated.

Structures

Planar elements. The foliation of both lower greenschist mylonites and ultramylonites in the northern

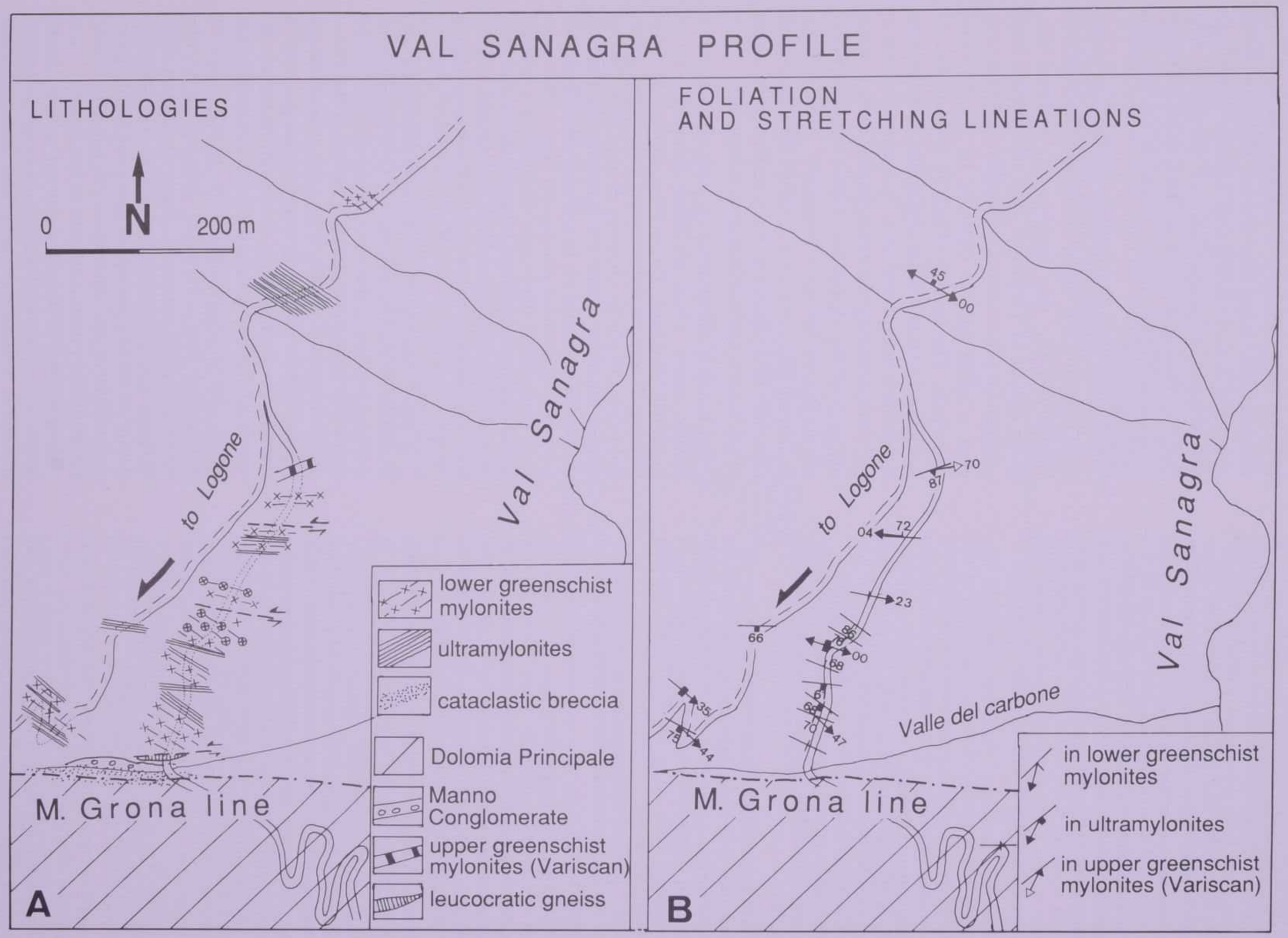

FIG. 39 - Val Sanagra profile. Lithologies (A) and structural elements (B).

part of the profile dips to the NE (Fig. 39 B). The ultramylonites show very well developed, NNW-dipping shear bands. South of the road junction to Alpe Logone, the strongly disturbed foliation of ultramylonites and lower greenschist mylonites is on the whole sub-vertical and strikes roughly E-W. Shear bands are rarely visible at a macroscopic scale. They are, however, commonly seen under the microscope. South of an E-W striking fault, the foliation in the lower greenschist mylonites is still sub-vertical but strikes NW-SE to WNW-ESE. Ultramylonites are either parallel to the mylonitic foliation or oblique to it with a roughly E-W strike. This configuration is similar to that found along the road to Logone and continues until the E-W striking Monte Grona line.

Stretching lineations. Stretching lineations are generally poorly developed in the fault-rocks of the Val Sanagra section. The widespread fracturing causes also a certain dispersion of the measurements. On the whole, however, stretching lineations show a general E-W, WNW-ESE trend (Fig. 39 B) compatible with the kinematic picture obtained from the better exposed sections (see Monte Grona and Camaggiore).

The only stretching lineation observed on a Variscan upper greenschist mylonite steeply dips to the E (Fig. 39 B).

Shear indicators. Kinematic criteria in the least disturbed lower greenschist mylonites and ultramylonites constantly show a sinistral movement along the fault zone. This corresponds to an eastward movement of the southern block along the WNW-ESE trending stretching lineation.

Shear criteria in Variscan upper greenschist mylonites indicate a southern block-downwards sense of movement.

Folds. Only microscopic folds have been observed. They typically fold mylonitic quartz bands and are therefore younger. They could be contemporaneous to or younger than ultramylonitization. In any case, the folds show a vergence compatible with the sinistral movement deduced from the other fault rocks.

Faults. The fault pattern is complex and not easily understandable. This is due to the superposition of two distinct tectonic "events" which both caused the formation of striated fault surfaces: the Mesozoic movement, which appears as sinistral in present day coordinates (Fig. 3) and a minor reactivation during Alpine steepening. The major faults found along the profile generally strike E-W. They show gently E- and W-dipping striations and sinistral sense of movement. They could be related to the Mesozoic phase of deformation. A more sophisticated treatment of the small scale fractures after the methods of Arthaud (1969) and Angelier (1975) did not produce clear distributions. I am therefore not able to confirm the reconstruction proposed by Perotti (1987).

Other outcrops between Val Sanagra and Val Rezzo

The next valley west of Val Sanagra is the Val Cavargna (Fig. 27). The first metamorphic rocks are found along the road ca. 100 m north of the Dolomia Principale.

The rocks of the fault zone show only a limited overprint and still preserve their Variscan paragenesis. They consist of a strongly fractured two-micas, staurolite-bearing gneiss. Fluid-controlled alteration is very strong and leads to widespread sericitization of plagioclase and staurolite. The Dolomia Principale south of the contact is strongly fractured. The contact itself is not exposed along the road.

In a small and poor outcrop ca. 500 east of the road, beneath the Sasso di Cusino (coordinates 733.13/103.30), the uppermost metamorphic rocks consist of a tectonic slice of Gneiss Chiari overlain by a few meters of breccias and followed, after a small gap in outcrop, by the Dolomia Principale. The breccia (Plate X, Figs. 1 and 2) has gained a certain fame in the last years because it has been interpreted as a "transgressive horizon" demonstrating the stratigraphic nature of the omission of the Upper Permian to Middle Triassic sediments between basement and Dolomia Principale (Gaetani *et al.*, 1986). As a matter of fact there is no evidence for a sedimentary origin of the breccia which I interpret as tectonic.

The breccia is composed of up to cm-sized clasts in a tectonic matrix. Clasts are often dolomitic and are commonly brecciated, fractured or crossed by veins pointing to repeated deformation. Also quartz grains and muscovite can be observed derived from the underlying basement. Clasts are generally angular; only the smaller ones tend to be sub-rounded. All dimensions are represented among the clasts which also show no preferred alignement. The matrix is strongly recrystallized; it can be dolomitic or calcitic with fine-grained micas, quartz grains etc. The breccia is texturally very similar to that found in Val Rezzo (section below).

A small outcrop of rocks overprinted by deformation related to the Monte Grona fault zone is found further to the west along the road in the lower Val Rezzo immediately north of the Dolomia Principale (Fig. 27). The contact itself is not exposed. The rocks are strongly folded gneisses to schists. Low-temperature overprint is shown by ductilely (?) stretched quartz grains in quartz-rich layers. Quartz grains also show, beside common deformation bands, incipient dynamic recrystallization (subgrain rotation). Fluid circulation must have been very intense and led to the advanced sericitization of plagioclase and to the chloritization of biotite and muscovite. Typical Stabbiello Gneiss follows tens of meters to the north.

This is the westernmost outcrop of the Monte Grona fault zone where signs of dynamic recrystallization in quartz are observed. Further to the west, only brittle deformation affected the Variscan rocks along the fault zone.

Upper Val Rezzo

Two very poorly exposed profiles from the Stabbiello Gneiss to the Dolomia Principale can be observed along two streamlets on the southern side of the upper Val Rezzo (Figs. 27 and 40). The profiles are strongly tectonized also by Alpine movements reactivating the segment of the Monte Grona line west of Val Rezzo (chapter 3.2). Observations from other scattered outcrops are also integrated in the description.

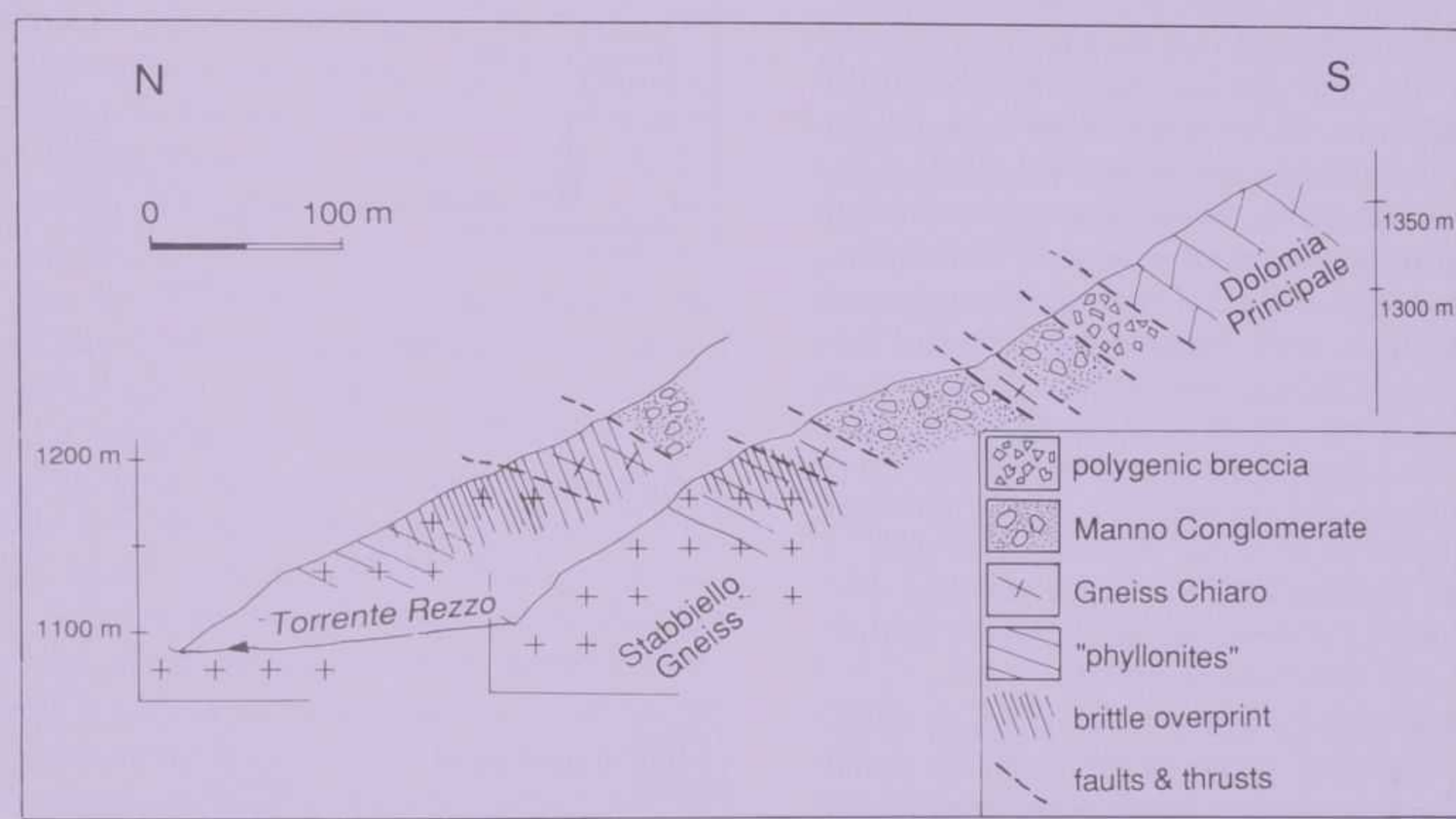

Fig. 40 - The two profiles of the upper Val Rezzo from the undisturbed Stabbiello Gneiss to the Dolomia Principale.

The area and the profiles have been described in detail by Lehner (1952). However, possibly because of the in the meanwhile worsened outcrop conditions and because of different interpretation, my description will differ from that given by Lehner. In particular I am of the opinion that Lehner understimated the degree of tectonization of the profile.

Typical Stabbiello Gneiss is found from the valley floor (Fig. 40) up to an altitude of ca. 1100-1150 m along both profiles. The upper part of the Stabbiello Gneiss is fine-grained, with a dm-spaced foliation gently dipping to the S-SE.

The gneiss has a very fine-grained (10-20 microns) groundmass of quartz, possibly feldspars, chlorite, epidote and sericite. A general crystallographic preferred orientation is observed. Tenths of mm-sized clasts are found in the groundmass. They generally consist of muscovite and smaller sericitized plagioclases. Folded and disrupted relics of dynamically recrystallized quartz ribbons are often found. Calcite is common throughout the thin sections. The calcium is probably derived from the alteration of plagioclase.

On the base of the fine grain-size and on the crystallographic preferred orientation in the matrix, I interpret these rocks as mylonites. Similar lithologies found along the Piancabella profile show stable syn-mylonitic biotite. Thus I interpret these mylonites to be of Variscan age.

These lithologies have been named phyllonites by previous authors (Lehner, 1952; Reinhard, 1964) and interpreted as weakly metamorphosed pelitic schists.

The "phyllonites" become pervasively fractured and crossed by numerous thin zones of ultracataclasites moving upsection. Above ca. 1150-1200 m altitude a macroscopic foliation is no longer visible. At ca. 1200 m a laterally discontinuous slice of Gneiss Chiari is observed. Both the top and the bottom contacts are tectonic. The Gneiss Chiari are tectonically overlain by several tens of meters of Manno Conglomerate. Its relevant thickness is partly due to tectonic repetition as shown by tectonically intercalated slices of Gneiss Chiari. Pebbles of the conglomerate often show pressure-solution pitting. Blocks of Servino quartzarenite are found along this segment of the section, however, not in place. The Manno Conglomerate is tectonically overlain by cataclastic breccias with sediment clasts (Plate X, Fig. 3).

The breccia consists of clasts ranging in size from cm to mm. The larger clasts are made up of dolomites, limestones, shales possibly derived from the Raibl Beds and a few metamorphic rocks. Casts of dissolved evaporites are common. On the whole, basement-derived clasts become less abundant upsection where dolomitic components are the dominant lithology. Clasts are generally angular. Dolomitic clasts show repeated fracturing. Smaller, mm-sized clasts consist of fairly well rounded quartz and feldspar, of idiomorphic

muscovite and, rarely, biotite. The strongly recrystallized cataclastic matrix is made up of fine-grained material of similar composition (calcite is abundant). No sorting is observed in the rock. A preferred orientation of elongated clasts is also not visible. Similar breccias in a similar tectonic position are found also further to the west (Bocchetta di S. Bernardo).

The breccia has been first described by LEHNER (1952). He interpreted it as a sedimentary breccia belonging to the Raibl Beds. On this interpretation he formulated the hypothesis of a transgression of the Dolomia Principale on a previusly exposed basement. In contrast to this, I interpret this breccia as tectonic. This interpretation is based on the strong microscopic similarities with the breccias found by FROITZHEIM (1989) along Liassic fault scarps in the Austroalpine domain. In general, extension-related Mesozoic breccias differ from shortening-related Alpine ones in that they have practically no sparitic cement. The tectonic origin of the breccia of the Val Rezzo is also supported by geometric relationship between the Dolomia Principale and the basement. The Dolomia Principale of the upper Val Rezzo - upper Val Colla region, is clearly oblique with respect to the basement/sediment contact; the polygenic breccia therefore cannot represent the stratigraphic base of the Dolomia Principale. I propose that the breccia formed in the Mesozoic during extensional movements along the Monte Grona/ Val Grande fault zone (section 2.3.2).

Some poorly exposed remnants of variegated shales and marls belonging to the Raibl Beds are found further to the west, more or less at the same structural level as the polygenic breccia (*i.e.* coordinates 727.55/103.05). Similarly to the breccia they have been regarded as the stratigraphic base of the Dolomia Principale (LEHNER, 1952). The shales are fault-bounded; they dip to the SW and thus are clearly discordant with respect to the sediment/basement contact. The bedding in the shales is also subparallel to bedding in the Dolomia Principale; the shales, however, are along strike with respect to bedding of the Dolomia Principale and therefore cannot be its stratigraphic base. The relationships are very similar to those depicted in figure 7. I interpret these relics of Raibl Beds to be extensional duplexes formed during Mesozoic movements along the Lugano normal fault.

The Dolomia Principale is found beginning at ca. 1300 m altitude. The base is strongly fractured. Bedding in the Dolomia Principale dips to the SW and is therefore clearly oblique with respect to the E-W striking Monte Grona line.

Piancabella profile

The Piancabella profile lies on Swiss territory some 500 m west of the Bocchetta di S. Bernardo (Fig. 27). Exposures along the streamlet descending towards the NE are fairly good but limited to the stream bed itself (Fig. 41). The streamlet shows a profile from the undeformed Stabbiello Gneiss to the Gneiss Chiari and, although with a covered contact, to the Dolomia Principale.

Similarly to the upper Val Rezzo sections, the Stabbiello Gneiss found at the base of the profile grades upwards into more massive gneisses with a more widely spaced foliation ("phyllonites"). This foliation dips to the SE (Fig. 41).

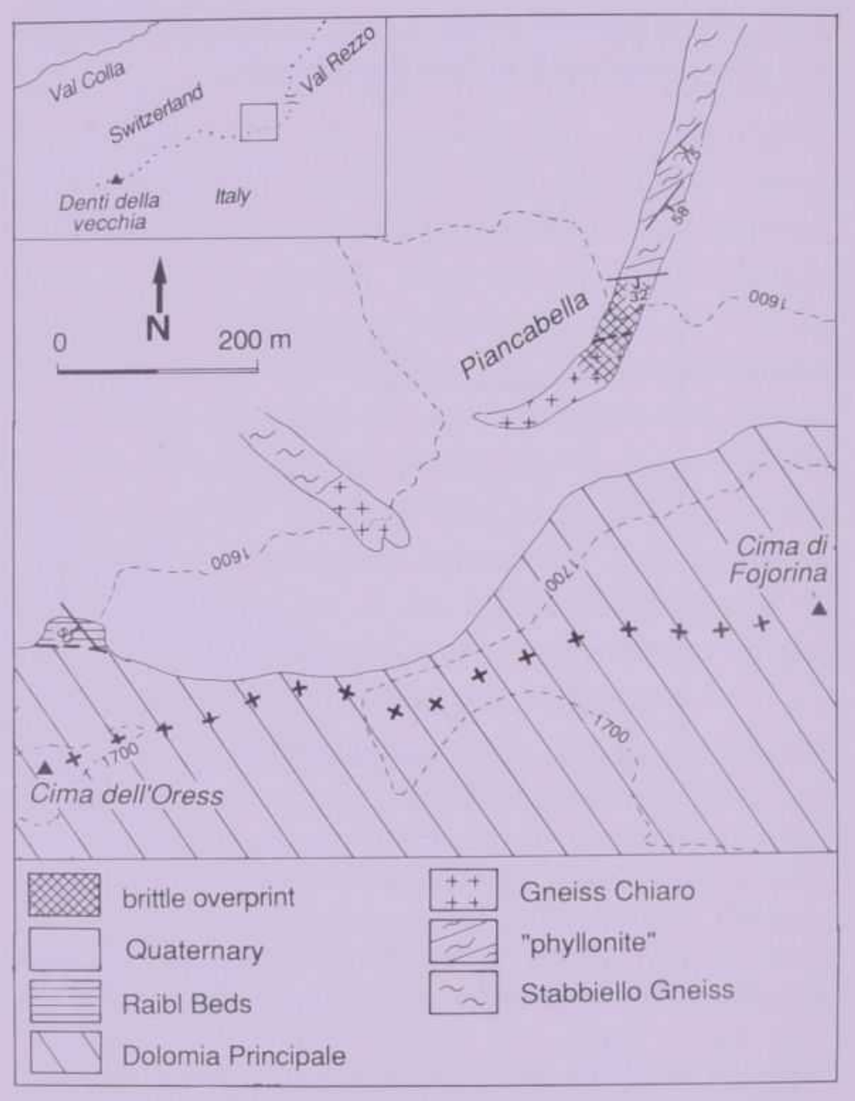

FIG. 41 - Map of the Piancabella area.

The "phyllonites" of the Piancabella section are perfectly comparable with those described in the upper Val Rezzo. They clearly show their mylonitic origin. They consist of a fine-grained (typically less than 20 micron) quartzo-feldspathic groundmass with sometimes very abundant tiny crystals of biotite and chlorite. Biotite is obiously stable and syn-kinematic. Clasts are generally ca. 0.1 mm large and consist of sericitized plagioclase, muscovite and rarely biotite. Ribbons of completely dynamically recrystallized quartz grains are common. Grains are typically 30-50 microns large and are polygonal. Both in the quartz ribbons and in the groundmass a good crystallographic preferred orientation is observed. Post-mylonitic calcite is common in thin section both as small scattered crystals and as long foliation-parallel crystals. On the base of the stability of biotite I consider these mylonites to have formed under upper greenschist conditions and therefore to be of Variscan age. Lower greenschist overprint is not observed along the Piancabella section.

"Phyllonites" are found till the foot-path at 1585 m altitude. Fracturing becomes progressive-

ly widespread moving up-section: small-scale faults, cataclastic breccias, veins still open and partially filled with iron oxides are commonly observed. Brittle overprint increases upwards until the tectonically emplaced Gneiss Chiari are found. The Gneiss Chiaro itself is intensively fractured. The section is then covered until at 1650 m altitude the Dolomia Principale is found. The Dolomia Principale dips to the SW and is therefore oblique with respect to the Monte Grona line.

Rocks belonging to the Raibl Beds are found in small, strongly tectonized outcrops east and west of the described section (Bocchetta di S. Bernardo and for instance coordinates 725.72/102.55). These occurrences are discontinuous and elongated parallel to the Monte Grona line, *i.e.* in a E-W direction. Bedding in the Raibl Beds dips to the SW (*e.g.* Fig. 41) and is thus discordant with respect to the Monte Grona line but parallel to bedding in the Dolomia Principale. These relationships are typical for fault-bounded slices emplaced along the Monte Grona line during Mesozoic extensional movements (extensional duplexes; Fig. 7). As in the upper Val Rezzo, the Raibl Beds have been considered to rest transgressively on the previously exposed basement and to be the stratigraphic base of the Dolomia Principale. This interpretation cannot be accepted because of the obliquity in strike between the Dolomia Principale and the basement/sediment contact.

4.3.4 *Some considerations on the evolution and the protolith of the fault rocks of the Monte Grona/Val Grande fault zone*

The variety of fault rocks and their distribution along the Monte Grona fault zone are due to two main factors:

a) the fact that the Lugano normal fault reached a depth of least 12-15 km, *i.e.* it crossed the thermal boundary zone across which the onset of quartz plasticity occurs. Rocks above the transition were brittely deformed; rocks below the zone were deformed by quartz intracrystalline plasticity. The consequence is that a variety of fault rocks formed along the fault zone (Fig. 28);

b) the fact that the warm footwall and the cooler hangingwall were juxtaposed during normal faulting. The juxtaposition of two blocks with differing temperature caused a cooling of the fault zone and therefore a change in deformational mechanisms with time while kinematic conditions remained constant. The deformation pattern across the Monte Grona/Val Grande fault zone is explainable only assuming that tectonic movement along the fault was faster than heat transmission, *i.e.* movement along the normal fault caused a distortion of the isotherms (Fig. 28).

Under these circumstances only in the first stages of deformation fault rocks of the same kind are formed on the two sides of the fault. Both hangingwall and footwall had cataclastic rocks in the upper part of the crust and mylonites below the limit for the onset of quartz plasticity. With proceeding extension, the evolution of the fault rocks of the two blocks diverges.

The lower greenschist mylonites formed in the *footwall* were progressively disactivated by a different deformational mechanism mainly controlled by plagioclase-derived aggregates (STÜNITZ, 1989) (chapter 4.3.2). This change occurred under a constant kinematic frame. I have interpreted this change as a consequence of deformation proceeding under decreasing temperatures and therefore decreasing quartz plasticity. Under these conditions shear bands are expected to develop. GAPAIS and WHITE (1982), discussing the meaning of shear bands, noticed that:

"Shear band development appears as an important process in accomodating large strain deformations at relatively low temperatures and shear bands are common structures in mylonitic zones... Both theoretical and experimental studies... show that a general shear band structure *corresponds to a strain enhanced rheological instability which reflects the onset of inhomogeneous deformation and allows the material to soften and accomodate further deformation. That is the development of ductile shear bands should result in replacement of the earlier microstructure or fabric by another which should be softer*" (italics are mine). With increasing difficulty to plastically deform quartz, deformation becomes localized and controlled by the breaking-down plagioclase. From the shear bands, resulting from this localized deformation, the ultramylonitic horizons evolved. It can be seen both in thin section as in the field (see for instance the Camaggiore profile) that ultramylonitic bands are both parallel to the mylonitic foliation and parallel to the shear bands. The onset of a new deformational mechanism allows for the accommodation of further strain. In fact, the last tectonic movements along the Monte Grona/Val Grande fault zone used a surface which was partially parallel to the mylonitic foliation (*e.g.* in the Monte Grona section) and partly parallel to the shear bands (e.g. in the Camaggiore section) (Figs. 27 and 42). With proceeding extension and therefore with decreasing temperature in the fault zone, brittle deformation is activated.

The fault rocks of the *hangingwall* west of Camaggiore (Fig. 27) consist only of cataclasites. Mylonites must also have been formed in the deeper part of the hangingwall but were later displaced by movements along the fault. In present day coordinates they should be found in the Monte Legnone area (Fig. 6) as required by the kinematics of movements along the fault (Fig. 3).

The question then arises of the possible protolith for the lower greenschist mylonites. Typical for these mylonites is the ubiquitous presence of large "augen" made up of microcline twinned K-feldspars and poikiloblastic plagioclases (chapter 4.3.2). The dimensions and the nature of these clasts strongly suggest an origin from an orthogneiss rather than from a paragneiss. In fact neither the Stabbiello Gneiss nor the Monte Muggio Gneiss show crystals with similar features.

Excluding the Gneiss Chiaro because of its limited extent, the only lithology which could have produced the clasts is the Monte Legnone Gneiss (El Tahlawi, 1965) which is part of the hangingwall of the Lugano normal fault. Thin sections of the Monte Legnone Gneiss contain feldspars perfectly comparable with those found in the mylonites (*e.g.* Plate X, Figs. 4 and 5).

4.3.5 *The age of the fault rocks of the Monte Grona/Val Grande fault zone*

The Monte Grona/Val Grande fault zone is discordant with respect to the Variscan foliation of the Val Colla zone, of the western Dervio-Olgiasca zone and of the Monte Muggio Gneiss (Figs. 2 and 27). The fault zone must therefore be younger than the Variscan orogeny. On the other hand, it is older than the Alpine orogeny since it has been steepened by Alpine folding. In fact, Alpine strike-slip kinematics, required by the described indicators, would not explain the decreasing syn-kinematic gradient observed across the fault zone and along the fault zone N of the Monte Grona/Val Grande line.

Direct evidence of the age of the mylonites does not exist since the mylonites have not been dated. The only published radiometric ages (Mottana *et al.*, 1985; and references therein) have been carried out on the Monte Muggio Gneiss and on the gneisses of the Dervio-Olgiasca zone. The main result obtained was the contrast between the basement rocks north of the Monte Grona/Val Grande line with Late Permian to Jurassic ages, and those south of the line with Variscan ages (Fig. 4).

At a closer look, the K-Ar datings of Mottana *et al.* (1985) show peculiar features. The most apparent is that biotite ages of the Dervio-Olgiasca zone are systematically older than the muscovite ages (Fig. 4). This is obviously inconsistent with the fact that the closing temperature for muscovite is higher than that of biotite (500°C and 300°C respectively) (Purdy and Jäger, 1976). I cannot say if this trend is significant or if it is a result of unclean sampling and dating. A certain bias on the quality of the published data is given by the absolute lack of any kind of structural analysis of the measured samples. Moreover, no description of the dated micas is given. This is particularly regrettable since the Dervio-Olgiasca zone underwent a long and complex history and different generations of micas are likely to be present (chapter 2.2.1); very careful structural and petrographic work would be therefore needed in order to correctly interpret the measurements.

In spite of the problems I consider the contrast between the Late Permian to Early Jurassic ages of the Dervio-Olgiasca zone and the Variscan ones of the Monte Muggio zone to be real. These results obviously go hand in hand with the presence of lower greenschist mylonites in the Dervio-Olgiasca zone (north of the line) and of only brittely deformed rocks in the Monte Muggio zone.

4.4 The geometry of the Lugano normal fault and the Monte Generoso basin; some general considerations about listric faults

4.4.1 *The geometry of the Monte Generoso basin*

On the base of the data discussed in the preceding chapter I propose a model for the geometry of the Lugano normal fault and for the evolution of the Monte Generoso basin (Fig. 42). I will use the observations made on the fault rocks to constrain the geometry of the fault zone at depth; data from the sedimentary cover allow for a detailed reconstruction of the hangingwall and for an independent check of the geometry of the fault as derived from the fault rocks.

The development of the basin is mostly controlled by tectonic activity along the east-dipping Lugano normal fault. The emergence of the Lugano normal fault at the surface roughly corresponds to the Lugano line. The uppermost leg of the normal fault was steep (probably steeper than 50°) as shown by relationships at the surface (Bernoulli, 1964) and as required by the thickness of syn-rift sediments immediately east of the fault (see below). A similar angle is formed by bedding in the Dolomia Principale and the base of the western part of the Alpine thrust sheets and of the sedimentary units belonging to their footwall. This obliquity was interpreted as con-

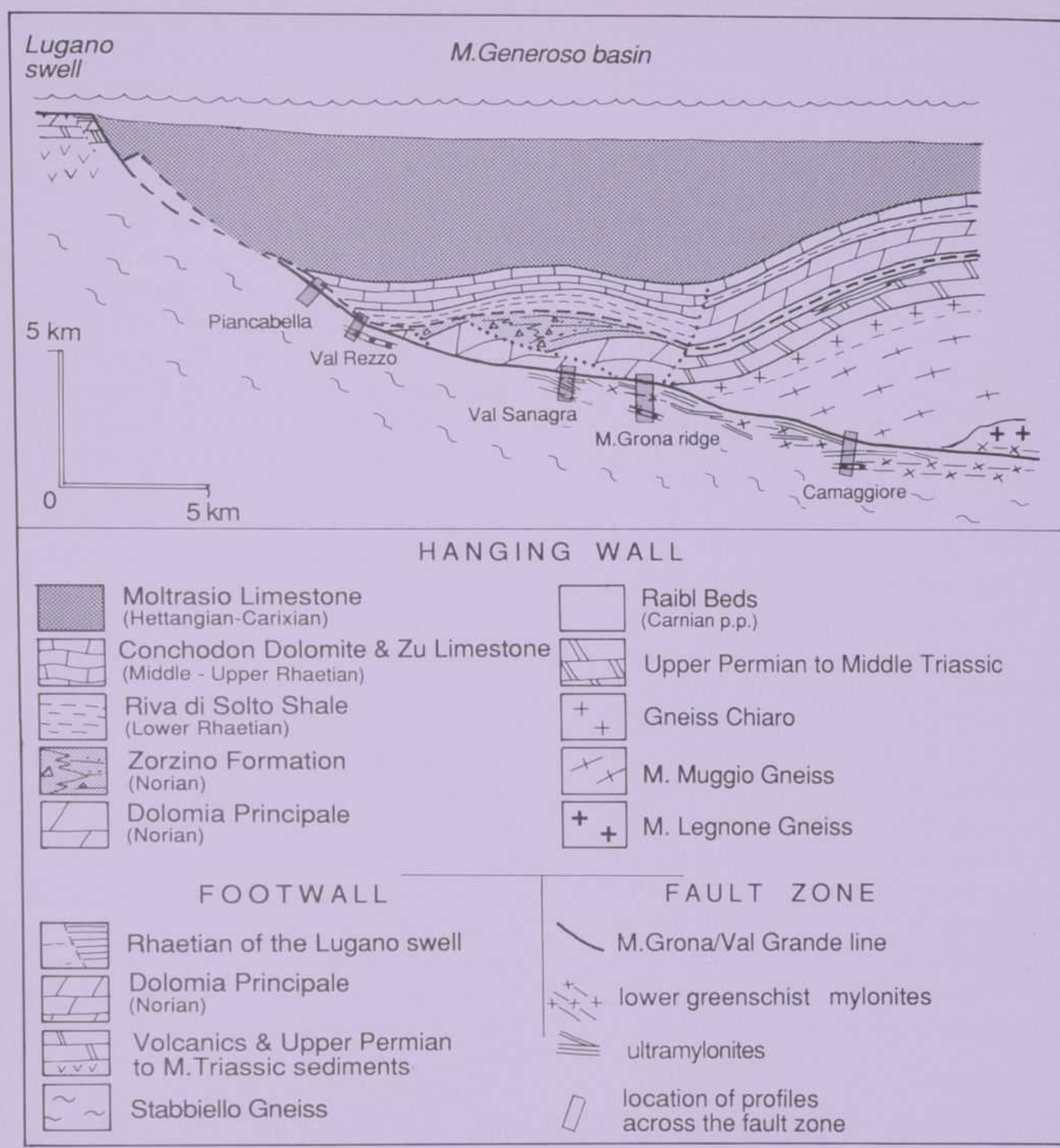

FIG. 42 - Profile through the Monte Generoso basin during the Liassic. The profile is constructed using the fault rocks of the Monte Grona/Val Grande fault zone to constrain the deep geometry of the Lugano normal fault; geometry of the hangingwall is derived from the Alpine thrust sheets and their footwall (see text).

trolled by the Lugano normal fault (Figs. 3 and 19): the angles between bedding and geometric base of the units are therefore cut-off angles (see chapter 3.2.2).

The shape of the Lugano normal fault at deeper crustal levels is constrained by the presence of lower greenschist mylonites in the footwall from the Val Sanagra eastward, that is ca. 14 km east of the Lugano line. These mylonites formed at depths around 10 km.

The temperature required for intracrystalline deformation in quartz is around 270°C-300°C (section 4.3.2). The translation of temperatures to depths is difficult because of the impossibility to know exactly the geothermal gradient during extension. Estimates from actualistic examples in similar geological settings assume depths of 10 km or less for the 300°C isotherm (SMITH *et al.*, 1975; LACHENBRUCH and SASS, 1979 for the Basin and Range Province; ROYDEN *et al.*, 1983 for the Pannonian basin).

The depth of 10 km of the normal fault some 14 km east of its emergence at the surface implies a decrease of dip with depth. The shape of the fault is therefore listric. The transition from the steep to the gently-dipping leg is not documented so that it cannot be said whether the transition was abrupt or gradual.

Further to the east, the Lugano normal fault must have maintained its gentle dip: the mylonites of the Camaggiore area, ca. 27-28 km east of the Lugano line, are better developed but not substantially different from those of the Val Sanagra. The depth of formation was therefore only slight-

ly larger. Further to the east the Monte Grona/ Val Grande fault zone disappears in the subsurface beneath the Monte Legnone so that no direct observations are possible.

The shape of the Lugano normal fault as reconstructed on the base of the fault rocks is compatible with the thicknesses of syn-rift sediments which were accommodated in the basin. In fact, the syn-rift sediments (from the Dolomia Principale to the Moltrasio Limestone) in the deepest part of the basin have a thickness of 8-9 km when decompacted (see section 5.1.1 for a discussion of the decompaction procedures). A gentle dip of the Lugano normal fault further to the east is also compatible with the reduced thickness of syn-rift sediments in the western Albenza region. Extension above a flat lying segment of a listric normal fault causes only limited subsidence.

The reconstruction of the geometry of the Monte Generoso basin allows for an estimate of the amount of extension. Comparing along the E-W striking profile the length of the base of the Dolomia Principale and that of the Conchodon Dolomite, an extension of ca. 6 km can be deduced for the Generoso basin from the Norian to the end of the Triassic. By comparing the length of the Conchodon Dolomite with that of the top of the Moltrasio limestone (Fig. 42) a further extension of ca. 7 km can be deduced for the Early Jurassic. This gives an approximate total of 13 km.

4.4.2 *Some general considerations on listric faults*

For the whole distance between the Lugano and Camaggiore regions deformation associated to the Monte Grona/Val Grande fault zone remains concentrated in a well defined stripe. This means that deformation along the Lugano normal fault from the surface down to at least 12-15 km was discrete at a crustal scale; the change in deformation mechanism from cataclasis to intracrystalline plasticity in quartz caused only a widening of the fault zone but not a transition from discrete to diffuse deformation, again on a crustal scale [1].

This brings us to the point of why (some?) faults which are steep at the surface attain gentler dips at depth. The explanation given by supporters of the pure-shear extension (see 5.2.1) is that crustal normal faults acquire gentle dips when they merge in the "ductile" layer of the crust where deformation is considered to be diffuse. Authors with a rather "simple shear" philosophy tend to attribute the gently-dipping attitude of low-angle normal faults of the reactivation of older surfaces (*e.g.* DYMENT, 1989 and references therein). In same cases (KLIGFIELD *et al.*, 1984) the change in dip has been explained with a change in deformation mechanism.

Evidence from this study indicates that the gently dipping position of the Lugano normal fault is probably not related to the transition from cataclasites to mylonites. The flattening of the fault could be related to the reactivation of a preexisting (flat-lying?) anisotropy surface possibly represented by the upper greenschist mylonites which have been found in different profiles across the fault zone. These mylonites are neither kinematically nor metamorphically compatible with the lower greenschist mylonites and were therefore interpreted as Variscan (chapter 4.3.2).

4.5 EXTENSION IN THE WESTERN LOMBARDIAN BASIN: ESTIMATE OF STRETCHING FACTOR

The derivation of a stretching factor for the Lombardian basin is important because it enables estimates of crustal thickness and helps in constraining the extension-related subsidence. Studies on extensional tectonics in the western Southern Alps are unfortunately only at their incipient stage. No analysis comparable to that presented above has been carried out on the major basins of the Southern Alps like the Monte Nudo basin and the Sebino trough. For this reason my estimates should be considered only as semiquantitative.

I will consider a section running from the Gozzano swell in the west to the Albenza zone in the east (Fig. 1). This section comprises the Monte Generoso basin, object of this study and the Monte Nudo basin; the section is long enough to include two major basins (longer than the average dimensions of the fault-bounded blocks) but at the same time short enough to exclude large lateral variations in the stretching factor. The M. Generoso basin has been exhaustively discussed in this study. I will now, on the base of published data, reconstruct the tectonic evolution of the M.

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[1] I think it is important to emphasize these observations because in a large number of articles (*e.g.* BRUN and CHOUKROUNE, 1983; KLIGFIELD *et al.*, 1984) the terms ductile and diffuse are considered, more or less implicitly, as interchangeable. In the view of these authors, the crust is subdivided in an upper part which deforms brittely and a lower one where deformation is diffuse and accommodated by intracrystalline plasticity. As shown by the Lugano normal fault, deformation can be accommodated ductilely (at the scale of the fault zone) but still remain concentrated, *i.e.* discrete at the crustal scale.

Nudo basin. A tentative upper crustal profile from the Gozzano swell in the west to the Albenza zone in the east will then be reconstructed. On the base of the profile, an estimate of the stretching factor experienced by the western Lombardian basin during rifting will be given.

4.5.1 *The Monte Nudo basin*

The M. Nudo basin (Fig. 1) has been studied in a number of publications (Van Houten, 1929; Nangeroni, 1932; Bernoulli, 1964; Kälin and Trümpy, 1977 etc.). Extension in the basin begins in the Early Liassic (Kälin and Trümpy, 1977 mention but not further describe extension in the Late Triassic) and lasts until the Toarcian (see chapter 2.3.1). Some 3000 m (thickness corrected for compaction) of syn-rift sediments were deposited during this time.

The main part of the syn-rift succession is made up by siliceous limestones of Liassic age very similar to the Moltrasio Limestone of the Generoso basin. The succession is followed by the S. Giulio Formation characterized by the abundance of terrigenous clasts. After the Domerian when pelagic sedimentation takes place, the fine-grained turbidites of the Valmaggiore Formation are found.

The evolution of the M. Nudo basin is controlled by extension along a single major normal fault, the Lago Maggiore fault. Similarly to the Lugano normal fault, it strikes N-S and dips to the east. An idea about the prolongation of the Lago Maggiore fault at depth can be gained because of Alpine shortening which caused a steepening of bedding and pre-Alpine structures similar to what observed for the M. Generoso basin. The Lago Maggiore fault has no direct prolongation in a N direction since lithologic boundaries on the N-NW side of Lago Maggiore are nowhere significantly displaced (Fig. 43). The only possibility is that the Lago Maggiore fault rotates into a E-W direction towards the north passing on land in the region of Luino. Unfortunately Alpine thrusts in the area do not allow a detailed reconstruction. However, the overall present day

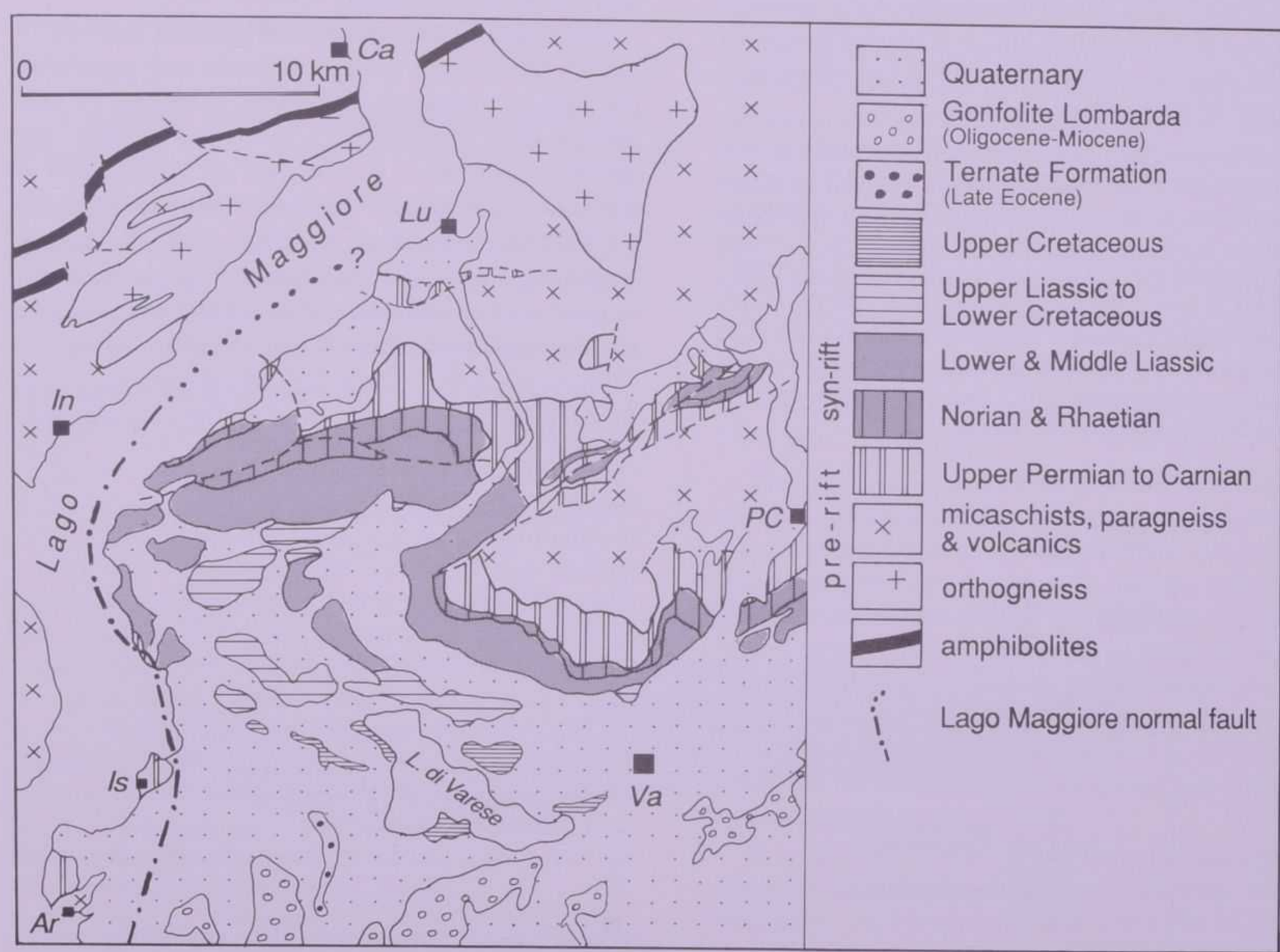

Fig. 43 - Schematic geologic map of the Monte Nudo area. Note how the Lago Maggiore fault, bordering towards the west the Monte Nudo basin cannot have a straight continuation towards the north and must therefore rotate in a E-W direction. The overall geometry closely resembles that of the M. Generoso basin. After Van Houten (1929), Nangeroni (1932) and Kälin and Trümpy (1977).

geometry closely recalls that of the M. Generoso basin. Therefore I interpret the Lago Maggiore fault as an E-dipping listric normal fault. The depth of flattening of the fault is poorly constrained because of Alpine overprint and the absence of observations on the fault rocks. Considering the thickness of pre- and syn-rift successions and of the basement south of the Luino area, a thickness of at least 5 km can be estimated for the hangingwall of the Lago Maggiore fault.

4.5.2 A stretching factor for the western Lombardian basin

A profile can thus be constructed between the Gozzano swell and the Albenza zone at the end of rifting (Fig. 44). The syn-rift succession is separated from the pre-rift basement including the crystalline basement with its pre-Norian sedimentary cover. The length of the base of the Dolomia Principale, *i.e.* of the oldest syn-rift formation, along the ca. 68 km long profile is about 46 km. This gives a stretching factor of 1.47. In the following discussion I will adopt a stretching factor of 1.5. This should be considered as a maximal estimate for the stretching factor experienced by the upper to middle crust of the considered segment of the Lombardian basin.

The stretching factor obtained here falls at the upper limit of the range between 1.3 and 1.5 proposed by HANDY (1986) for the Monte Generoso and Monte Nudo basins respectively. My stretching factor is larger than the 1.25 proposed by SCHMID *et al.* (1987). Both HANDY (1986) and SCHMID *et al.* (1987) derive their stretching factors assuming a dominio-like style of deformation. LEMOINE and TRÜMPY (1987) postulated a stretching factor of 1.1, however without specifying how they derived this figure.

5 SUBSIDENCE HISTORY OF THE LOMBARDIAN BASIN: IMPLICATIONS ON THE SEDIMENTARY EVOLUTION AND POSSIBLE MODES OF LITHOSPHERIC EXTENSION

INTRODUCTION

The previous chapters have allowed for the reconstruction of the geometry and evolution of the Monte Generoso basin and eventually for the derivation of a stretching factor for the western Lombardian basin. In this fifth part of the study, a reconstruction of the vertical movements of the same area from the rifting to the drifting stage is attempted. This allows me not only to estimate the depositional depths of syn- and post-rift lithologies, but also to discuss one of the presently most debated geodynamic issues, *i.e.* how the substantial extension observed in the upper crust is compensated at depth. This will be done by analyzing the upper crustal vertical movements as documented by the sedimentary record and by considering the implications of the tectonically-derived stretching factor.

After having introduced the general notions of subsidence analysis (chapter 5.1.1) and presented the data set (duration of rifting, thickness of sediments etc.) (chapter 5.1.2) I will derive the depths of the basin at the end of rifting and at the end of the passive margin evolution (chapter 5.1.3). These estimates, together with the crustal stretching factor derived from tectonic analysis, will be used to constrain the subsidence curve of the western Lombardian basin (chapter 5.1.4).

The classical way to derive subsidence curves is to construct depth/time diagrams for the various sedi-

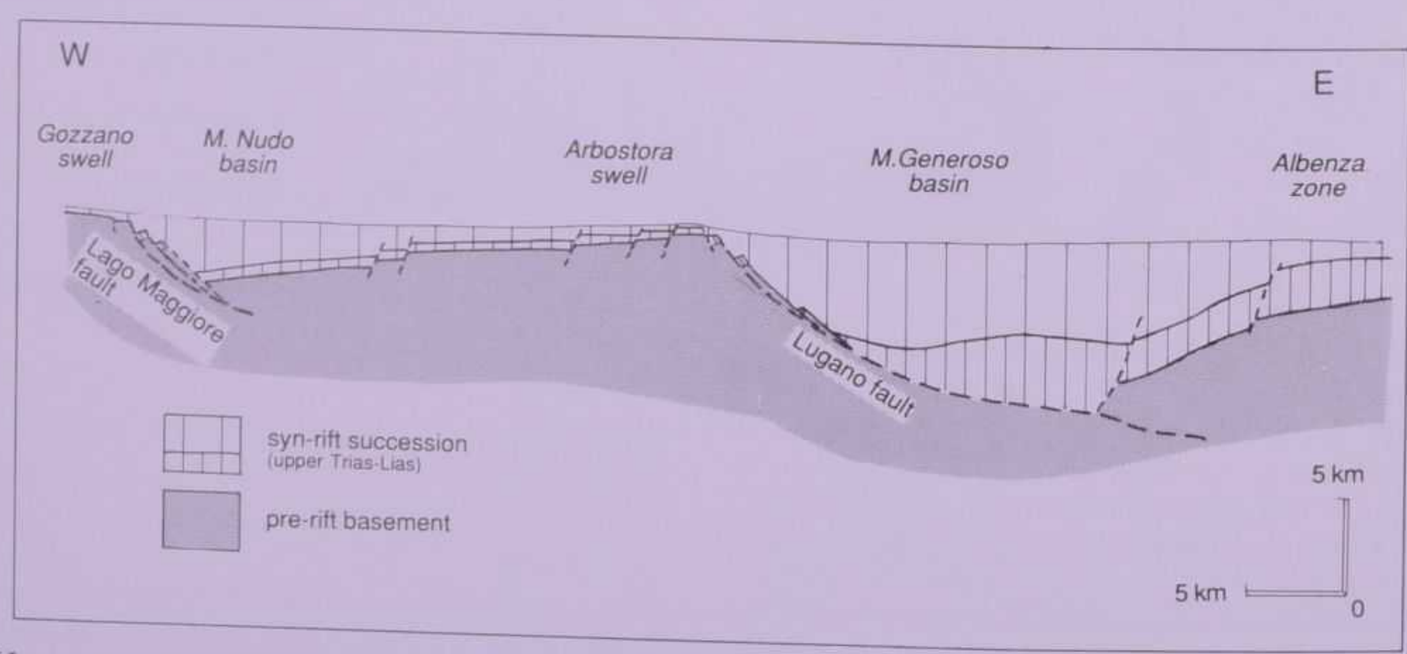

FIG. 44 - Upper crustal profile across the western Lombardian basin at the end of rifting (Toarcian). The pre-rift basement comprises the crystalline basement and its pre-Norian sedimentary cover; in the syn-rift succession the Upper Triassic has been distinguished from the Liassic.

mentary units (geohistory diagrams of VAN HINTE, 1978) and, assuming isostatic compensation (Airy), to deduce the history of the pre-rift basement (BESSIS, 1986 and reference therein). This technique, known as back-stripping, critically depends on precise estimates of many factors among which the water depth at the moment of deposition is particularly relevant. In the Southern Alps depositional depths are fairly well constrained during the Triassic; paleo-bathymetry of the Liassic formations can also be estimated (see below). On the contrary, nothing can be said about the post-rift formations which were deposited well below the photic zone. A back-stripping procedure therefore is, at least at the present state of knowledge, not applicable in the Lombardian basin.

The shape of the subsidence curve is obviously dependent on the mode of lithospheric extension. I will therefore use the subsidence curve which best fits the sedimentary and tectonic data, to speculate on possible models of extension of the South-Alpine lithosphere (chapter 5.2).

5.1 SUBSIDENCE ANALYSIS: RECONSTRUCTING THE HISTORY OF VERTICAL MOVEMENTS OF THE LOMBARDIAN BASIN

5.1.1 *General remarks on subsidence and related procedures*

General notions and definitions

Since the pioneer work by MCKENZIE (1978), it has become generally accepted that extensional sedimentary basins are produced by stretching of the crust and concomitant rise of the asthenosphere/lithosphere boundary. The extensional phase is then followed by long-term re-equilibration of the system to regional thermal conditions. In order to maintain isostatic equilibrium during these stages, changes in elevation of the different crustal levels occur, the signs and magnitudes of which depend primarily on the net density changes in the lithospheric column. The first quantitative model describing the evolution of extensional basins (MCKENZIE, 1978) was based on a "pure shear" geometry and instantaneous stretching; it has been followed by a series of refinements and integrations in order to take into account depth-dependent extension, flexural effects, finite rifting time, two-dimensional heat flow and blanketing effects of the sediments (list of references in ISSLER *et al.*, 1989).

The depth-dependent extensional model has been developed in order to consider situations in which crustal and subcrustal stretching factors differ (HELLINGER and SCLATER, 1983). Although it was developed when the simple shear model had not yet been developed, it can also be used to describe simple-shear-like geometries (ROYDEN, 1986).

The recognition that heat loss during extension occurs not only in a vertical direction but also laterally towards the cooler stable continent (COCHRAN, 1983) was also important: under these circumstances thermal subsidence becomes significant already 10 Ma after the beginning of rifting (differently from what previously assumed by JARVIS and MCKENZIE, 1980).

The initial response of the upper crustal levels to active extension can be uplift or subsidence depending on the crustal structure (LE PICHON and SIBUET, 1981) and on the mode of extension (WERNICKE, 1985; BUCK *et al.*, 1988; ISSLER *et al.*, 1989); it is commonly referred to as *fault-bounded, initial* or *syn-rift subsidence* (with the depth reached at the end of extension being the *initial depth*). Initial subsidence (positive or negative) is followed by a long-term subsidence (long with respect to the time constant of oceanic lithosphere which is 62.8 Ma; MCKENZIE, 1978 and references therein). This is the *thermal* or *post-rift subsidence* and is due to the isostatic response to slow conductive cooling and consequent progressive thickening of the previously heated lithosphere. Once the thermal anomaly has recovered and the lithosphere has returned to its equilibrium thickness, tectonic subsidence (this is the sum of initial and thermal subsidence and is also called total subsidence) ceases and the basin reaches its *final depth*. After disappearance of the thermal effects on basement subsidence, the final depth should only be a function of the crustal thinning that has occurred during rifting (ROYDEN, 1986).

Filtering out the factors controlling sediment thickness but not related to subsidence

The thickness of sediments deposited in an extending rift zone and successively on the passive continental margin is only partially controlled by the deep-seated density changes and the related crustal vertical movements. A significant control on sediment thickness is exerted by the local sea-floor morphology (changing during rifting) and by paleogeographic and paleo-oceanographic factors. These effects have to be removed in order to study the subsidence related to crustal and lithospheric extension.

It is apparent that the sea-floor morphology has a strong influence on sediment dispersal and thickness; this is especially true when, like in the Southern Alps, most of syn-rift sediments are allochthonous. Ongoing extensional faulting causes a changing morphologic relief and therefore thickness variations which are not a consequence

of variable thinning factors. The only way to minimize the error is to average thicknesses over distances larger than fault spacing. In this way the presence of different tectonic situations can smear out the corresponding deviations. As clearly demonstrated by SAWYER (1986), the use of one-dimensional data sets (sections or wells) can lead to substantial errors in subsidence and therefore crustal thinning estimates ([1]). Consequently I will derive sediment thickness and initial subsidence values averaged across the whole western Lombardian basin.

It is also straightforward that the depositional environment (*e.g.* shallow water vs. pelagic) exerts a strong influence on sediment thickness. To take into account these factors and to allow a comparison among different passive continental margins, the sediments are removed and their loading effect corrected. To do this, the sediments have to be decompacted and then replaced by an isostatically equivalent water column. The basin devoid of sediments is referred to as the *water loaded-basin*. The decompacting and compensation procedures are here briefly illustrated.

The principal assumption used in *decompacting procedures* is that the porosity of the rock depends only on the depth of burial. Effects of time and input of cements are therefore excluded. The procedure adopted to decompact the stratigraphic columns is described by SCLATER and CHRISTIE (1980), BOND and KOMINZ (1984) etc. A porosity/depth curve has to be constructed or chosen. Two different curves have been adopted in this study in order to obtain a range of values: the curve proposed by SCHMOKER and HALLEY (1982) for lithologies with at least 75% carbonates and the average/mean curve of the same authors. The decompaction program has been written by R. HEILBRONNER-STÜNITZ.

The *isostatically equivalent water column* is derived assuming point-wise, Airy-type isostatic compensation (discussion in SCLATER and CHRISTIE, 1980):

$$Y = S[(\rho_a - \rho_s)/(\rho_a - \rho_w)] + W_d$$

where

S = thickness of considered sediments
W_d = water depth
$\rho_a = \rho_m(1 - \alpha T)$ = density of the asthenosphere (see table 3)

[1] Sections localized in troughs systematically overestimate the thinning factors while the opposite occurs for sections localized on highs.

TABLE 3 - Constants adopted in the performed calculations

Lithospheric thickness	h_l	125 km
crustal thickness	h_c	31.2 km
crustal density	ρ_c	2.8 g/cm^3
mantle density	ρ_m	3.33 g/cm^3
sediment density	ρ_s	2.40 g/cm^3
water density	ρ_w	1.03 g/cm^3
coeff. ther. expansion	α	3.4x10-5 °C^{-1}
asthenosph. temperature	T_a	1333 °C
thermal diffusivity	K	0.008 cm^2/sec
thermal conductivity	k	0.0075 cal/deg cm sec
lith. therm time const	t	62.7 Ma.
sediment thickness	S	
final depth	Z	
crustal stretching factor	δ	
sub-crustal stretching factor	β	

5.1.2 *Data set in the Lombardian basin*

Critical factors in a subsidence analysis are (ROYDEN, 1986):
1) duration of rifting;
2) thickness of pre-, syn- and post-rift sediments;
3) water depths at the beginning and at the end of rifting and during drifting, *i.e.* during thermal subsidence.

Duration of rifting

A reliable definition of the rifting time can only be based on the determination of when extensional faulting began and ceased. Non-abrupt thickness variations imply only subsidence differences and not necessarily extension. Based on this premise, I assume that rifting started in the studied area, and in the Lombardian basin, during the Norian (chapter 2.3.1). I thus take 220 Ma before present (HAQ *et al.*, 1987) for the onset of rift-related tectonic activity.

An earlier beginning of the rifting phase (WINTERER and BOSELLINI, 1981 and others) is not supported, at least in the Lombardian Basin, by the required tectonic evidence. The troughs which in the Late Carboniferous and in the Permian controlled the sedimentation of clastic and volcanic deposits are not necessarily directly related to the rifting event (HANDY and ZINGG, in press) but rather related to late- or post-Variscan large-scale strike-slip faulting (ARTHAUD and MATTE, 1977).

A precise dating of the end of rifting is more difficult because tectonic evidence is less clear-cut. However, Toarcian formations seem to seal the underlying normal faults (Fig. 45) (chapter 2.3.1). I assume that normal faulting in the Lombard-

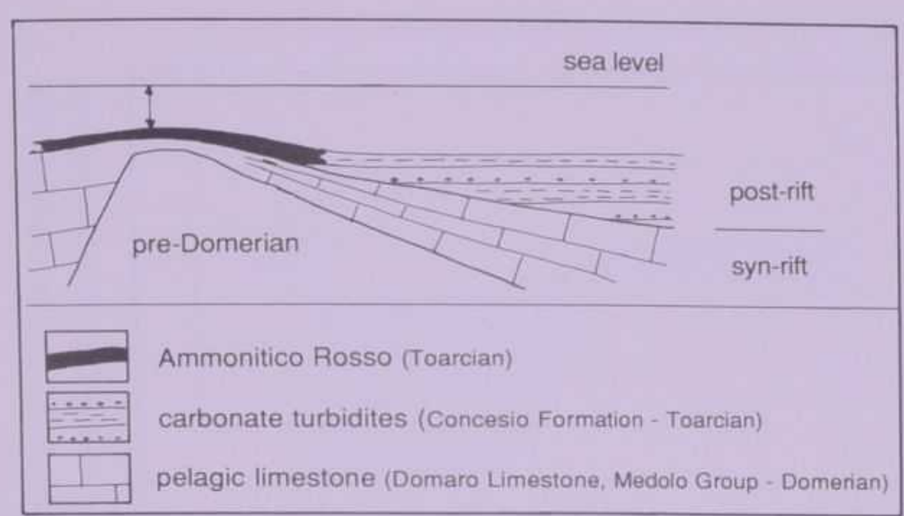

FIG. 45 - Illustration of the relationships between early post-rift formations at the end of the Toarcian: on the highs Rosso Ammonitico-like facies were deposited, while in the basin turbidites are found. The Toarcian formations seem not to be displaced by normal faults thus constraining the end of rifting. The sketch, redrawn after GAETANI (1975), illustrates the situation at the Monte Cavallo high. The real width of the figure is about 10 km.

ian basin came to an end at the beginning of the Toarcian. The proposed age corresponds to ca. 188 Ma before present (HAQ *et al.*, 1987). This age is similar to that proposed by WINTERER and BOSELLINI (1981) and is not in contrast with radiometric dating of ophiolites of the Ligurian-Piedmont ocean.

Surprisingly enough, radiometric dating on ophiolites has not attracted much attention. In fact only relatively few samples, often from tectonized and deformed zones, have been measured. CARPENA and CABY (1984), on the base of zircons fission tracks from plutonic rocks, deduced cooling at 212-192 Ma. U/Pb ages on zircons in albitites gave ages around 161 Ma (OHNENSTETTER *et al.*, 1981). Fission tracks in zircons measured in a gabbro of the northern Apennines similarly produced an age of 165 Ma (BIGAZZI *et al.*, 1972). K-Ar on hornblende of gabbros and diabases have produced ages ranging from 180 to 135 Ma (BERTRAND and DELALOYE, 1976) which have been interpreted as crystallization ages.

The age adopted for the end of rifting is considerably older than the middle Callovian (corresponding to 155 Ma before present) age of the radiolarites, *i.e.* the oldest sediments deposited on oceanic crust (BAUMGARTNER, 1987) [(1)].

I adopt for the rifting stage in the Lombardian basin a duration of ca. 30 Ma (minimal estimate). In accord with the reconstruction presented in chapter 4.2.2 I consider that rifting occurred in one single, continuous phase.

Thickness of syn- and post-rift sediments

My data base consists of stratigraphic columns across the Lombardian basin (Fig. 46) compiled by Otto KÄLIN on the base of published literature and his own observations. Localities of the columns and bibliographic references are given in the appendix.

The columns have been compiled along an E-W profile. This is a consequence of the present day location of the sediments outcrops; it is obviously impossible to say how representative this profile is since there is only very limited control in the N-S direction.

The average thickness of the sediments across the basin is obtained by dividing the surface of syn-rift sediments along the profile by the length of the profile. The average thickness of syn-rift sediments, not corrected for compaction, is:

$t_{syn\text{-}rift} = 3220$ m.

The thickness of post-rift sediments is fairly constant all across the basin and is estimated at 250 m.

The measured thickness of the sediments has also been corrected for compaction (chapter 5.1.1). The thicknesses of the syn-rift succession corrected for compaction and for the effect of the overlying water column (see below), are given in figure 46.

Water depths at the beginning, at the end of rifting and during the drifting stage

Deposition in the Lombardian basin took place near sea level for the whole pre-rifting stage. Water depths in the pre-rift stage are thus fairly well constrained.

Water depths at the end of rifting are more difficult to estimate since sedimentation took place below the photic zone and direct control is poor.

An indirect estimate of water depth at the end of rifting can be attempted considering the Toarcian sediments, *i.e.* the first post-rift formations (chapter 2.3.1). A relic relief persisted in the

(1) It is still not clear if this contradiction is real or not. If the first appearance of oceanic crust is significantly younger than the assumed age for cessation of normal faulting in the Lombardian basin, then, similarly to what postulated by FROITZHEIM and EBERLI (in press), a lateral shift of the zone undergoing rifting could be envisaged. FROITZHEIM and EBERLI (in press) found in the Austroalpine that rifting went on until the Toarcian mainly in an internal part of the margin (the Central Austroalpine). From the Toarcian to the Middle Jurassic, rifting went on along a more distal part of the margin (the Lower Austroalpine) while the Central Austroalpine experienced no further extension.

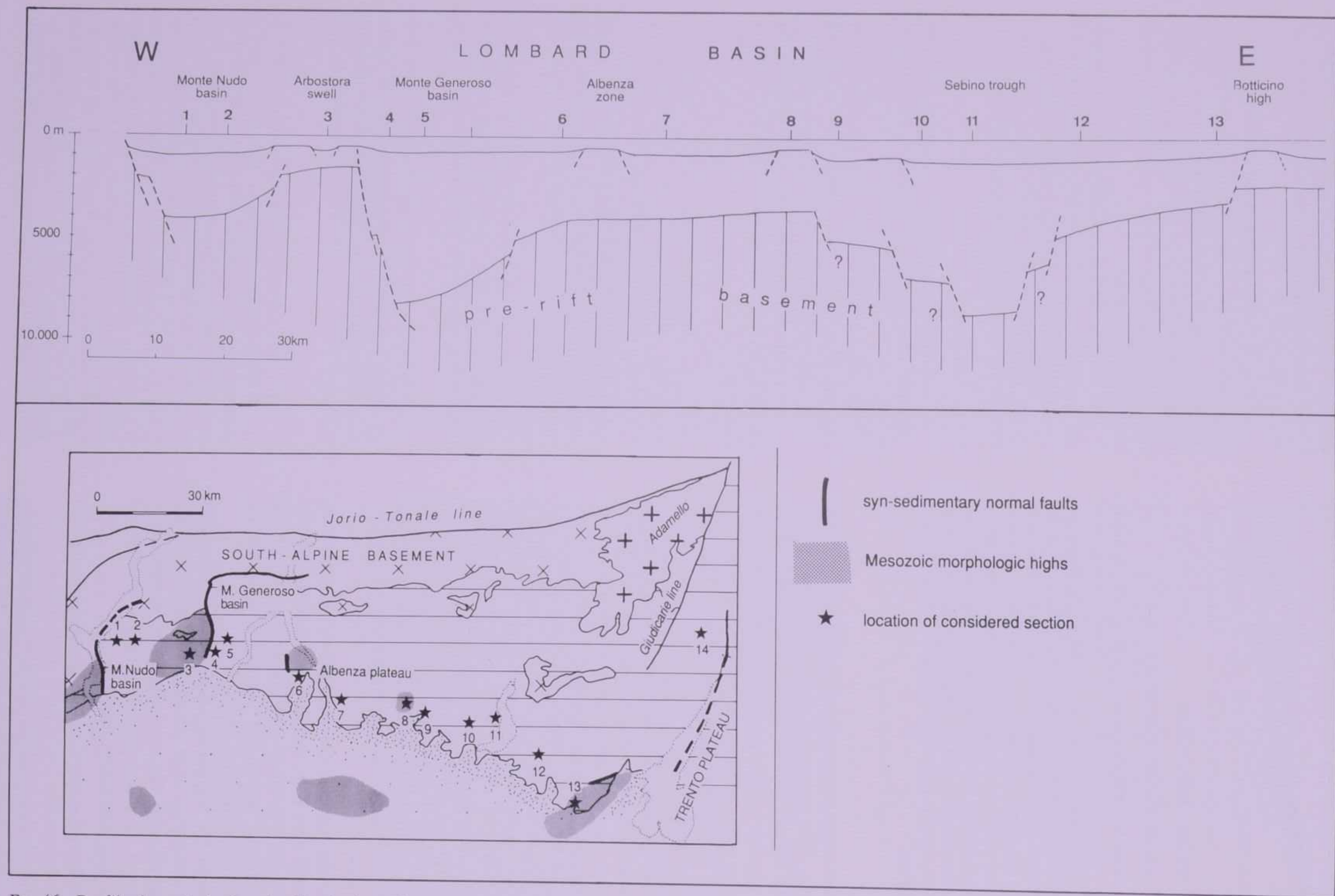

FIG. 46 - Profile through the Lombardian basin at the end of rifting (Toarcian) before the deposition of the Rosso Ammonitico Lombardo and of the turbiditic formations (see text). Thickness of syn-rift sediments (from the Dolomia Principale to the siliceous limestone) is corrected for compaction. Numbers refer to the stratigraphic columns used for the compilation. In the lower part of the figure the locations of the columns are given. Redrawn after a compilation by O. KÄLIN.

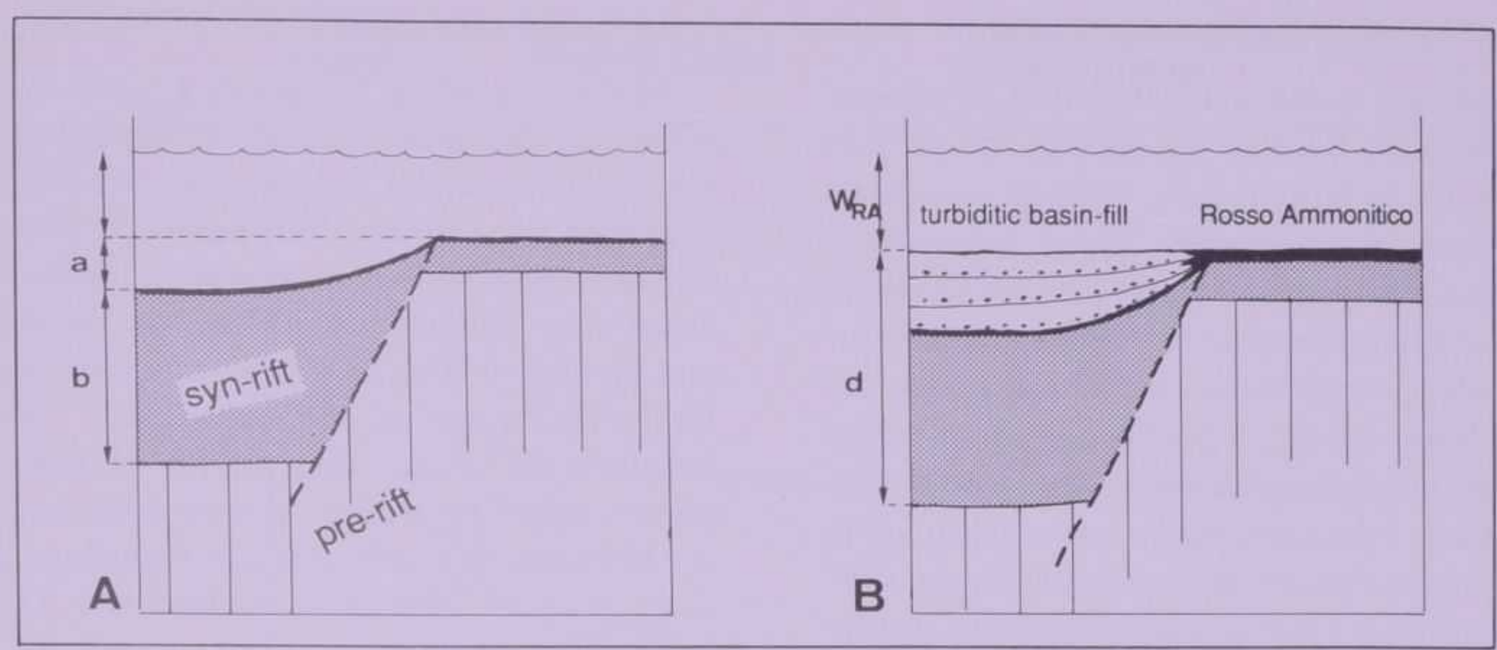

FIG. 47 - Sketch illustrating the method used to reconstruct the sea-floor morphology at the end of rifting (see text for explanation). The sketch is not to scale.

Toarcian, and partly later, after the cessation of extensional faulting. On the highs nodular limestones in Rosso Ammonitico-like facies were deposited while calcareous turbidites are found in the intervening basins (Figs. 4 and 47). The sea-floor morphology was more or less smoothed before the Callovian when the radiolarites, which show fairly constant thicknesses across the whole Lombardian basin, were deposited. The idea, suggested and developed by O. KÄLIN, is then to estimate the depth of deposition of the Rosso Ammonitico facies on the highs and then reconstruct the sea-floor morphology relative to the highs removing the turbiditic basin-fill.

The absolute depth of the highs on which the Toarcian Rosso Ammonitico was deposited (W_{RA} in Fig. 47) can be sedimentologically estimated to be between 250 and 500 meters.

JENKYNS (1974) argues that deposition of a Rosso Ammonitico like the one found in Lombardy should have taken place below the photic zone but not deeper than 800-1000 meters. The upper bracket is given by the absence of stromatolites and other features indicative of shallow water. The lower bracket is given by the common presence at the sea-floor of organisms with aragonitic hard parts (ammonites and bivalves). The given estimate is well in agreement with the recent interpretation of the black shales which underlie the Rosso Ammonitico in several places of the Lombardian basin. These shales have been interpreted as being deposited within the minimum oxygen zone of the Jurassic ocean (JENKYNS, 1988).

To reconstruct the depth of the basins with respect to the highs, the turbiditic basin fill has to be removed and replaced by an equivalent water column.

In more detail this is done performing the following calculations:

i) calculate the decompacted thickness of the syn-rift succession plus the post-rift sediments underlying the radiolarites (mainly the Rosso Ammonitico and the turbidites) (segment d in Fig. 47 B);
ii) calculate the decompacted thickness of the syn-rift sediments at the end of rifting, *i.e.* before the deposition of the Rosso Ammonitico and of the basin-fills (b in Fig. 47 A);
iii) the height difference of the top of the considered column with respect to the highs on which the Rosso Ammonitico was deposited is eventually obtained subtracting b from d (a in Fig. 47 A).

A profile can thus be constructed representing the sea floor morphology at the end of rifting (Fig. 46). On the same profile the syn-rift succession is represented.

Water depths during the drifting stage cannot be estimated on a sedimentological or paleontological basis. Sedimentation took place well below the photic zone so that no direct control is possible. The utilization of paleo-oceanographic indicators like depths of the CCD and/or of the ACD derived from other oceans is not very reliable. The local control on the depths of these horizons makes their extrapolation in time and in space very insecure (discussion in BAUMGARTNER, 1987).

5.1.3 *First constraints on the vertical evolution of the Lombardian basin: initial subsidence and final depth*

On the base of the tectonic analysis (chapter 4.5.2) and on the sedimentary data presented above some constraints can be given on the vertical evolution of the western Lombardian basin. I shall now reconstruct the initial subsidence and the final depth of the water-loaded western Lombardian basin.

Initial subsidence

The first step to calculate the initial subsidence of a sedimentary basin is to remove the syn-rift sediments and to replace them with an isostatically equivalent water-column. These calculations have been carried out separately for all considered columns. Adding the local bathimetry to the single equivalent water columns and compiling the resulting water depths, a profile is then obtained for the sediment-free Lombardian basin at the end of rifting. I then calculate a water depth value averaged across the whole Lombardian basin dividing the surface of the water loaded basin on the profile by the length of the profile. This is the initial subsidence for the water-loaded Lombardian basin.

The values obtained lie between 1.4 and 1.9 km (stippled bar in Fig. 48). In this bracket the uncertainties in the water depths of the Rosso Ammonitico deposits and different ways of carrying out the decompaction are taken into account.

Estimate of the final depth

The control on post-rift paleo-depths is far from being good enough to enable a direct estimate of the final depth of the Lombardian basin. This, however, can be estimated using the stretching factor deduced in chapter 4.5.2. In fact, the final depth of a basin depends, at first approximation, only from the crustal stretching factor and is independent from the mode of extension in the subcrustal lithosphere.

When initial elevation is at sea-level, the final depth of the water-loaded basin is (Le Pichon and Sibuet, 1981):
$Z = C\,(1-1/\delta)$ where
δ = crustal stretching factor
and

$$C = h_c \frac{(\rho_m - \rho_c)\,[1 - (\alpha/2)T_a(h_c/h_l)]}{\rho_a - \rho_w}$$

The constant C physically corresponds to the elevation reached by the top of the asthenosphere with infinite stretching of the lithosphere and is derived from mid-oceanic ridges where the asthenosphere nears the surface. The other constants and symbols adopted are given in table 3.

The final depth of the water-loaded basin with a crustal stretching factor of 1.5 is:

$$Z = C(1-1/\delta) = 2.53 \text{ km.}$$

This is the depth reached by the water-loaded basin at infinite time (Fig. 48).

5.1.4 *Subsidence curves*

In the preceding chapters I have derived estimates for the upper/middle crustal stretching factor, for the initial subsidence and for the final depth of the Lombardian basin [1]. These quantities are obviously not independent from each other since they reflect the same extensional history and geometry. They can therefore be used to constrain the shape of the subsidence curve of the Lombardian basin. The shape of the curve depends on the geometry of extension at the different lihospheric levels and can therefore give some clues on how extension below the Lombardian basin took place.

The Royden curves

One of the currently used models is the depth-dependent, instantaneous stretching model (Royden, 1986; and references therein). From the equations given in this paper, I take the subsidence curves of figure 48 (constants are made compatible with those chosen by Cochran; see table 3; the program has been written by Martin Casey). The curves (1 to 4 in Fig. 48) have been constructed maintaining constant the final depth derived from the crustal stretching factor. All the curves have therefore the same crustal stretching factor δ. The shape of the curve moving backwards away from the final depth depends on both the crustal and the subcrustal stretching factor. The curves plotted in the diagram represent the subsidence curves for different subcrustal stretching factors β (values in table 4). Curves from 1 to 4 depict the transition from very a hot extension ($\delta \ll \beta$) to a very cold one ($\delta \gg \beta$). Curve 3 is representative of a more or less uniform stretching situation. It is apparent from figure 48 that curves 1, 2 and 3 are not compatible with the initial subsidence values (stippled bar) derived from sedimentary record. Only curve 4 fits the reconstructed initial subsidence values. Curve 4 represents a very asymmetric geometry of extension with practically no subcrustal stretching (Table 4). This would be compatible only with a simple-shear style of extension with the crust of the Lombardian basin being detached from the underlying unstretched lithosphere by a major low angle normal fault. As will be discussed in chapter 5.2.2 this geometry is not plausible.

I conclude from the presented discussion that none of the Royden curves seems to really fit the

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[1] While the initial subsidence has been derived from the sedimentary record, the final depth has been deduced on the base of the crustal stretching factor.

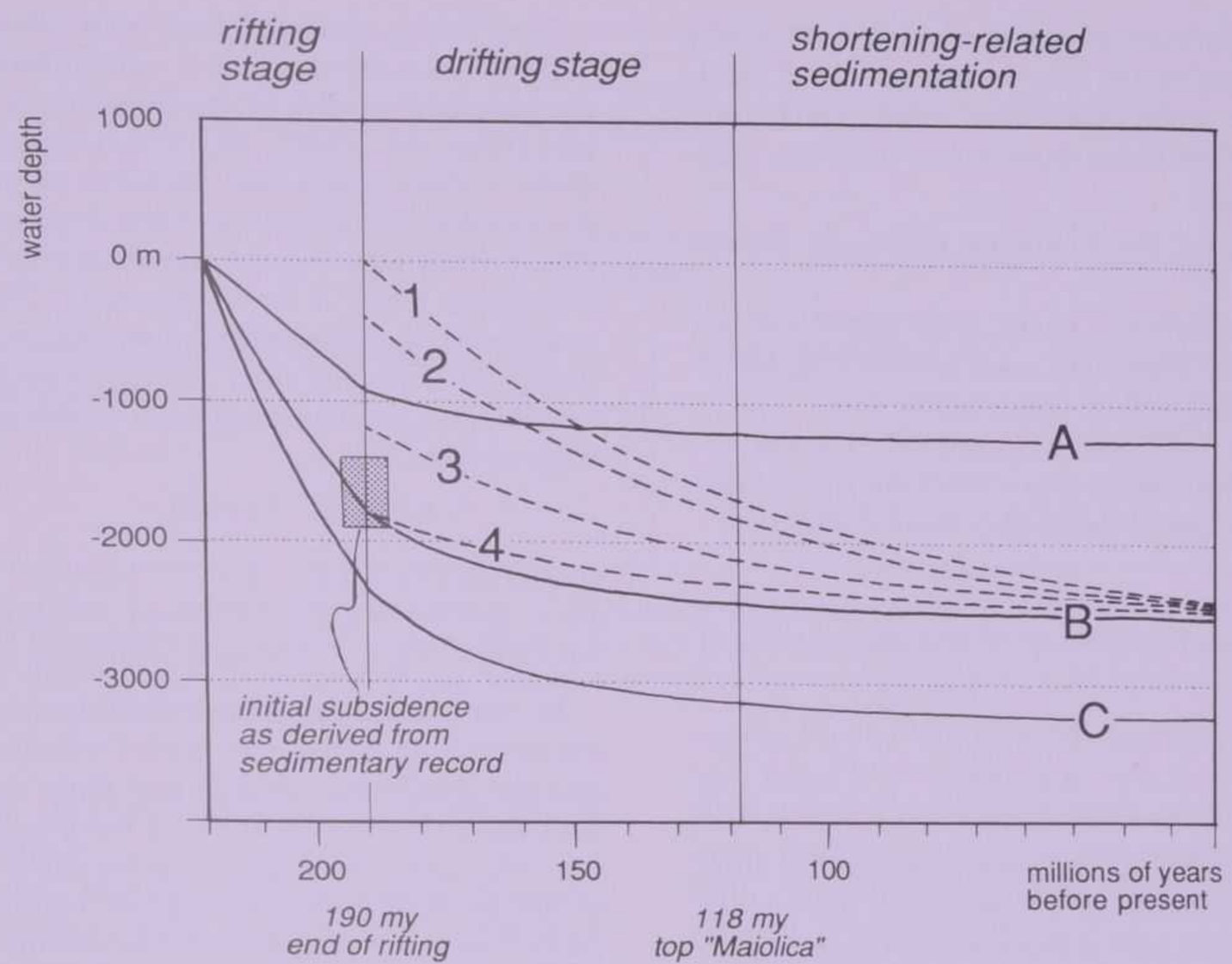

FIG. 48 - Subsidence curves for the Lombardian basin. The stippled bar represents the range of possible initial subsidence values as derived from the sedimentary record. The dashed curves are subsidence curves assuming an instantaneous rifting (ROYDEN model) and are constructed in such a way to be compatible with the independently-derived final depth. Values of δ and β are given in table 4. Only a very cold extension (δ»β) would be compatible with the initial subsidence values. Solid curves depict a uniform stretching situation with two-dimensional heat loss (COCHRAN model) for various stretching factors. Curve B is compatible with the tectonically-derived stretching factor and predicts initial subsidence values compatible with the ones derived from the sediments.

TABLE 4 - Numerical values for the crustal (δ) and subcrustal (β) stretching factors, for the initial subsidence (Si) and final depth (Z) of the subsidence curves shown in Fig. 48

ROYDEN CURVES					COCHRAN CURVES				
#	δ	β	Si	Z	#	δ	β	Si	Z
1	1.5	13.3	0	2500	A	1.2	1.2	923	1273
2	1.5	3.70	400	2500	B	1.5	1.5	1791	2546
3	1.5	1.5	1200	2500	C	1.75	1.75	2347	3274
4	1.5	1.04	1800	2500					

data. This is mainly a consequence of the assumption of instantaneous stretching at the base of the model [1].

Cochran curves

The model presented by COCHRAN (1983) differs from that of ROYDEN (1986) in that it considers heat loss not only on the vertical but also on the horizontal dimension. This leads to an amplification of the effects of thermal subsidence which becomes significant already during rifting stages longer than 10 Ma.

The subsidence curves A, B, C (kindly provided by J. COCHRAN) shown in figure 48 represent the vertical motion of a point in the center of a basin (box C of the COCHRAN, 1983 model) extending with an overall pure shear geometry. Three curves are given for different stretching factors (Table 4). Curves A and C are not only incompatible with the stretching factor observed at the surface, but also predict initial subsidence values in contradiction to the observed ones (stippled bar in Fig. 48). I consider curves A and C not to be applicable to the Lombardian basin. Curve B on the

[1] This also means that the ROYDEN model should be used only for very short rifting stages. Otherwise stretching factors will be reconstructed which are generally larger than those actually occurring.

contrary, which was constructed with a stretching factor similar to the one observed in the field, predicts an initial subsidence value which falls well within the range determined from the sediment record.

Summing up, the following points are important:

a) the subsidence curve constructed for an homogeneous stretching with a stretching factor of 1.5 seems to well fit the available data;

b) lateral heat loss is important, *i.e.* the assumption of instantaneous stretching is not valid for rifting stages of the kind described above.

5.1.5 Geologic implications of the analysis of subsidence

Implications for the sedimentary history of the Lombardian basin

The Southern Alps never reached the final depth reconstructed in the preceding sections. In fact convergent movements along the Austroalpine/South Penninic active margin caused the onset of terrigenous flysch sedimentation in the middle Cretaceous (CASTELLARIN, 1976). I consider the Barremian pelagic limestones of the Maiolica Formation to be the last lithology truly representative of the passive margin evolution. Thermal subsidence thus went on only for 72 Ma after the end of rifting. The depth reached at this time by the water-loaded basin can be read from COCHRAN's curve (B in Fig. 48) and is 2460 m [1].

Adding the syn-and post-rift sediments to the calculated final depth, and correcting for the associated loading effect, the actual water depth at time of deposition of the Maiolica turns out to be between 1200 and 1500 depending on the densities used.

In detail:

$W_{72} = W - S\,(\rho_a - \rho_s) / (\rho_a - \rho_w)$

ρ_a = density of the asthenosphere

W_{72} = water depth of the water-loaded basin in the Barremian

$\rho_a = \rho_m\,(1 - \alpha T_a) = 3.1877$ g/cm³.

using a mean density of 2.4 g/cm³ for the sediments

W = 1182 m

with a density of 2.6 g/cm³.

W = 1506 m.

[1] Since thermal relaxation is at this time substantially accomplished, the depth at 72 Ma is, at first approximation, independent from the geometry of stretching.

The values obtained here are lower than the 3000 m reconstructed in a very indirect way for the paleobathymetry of the Maiolica by WINTERER and BOSELLINI (1981). My reconstructed depth is more realistic since the Maiolica of this part of the Lombardian basin has been deposited on a ca. 20 km thick continental crust (as required by the stretching factor I have derived). The 3000 m postulated by WINTERER and BOSELLINI (1981) would be expected only on an oceanic crust or, possibly, on a continental crust thinned to a much larger degree.

Implications for lower crustal and subcrustal extension below the Lombardian basin

In view of the several uncertainties regarding the stretching factor, the initial subsidence values and the subsidence models and their equations, it is obvious that the estimates I have been presenting only have a semi-quantitative value. However, on the basis of these preliminary results it is concluded that the vertical evolution of the Lombardian basin seems to be best described by a uniform stretching geometry with a stretching factor of 1.5. This implies that the stretching factor derived for the upper/middle crust is a stretching and therefore thinning factor also of the lithosphere underlying the basin.

5.2 EXTENSION IN THE LOWER CRUST AND LITHOSPHERE OF THE LOMBARDIAN BASIN: A DISCUSSION

5.2.1 *Extensional models*

The mode of continental extension leading to the formation of a passive continental margin is presently highly debated. Two end-member models face each other: the pure shear (homogeneous stretching) and the simple shear model (non-homogeneous stretching) (Fig. 49).

In the pure shear model (MCKENZIE, 1978; ROYDEN, 1986 and others) the extension observed at the surface is equal to that of the underlying crust and lithospheric mantle; the thinning caused by extension is therefore constant on a vertical section throughout the lithosphere. The whole lithosphere is thinned with a pure shear geometry. Deformation in the upper crust is discrete [2], while lower crust and lithospheric mantle

[2] The detailed geometry of faulting in the upper crust is not relevant in MCKENZIE's model. As long as the extension at a crustal scale has a pure shear geometry, every geometry is possible. This means that asymmetric fault patterns (like domino) by no means imply an asymmetric extension at depth.

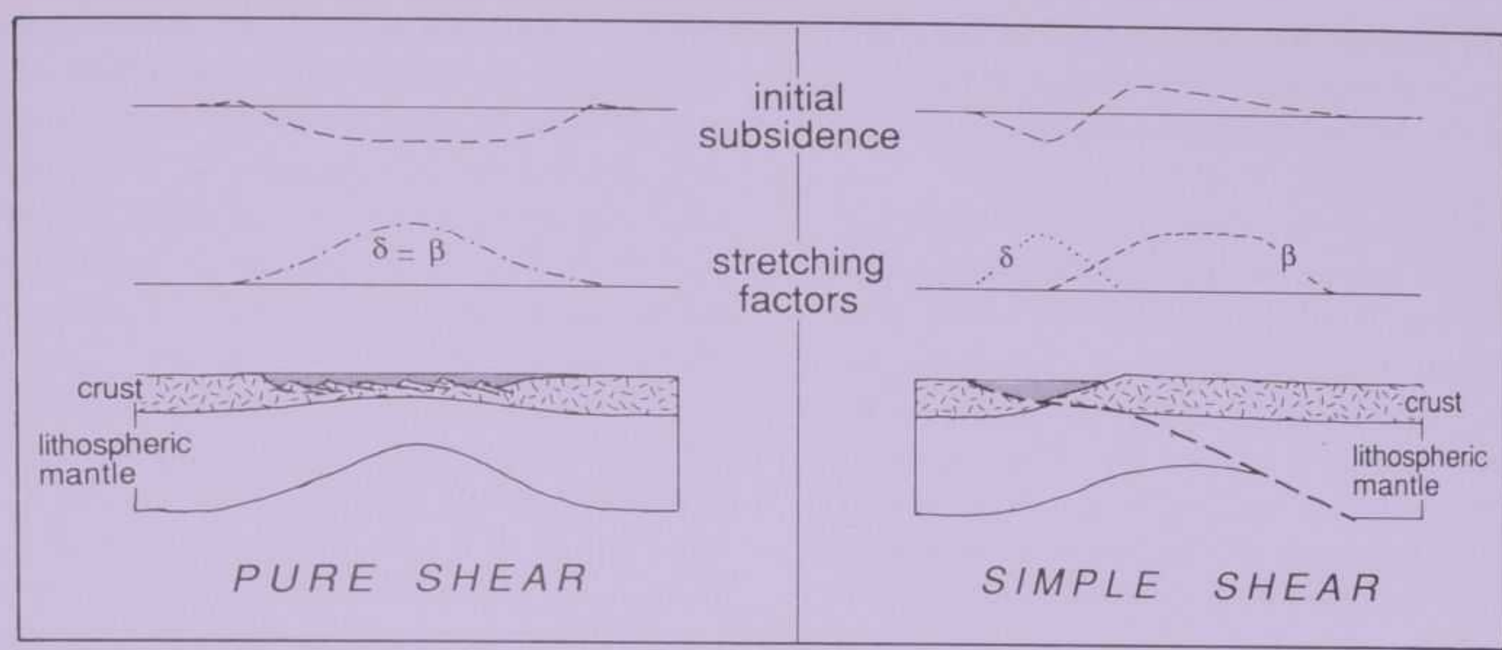

FIG. 49 - Pure shear vs. simple shear. Schematic representation of the fault geometry, of stretching factor variations and of initial subsidence for the two models. Fault geometry in the upper crust in the pure shear models has been intentionally drawn as asymmetric although still with an overall pure shear geometry. Modified after WERNICKE (1985), BUCK *et al.* (1988) and ISSLER *et al.* (1989).

behave like rubber bands and deformation is therefore diffuse.

According to the simple shear model (RAMSAY, 1980; WERNICKE, 1985; LISTER *et al.*, 1986 and others), on the contrary, extension in the upper crust is laterally transferred by a major low angle normal fault cutting through the entire lithosphere. Segments of extended, and therefore thinned crust, overlie columns of lithospheric mantle which preserve their original thickness and viceversa (Fig. 49).

Since the early eighties, a number of models have been proposed which somehow combine the models mentioned above: the lithosphere is subdivided in an upper part where extension is controlled by a major low angle normal fault and has therefore a simple shear geometry, and a lower part where deformation is diffuse and has a pure shear geometry. The transition between the two layers has been set at different levels by different authors: *e.g.* between upper and lower crust (LE PICHON and BARBIER, 1987; GIBBS, 1987) or at the base of the crust (LISTER and DAVIS, 1989).

The thermal and isostatic behaviour of the two end-member models has been recently modeled (VOORHOEVE and HOUSEMANN, 1988; BUCK *et al.*, 1988; ISSLER *et al.*, 1989). The results have been compared with real situations in order to try to recognize the mechanisms and geometries actually going on in areas of active rifting (for instance in the Red Sea-Gulf of Suez region, JACKSON *et al.*, 1988; BUCK *et al.*, 1988). Zones of active rifting have obviously the advantage of allowing a control on important geophysical quantities like crustal thickness, heat flow, movements of the earth surface etc. It is interesting to note how carefully investigated real situations are generally not explainable with simple models. They rather require combinations of the end-members, changes with time etc. (BUCK *et al.*, 1988). DUNBAR and SAWYER (1989) well showed how important variations in style and amount of extension can occur also along the same passive continental margin.

5.2.2 *Geologic evidence from the western Southern Alps. Discussion of possible models for the Mesozoic extension*

As a result of their complex and polyphased history of deformation, different lithospheric levels are presently exposed in the western Southern Alps from the upper crustal horizons of Lombardy to the lower crustal to upper mantle levels of the Ivrea zone. This offers the possibility to investigate continental extension at different levels.

The most relevant data derived from the tectonic analysis, from the sedimentary record and from the study of subsidence are here integrated in order to propose a model for the extension of the South-Alpine lithosphere (Fig. 50).

Extension in the upper and middle crust of the western Lombardian basin is controlled by two major, E-dipping normal listric faults: the Lago Maggiore fault and the Lugano fault. The Lugano fault attains a gentle eastward dip at about 8-10 km of depth and can be followed to a paleo-depth of ca. 12-15 km. Deformation along the fault zone remains discrete, *i.e.* concentrated along a well defined band, at a crustal scale. The metamorphic rocks below the fault are not deformed. The Lago Maggiore fault seems to have a similar geometry. To a paleo-depth of ca. 15 km deformation was

therefore discrete at a crustal scale. The two normal faults are however not connected and do not merge into a common detachment horizon (Fig. 50).

Rocks belonging to deeper crustal levels are found, further to the west in the Strona-Ceneri and Ivrea zones (Fig. 1) (*e.g.* HANDY, 1987). The Strona-Ceneri zone consists of paragneisses, schists and orthogneisses and is considered to be a segment of middle crust. The Ivrea zone is made up by metabasic rocks with ultrabasic lenses in the north and paragneisses in the south and is considered to represent a section of lower continental crust (HANDY, 1987). The contact between the Strona-Ceneri and the Ivrea zones is represented by a Mesozoic low angle normal fault (HODGES and FOUNTAIN, 1984; HANDY, 1987).

The fault zone is well defined with respect to the hangingwall (Strona-Ceneri zone). Towards deeper levels the fault zone grades into a zone of diffuse deformation. HANDY and ZINGG (in press) assume therefore a diffuse deformation of the lower crust. In the Ivrea zone itself, conjugate sets of high-temperature shear zones have been described which have been tentatively correlated to the extension leading to the formation of the passive continental margin (BRODIE and RUTTER, 1987). These discrete shear zones accommodate at the scale of the lower crust a diffuse deformation. If the shear zones of the Ivrea zone are actually related to rifting (for a discussion of the problem see HANDY and ZINGG, in press), they would document a diffuse deformation of the lower crust and uppermost mantle during the Mesozoic extension. The crustal normal faults described in the upper and intermediate South-Alpine crust would then splay towards deeper levels and grade into a zone of diffuse deformation which will eventually involve the whole lithospheric mantle. With this model the extension observed in the upper and middle crust would not be transferred laterally and would be accommodated in the underlying lithospheric column. This interpretation well fits wits the results of the analysis of subsidence carried out in chapter 5.1 [1].

I find at the moment no founded objection to this model for the late Triassic to Liassic evolution of the Lombardian basin.

[1] The suggested model is similar to what proposed by LE PICHON and BARBIER (1987) and by LISTER *et al.* (1986).

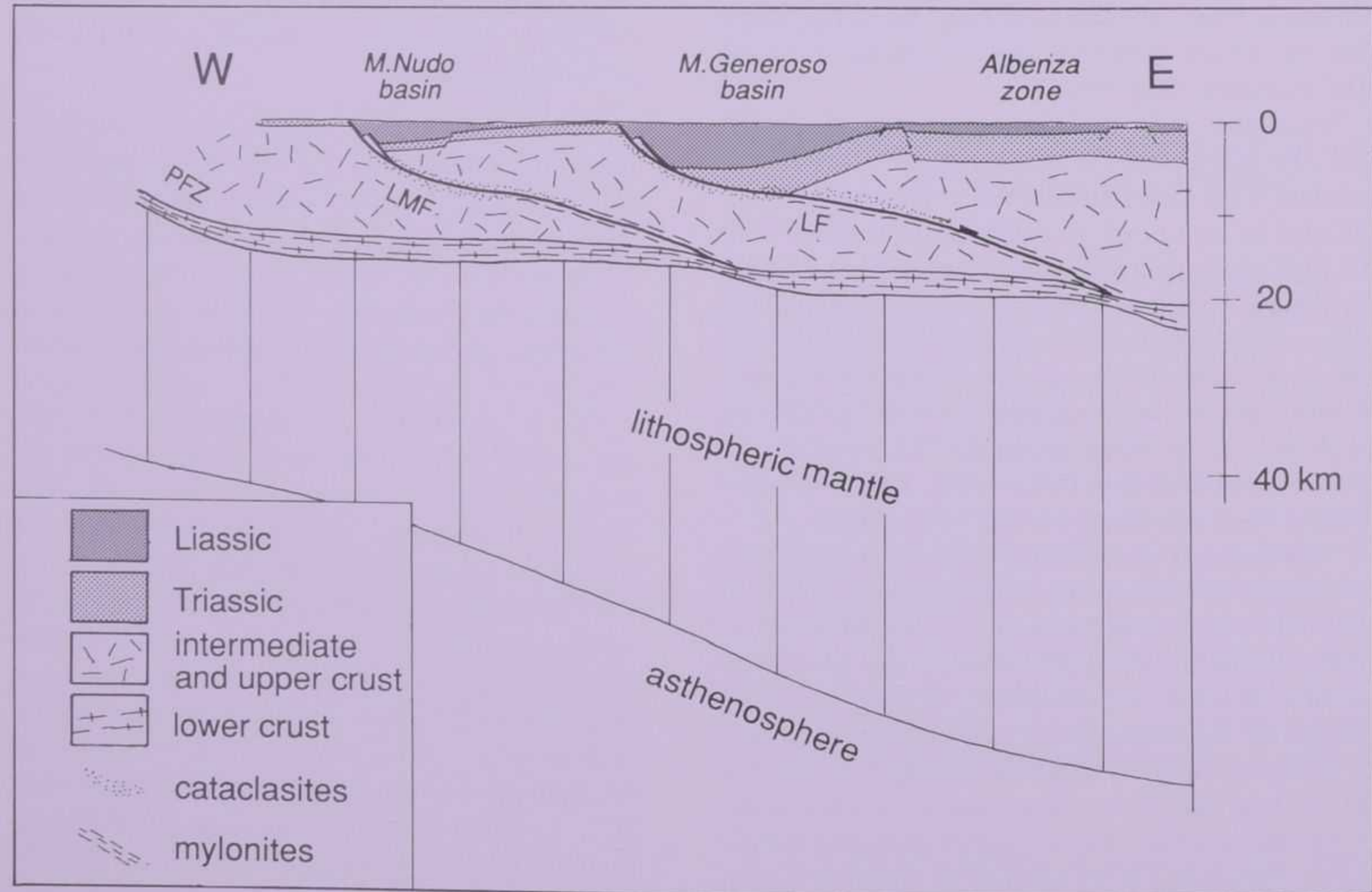

FIG. 50 - Lithospheric profile across the South-Alpine segment of the Apulian passive continental margin. Extension in the crust is accommodated by major low-angle normal faults. The faults splay downwards and deformation in the lower crust and in the lithosphere is diffuse resulting in an overall pure shear geometry. PFZ = Pogallo fault zone; LMF = Lago Maggiore fault; LF = Lugano fault.

Two attempts to apply simple shear (Wernicke-type) models to the Southern Alps have been recently made. LEMOINE *et al.* (1987), mainly on the base of petrographic data from the Ligurian ophiolites, proposed that the opening of the Ligurian-Piemont ocean was controlled by a W-dipping low-angle normal fault. This model provides an elegant explanation for the evolution of the Briançonnais domain of the western Alps but does not seem to be compatible with the sedimentary record in the eastern part of the system (Apulian plate) and with what is known about extension in the middle and lower South-Alpine crust (this work, HANDY 1987). Moreover, according to the geometry proposed by LEMOINE *et al.* (1987), the Lago Maggiore and the Lugano faults would merely represent antithetic faults with respect to the major, west-dipping low angle normal fault. I do have some difficulties in accepting this.

SARTI *et al.* (1989) published a section across the Southern Alps where extension occurs along a E-dipping, low-angle normal fault. Beside internal incongruities, the vertical movements resulting from the proposed configuration are in contrast with the established geologic record. As shown by many investigations (*e.g.* BUCK *et al.*, 1988; ISSLER *et al.*, 1989), the upper plate of a gently dipping low angle normal fault (the Southern Alps in our case), would experience important uplift during rifting and then slowly return to the initial elevation. While some local uplift took in fact place (however mostly concentrated in the westernmost segment of the passive margin; Gozzano swell, KÄLIN and TRÜMPY, 1977), it is obvious that the Southern Alps on the whole, at least from the Lombardian basin eastwards, underwent substantial subsidence.

On the base of the presented discussion I consider the simple shear model not to describe adequately the Triassic to Liassic extension of the Lombardian basin.

APPENDIX

References used in the compilation of stratigraphic columns

Sections 1, 2 and 3
location:
1 = Cittiglio
2 = Arcumeggia
3 = Saltrio
References: BERNOULLI, 1964; WIEDENMAYER, 1963; KÄLIN and TRÜMPY, 1977 with references therein.

Sections 4 and 5
location:
4 = western Generoso trough
5 = central Generoso trough
References: BERNOULLI, 1964; LEHNER, 1952; GNACCOLINI, 1965; WIEDENMAYER, 1980; GAETANI and FANTINI SESTINI, 1978 and own data.

Section 6
location: Corni di Canzo, Val Ceppelline/Monte Cornizzolo
References: CANTALUPPI and MONTANARI, 1969; BERNOULLI, 1964; LICHTENSTEIGER, 1986; GNACCOLINI, 1965a; KÄLIN, unpublished.

Section 7
location: Sogno/Val Cava
References: DESIO, 1929; GAETANI and POLIANI 1978; KÄLIN, unpublished.

Section 8
location: Monte di Nese/Monte Cavallo.
References: ASSERETO and CASATI, 1965; CASATI and GAETANI, 1968.

Section 9
location: Pradalunga/ Monte Rena
References: ASSERETO and CASATI, 1965; KÄLIN, unpublished; GAETANI, 1970; FRANCANI, 1967.

Sections 10, 11 and 12
location:
10 = Val Adrana, S. Fermo, Torrezzo.
11 = Lake Iseo
12 = Val Trompia, Monte Domaro, Noboli
References: VECCHIA, 1949; GAETANI, 1970; CITA *et al.*, 1961; KÄLIN, unpublished.

Section 13
location: Botticino
References: ASSERETO and CASATI, 1965; CASSINIS, 1968; BONI and CASSINIS, 1978; CASSINIS, 1978 and references therein; CASSINIS and VERCESI, 1982; CANTALUPPI and CASSINIS, 1984; KÄLIN, unpublished.

ACKNOWLEDGEMENTS

This Thesis has been started under the tutorship of Daniel BERNOULLI and H.P. LAUBSCHER. Its main goal was the investigation of the Alpine and possibly, of the pre-Alpine history of the area between Lugano and Menaggio. During the development of the work the attention was progressively shifted to the Mesozoic extensional tectonics. As a consequence, the role of H.P. LAUBSCHER became less important. However his way of thinking, his lectures in Basel and the discussions we had, have always been very stimulating and useful. An important part of the Thesis is dedicated to the analysis of the fault rocks formed along the Monte Grona/

Val Grande fault zone. This would not have been possible without the teaching and the direct interest of Stefan SCHMID. His willingness to help has been constant in spite of his and my difficult moments. During the last stages of the Thesis we found an interested and important partner in Alberto CASTELLARIN. His encouragement has been very important to me also in trying to reestablish contacts with the academic world of my country. A similar role has played G.V. DAL PIAZ. Although he was not involved directly in the investigations I thank him for his support and help. One person has remained throughout all these changes: Daniele BERNOULLI. A sincere and respectful relationship has developed already from the first months of my stay in Switzerland. His jokes, his love for the absurd side of reality beside his competence have always been important to me.

A lot of persons have been important during these years, both in Basel and in Zürich. I want here only to mention some of them: thanks to P. LÜDIN, T. LICHTENSTEIGER, M. HANDY, P. KELLER; to Holger STÜNITZ, Niko FROITZHEIM, Neil MANCKTELOW, Eva SAUTER, W. WINKLER, G. SCHREURS, P. RUEGG, Stefano MAZZOLI, Helmi WEISSERT, Stefano BERNASCONI and many others.

Various persons have helped me scientifically and technically: the people of the thin sections lab, A. UHR with his drawings, U. GERBER with his photographs, Martin CASEY and R. HEILBRONNER for their help with the computer, Stefano BERNASCONI with the measurements of the stable isotopes, D. DIETRICH with the quick introduction to the world of the SEM and last but not least, J. COCHRAN who very kindly sent me some of the subsidence curves I have used in this work.

A special thank goes to Otto KÄLIN: his vast knowledge of the South-Alpine sediments as well as his perfectionism have been very useful and taught me a lot. A last thank goes to Prof. G.B. DAL PIAZ for having accepted to publish this Thesis in the Memorie di Scienze Geologiche.

Questa tesi è però dedicata ai compagni di tante bellissime giornate: Lorenzo e tutti gli altri bambini ed adulti dei Monti di Gottro.

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EXPLANATION OF PLATE I

FIG. 1 - Asymmetric sequence in the transition between Riva di Solto Shale and Zu Limestone. Note thickening-upward of limestone layers and disappearance of shaly intercalations. SE of Dasio, Valsolda.

Cunardo Formation

FIG. 2 - Dark chert (white arrow) at the top of the Cunardo Formation. Sample Gi 102, SE of Plesio (coordinates 739.17/100.70).

FIG. 3 - Chert (light-coloured) growing discordantly across the sedimentary lamination. Sample Gi 102, SE of Plesio (coordinates 739.17/100.70); // nichols, scale bar = 1 mm.

FIG. 4 - Graded peloids in peritidal carbonates of the Cunardo Formation. Sample Gi 102, SE of Plesio (coordinates 739.17/100.70); // nichols, scale bar = 1 mm.


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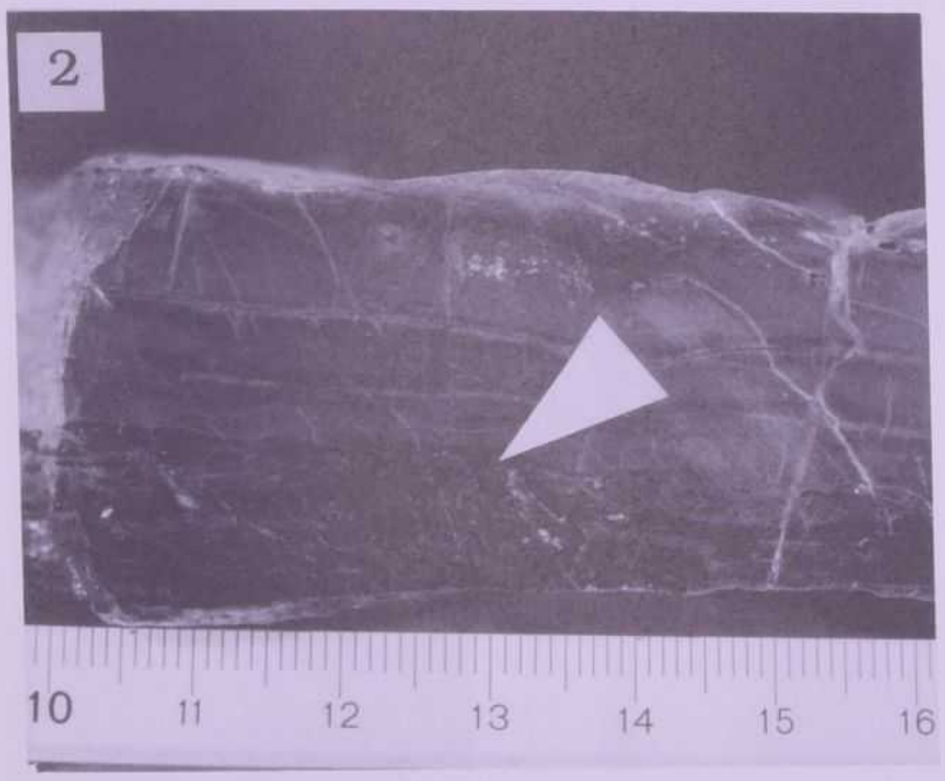
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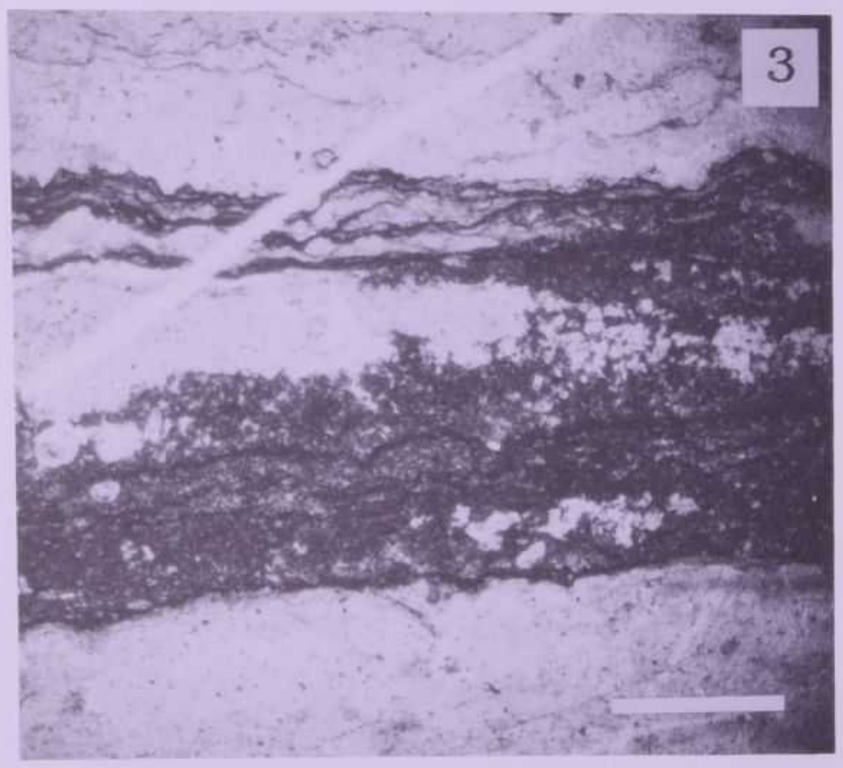
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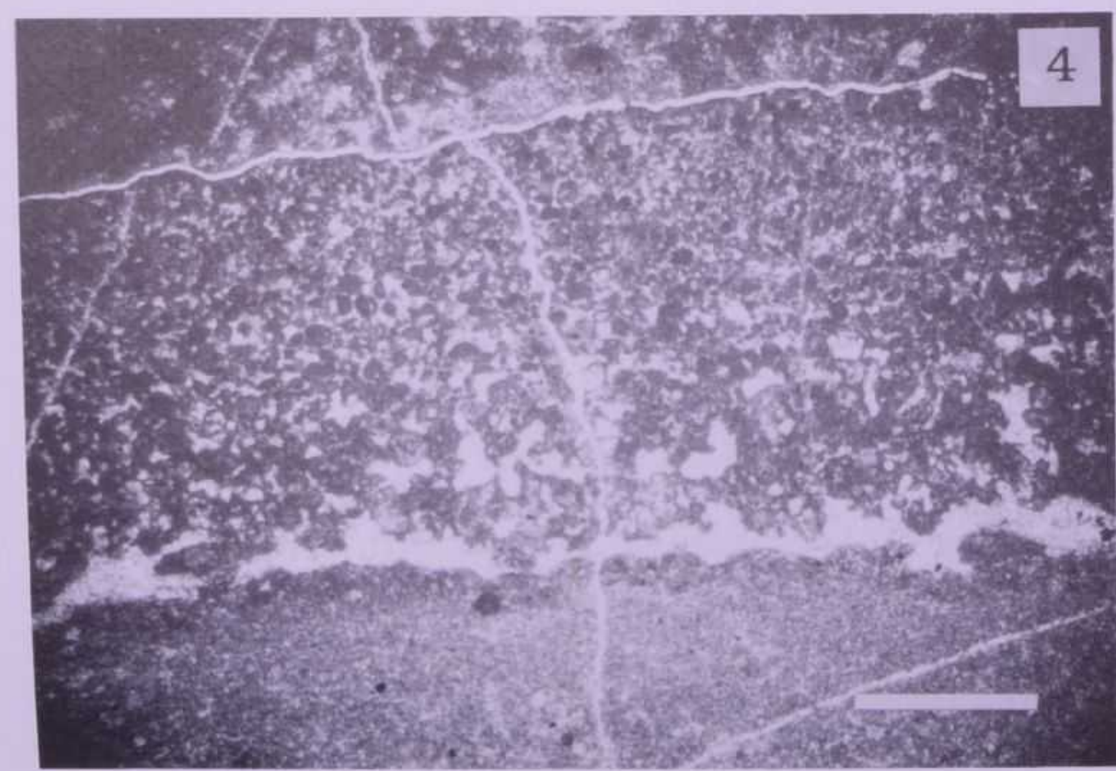
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EXPLANATION OF PLATE II

Ligomena Breccia

FIG. 1 - Macroscopic aspect of the Ligomena Breccia. Note clasts (1) in reddish silt (2). Sample Gi 214, Ligomena.

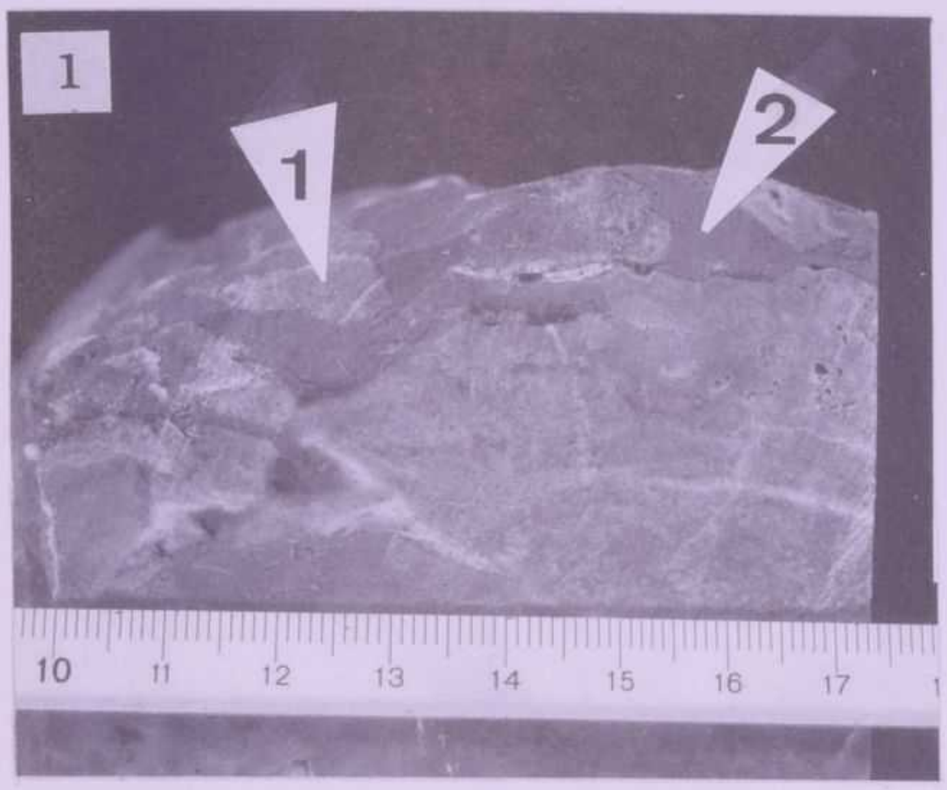
FIG. 2 - The red silt (above) partially geopetally (?) fills cavities left by columnar sparite (below). Sample Gi 214, Ligomena; // nichols, scale bar = 0.5 mm.

FIG. 3 - Columnar sparite nucleates and grows from the geopetal red silt. Relationships are the opposite of Fig. 2. Black arrow indicates growth zonation discordantly cut by new grains. Sample Gi 214, Ligomena; // nichols, scale bar = 0.5 mm.

FIG. 4 - Colourless microsparite with very common inclusions of iron oxides drapes a clast of dedolomitized Dolomia Principale. Arrow points to onlap. Sample Gi 93, Ligomena; // nichols, scale bare = 2 mm.

FIG. 5 - Large dolomite rombohedra grown in the colourless microsparite. The whole section is presently calcitic. Iron oxides are arranged along crystallographic faces of older dolomite rombohedra. The iron oxides are probably derived from the oxidation of bivalent iron incorporated in dolomite or ankerite crystals. Sample Gi 93, Ligomena; // nichols, scale bar = 0.5 mm.

FIG. 6 - Large crystals of columnar sparite growing from a clast. Sample Gi 125, Ligomena; // nichols, scale bar = 0.5 mm.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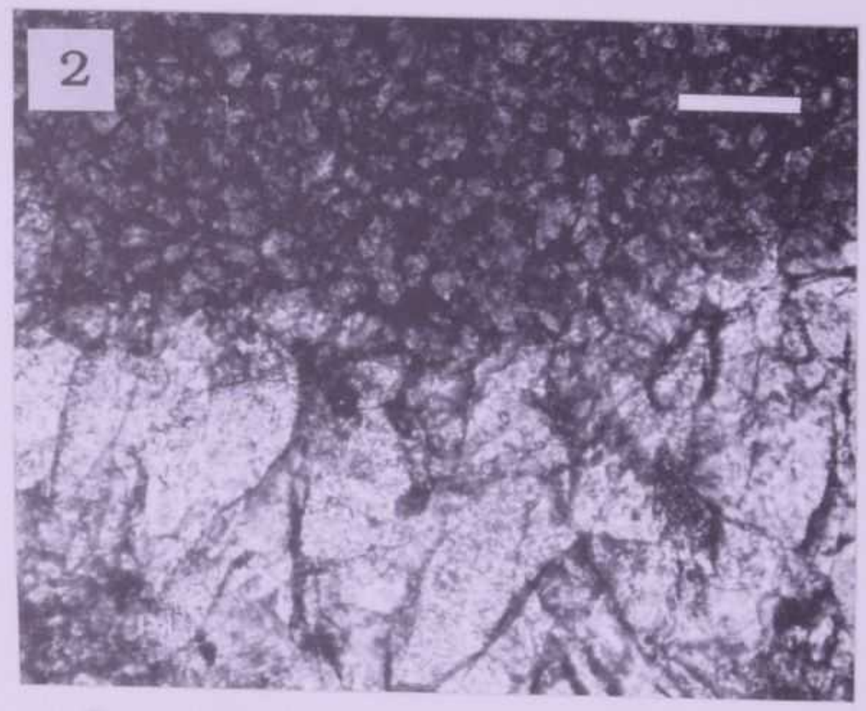
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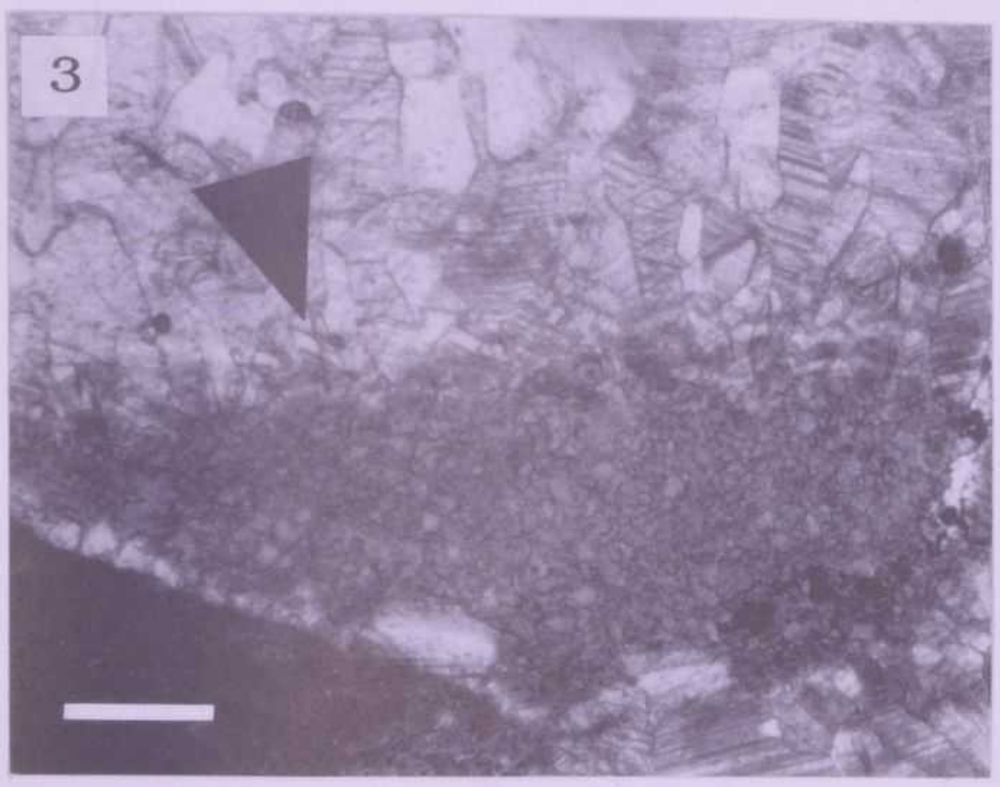
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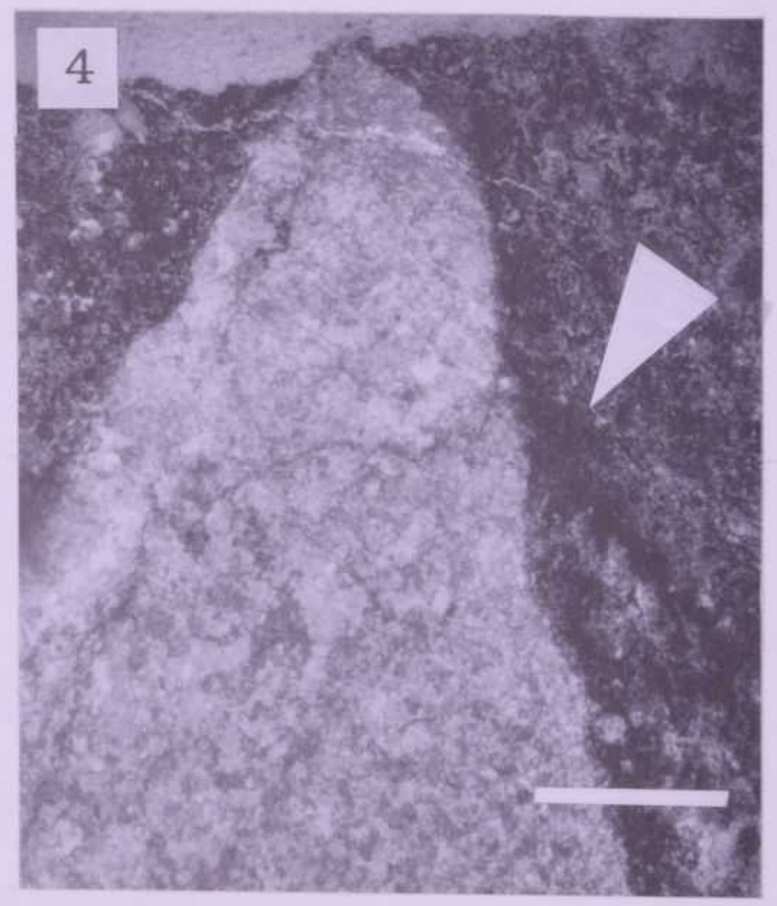
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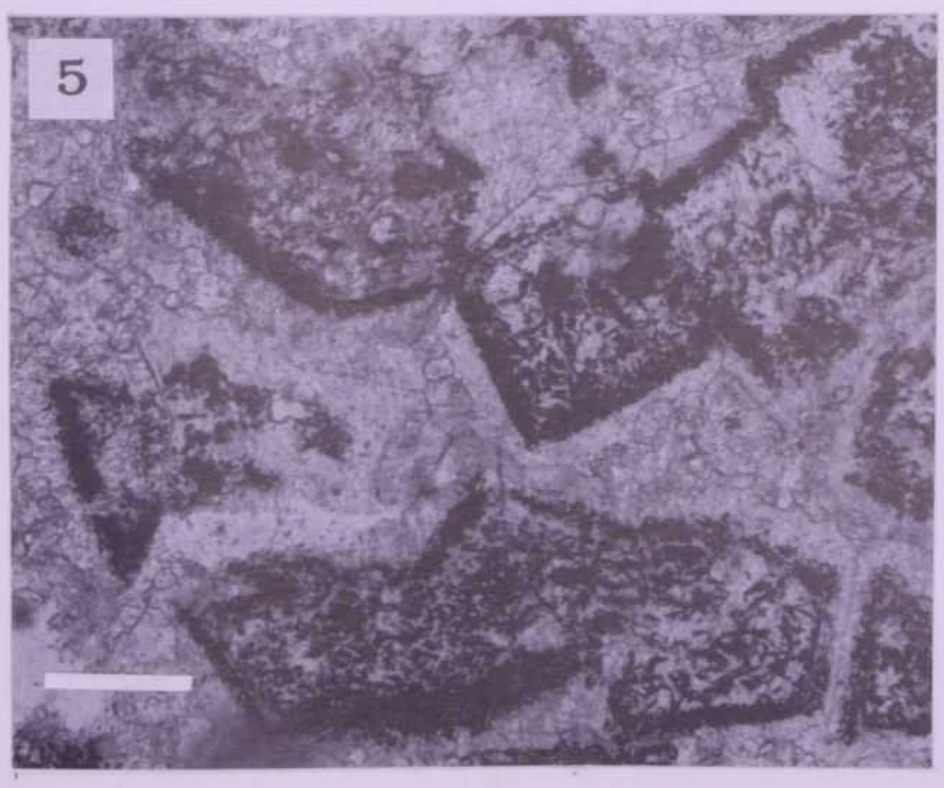
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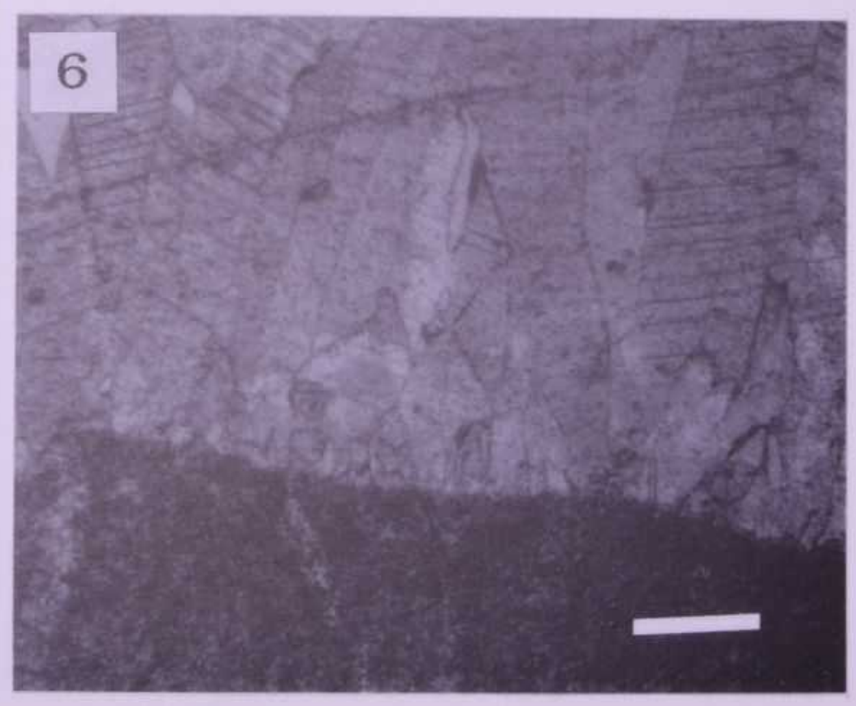
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EXPLANATION OF PLATE III

Zorzino Formation

FIG. 1 - Massive breccia. Light-coloured clast are platform derived. Note hammer on the right for scale. E flank of M. Mugetto, Val Rezzo.

FIG. 2 - Fine-grained dolomitic sand onlapping onto a dm-sized platform-derived block. Massive breccias; S of Monte Pidaggia.

FIG. 3 - Mud-supported conglomerate. Stratified breccias; Colmen dei Carac, upper Val Rezzo.

FIG. 4 - Meter-thick bed of stratified breccias overlying a succession of thinner turbiditic beds. Note very limited erosion at the base of the coarse-grained bed. Cime di Bronzone.

FIG. 5 - Finely laminated dolomitic arenites overlain by a mud-supported breccia. Note very limited erosion at the base of the breccia. Colmen dei Carac, upper Val Rezzo.

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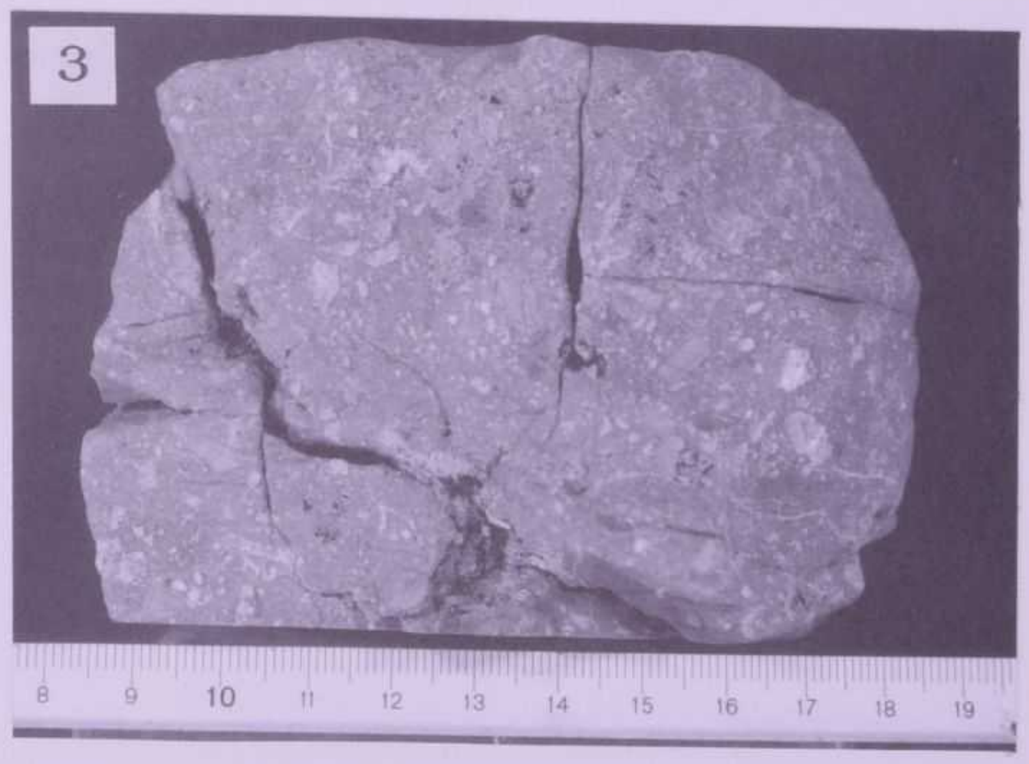
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EXPLANATION OF PLATE IV

Zorzino Formation

FIG. 1 - Graded bed. Clasts are mainly platform-derived. Velzo, Val Sanagra.

FIG. 2 - Slump in thin-bedded turbiditic dolomites. Val Sanagra, road junction to Monti di Gottro.

FIG. 3 - Small-scale syn-sedimentary normal fault in laminated dolomites. Val Sanagra, S of the road junction to Monti di Gottro.

FIG. 4 - Finely laminated dolomites; dark laminae are rich in organic matter. Colmen dei Carac.

Zu Limestone

FIG. 5 - Calcareous sand eroding plane-parallel lamination (storm-deposit). Monti di Nava, Sasso di S. Martino.


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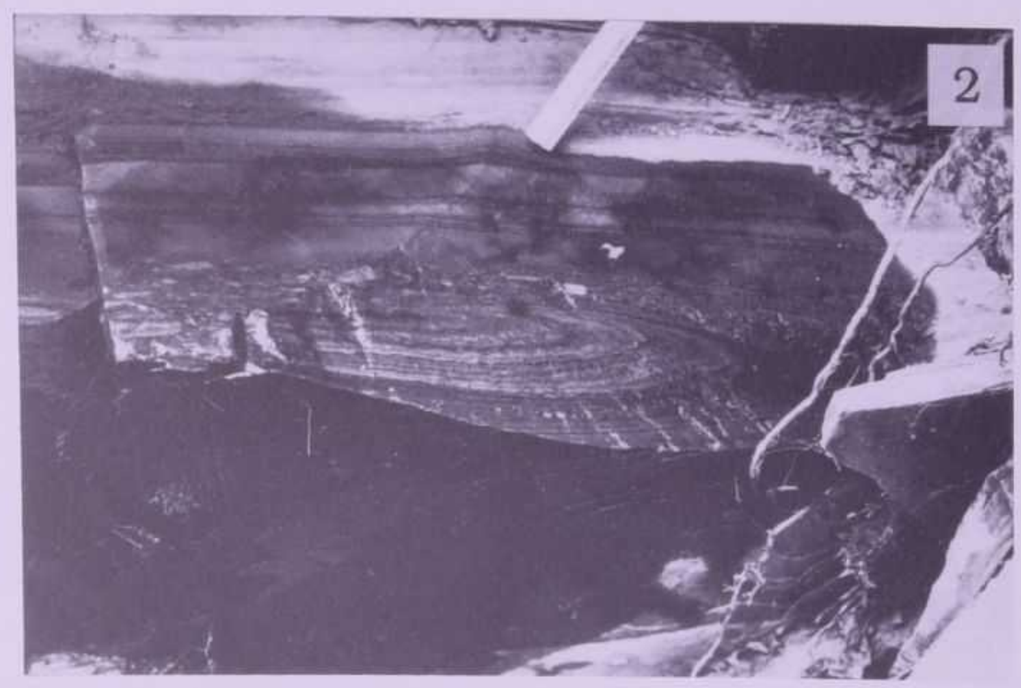
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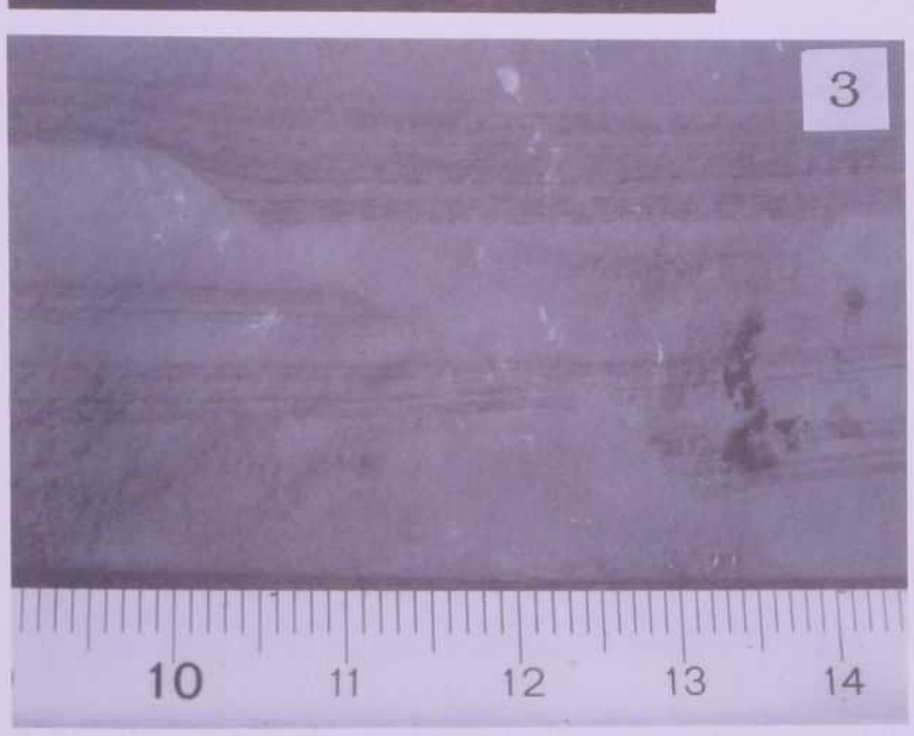
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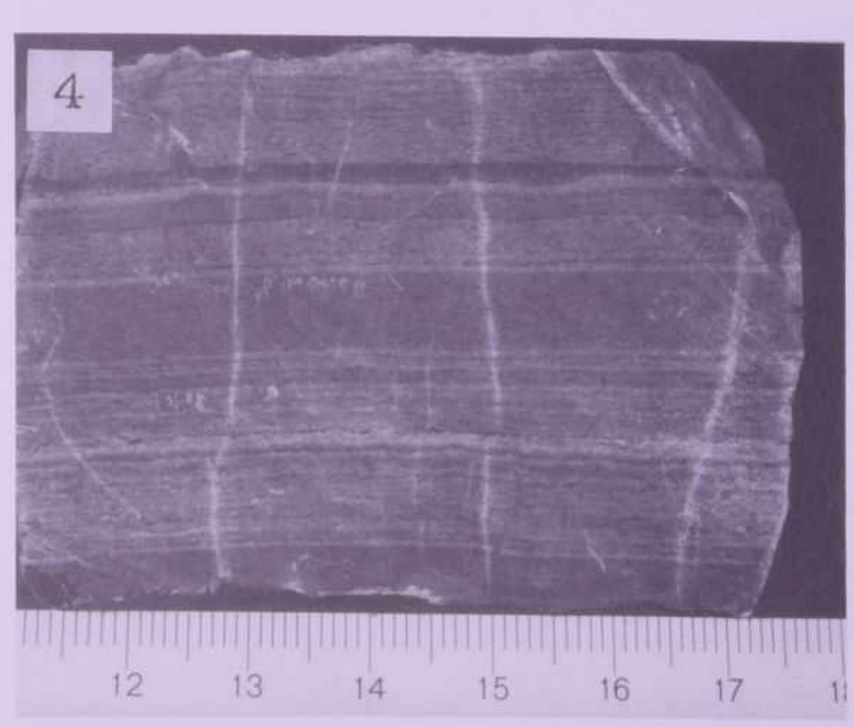
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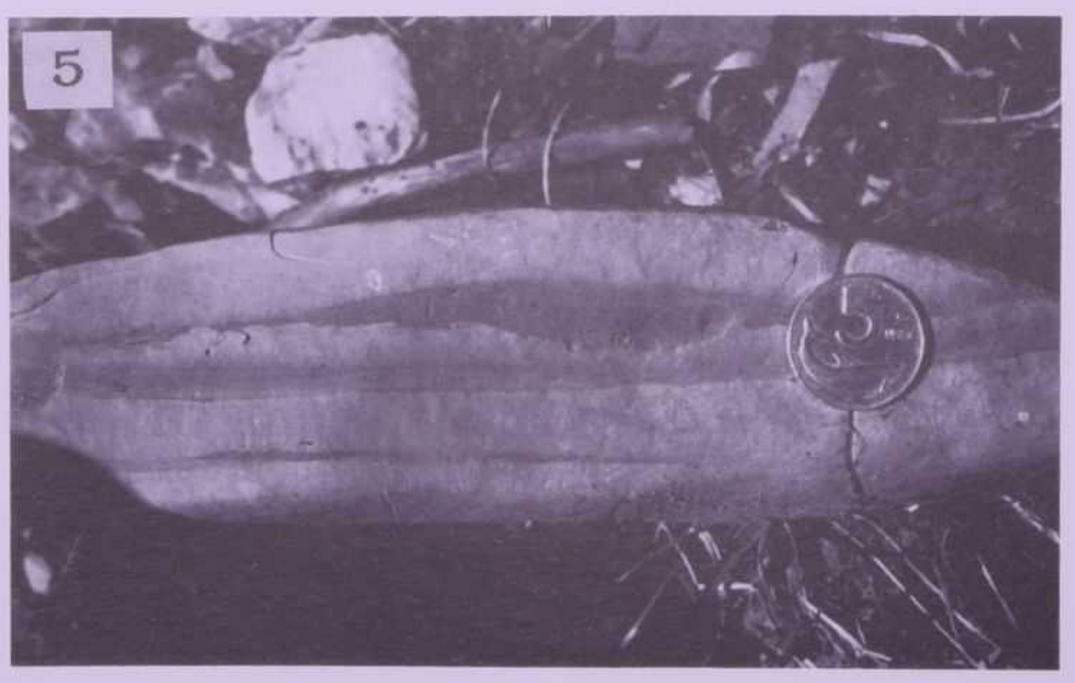
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EXPLANATION OF PLATE V

Lower greenschist mylonites

FIG. 1 - Typical macroscopic aspect of lower greenschist mylonites. Plane of photograph is subhorizontal; W to the left. Note sinistral shear bands parallel to the pencil. Camaggiore profile.

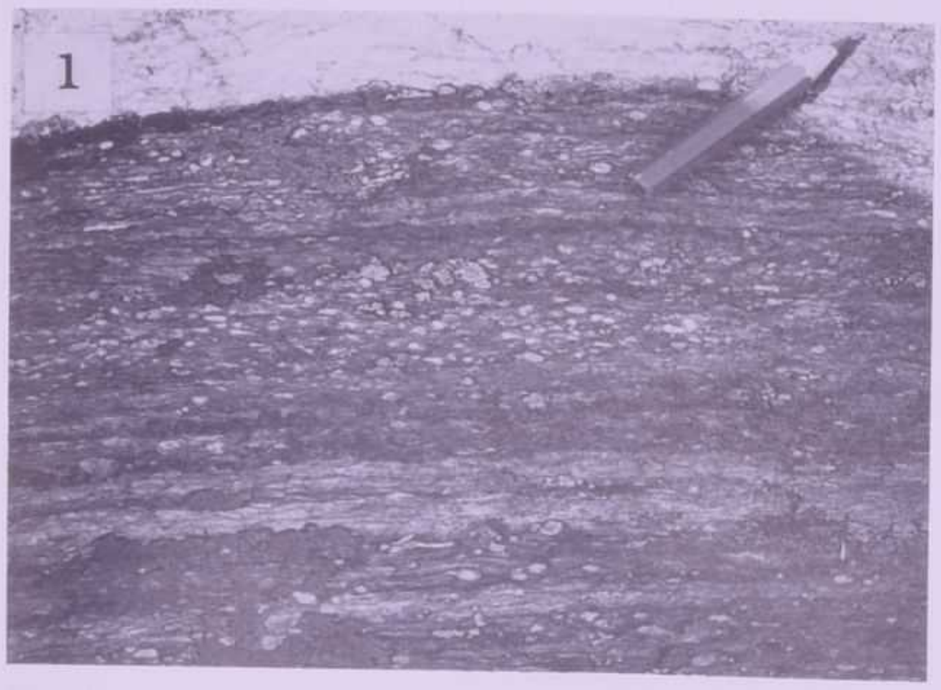
FIG. 2 - Lower greenschist proto-mylonite. Note large feldspathic augens. Sample Gi 314; NW of Acquaseria (coordinates 740.54/102.09).

FIG. 3 - Micrograph of lower greenschist mylonite. Note ribbons of stretched and dynamically recrystallizing quartz grains. Feldspars (indicated by the arrow) behave brittely. Sample Gi 279; W of M. Grona. Section is parallel to the XZ plane which dips 45° to the E (right), X nichols, scale bar = 1 mm.

FIG. 4 - Micrograph of quartz vein in lower greenschist mylonite. Dynamic recrystallization occurs by polygonization. Camaggiore profile; sample VG 93; thin section cut parallel to the XZ plane, which dips 20° to the E (left), X nichols, scale bar = 0.5 mm.

FIG. 5 - Micrograph of quartz vein in lower greenschist mylonite. Dynamic recrystallization by polygonization is almost complete. Camaggiore profile; sample Gi 326; thin section cut parallel to the XZ plane which dips 9° to the E (left), X nichols, scale bar = 0.25 mm.

FIG. 6 - Detailed picture of the same sample and thin section as Fig. 5. Scale bar = 0.1 mm.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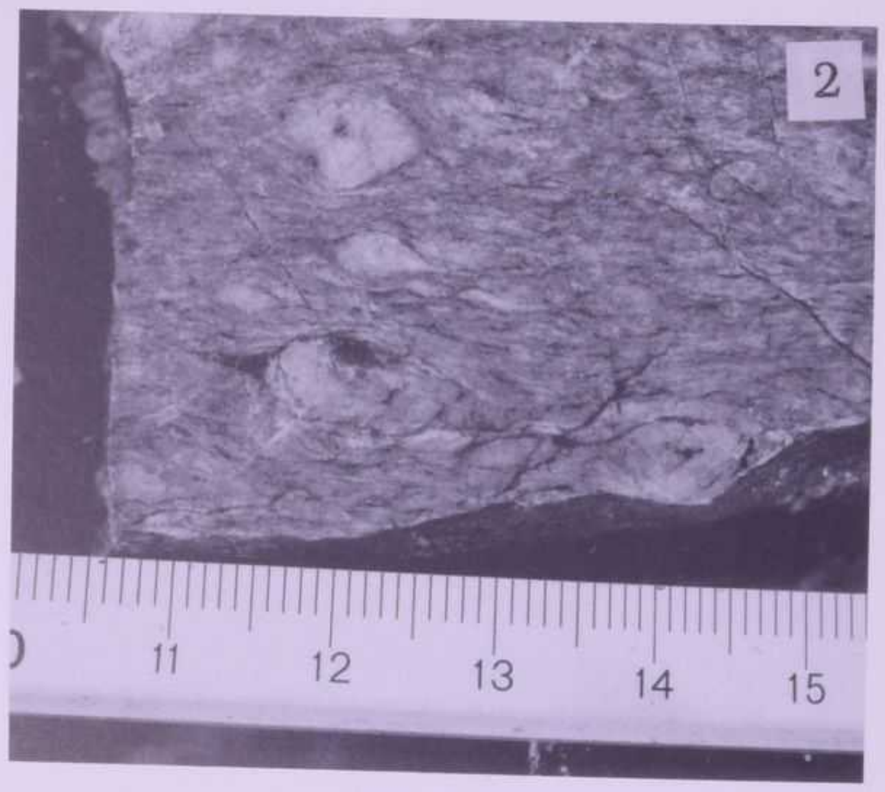
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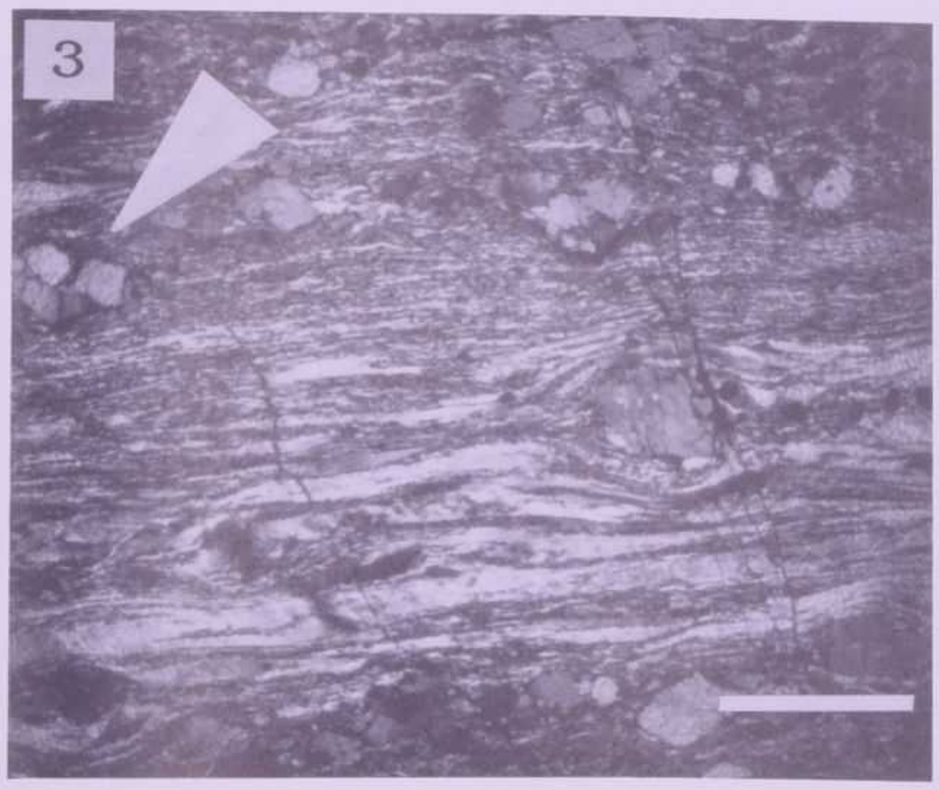
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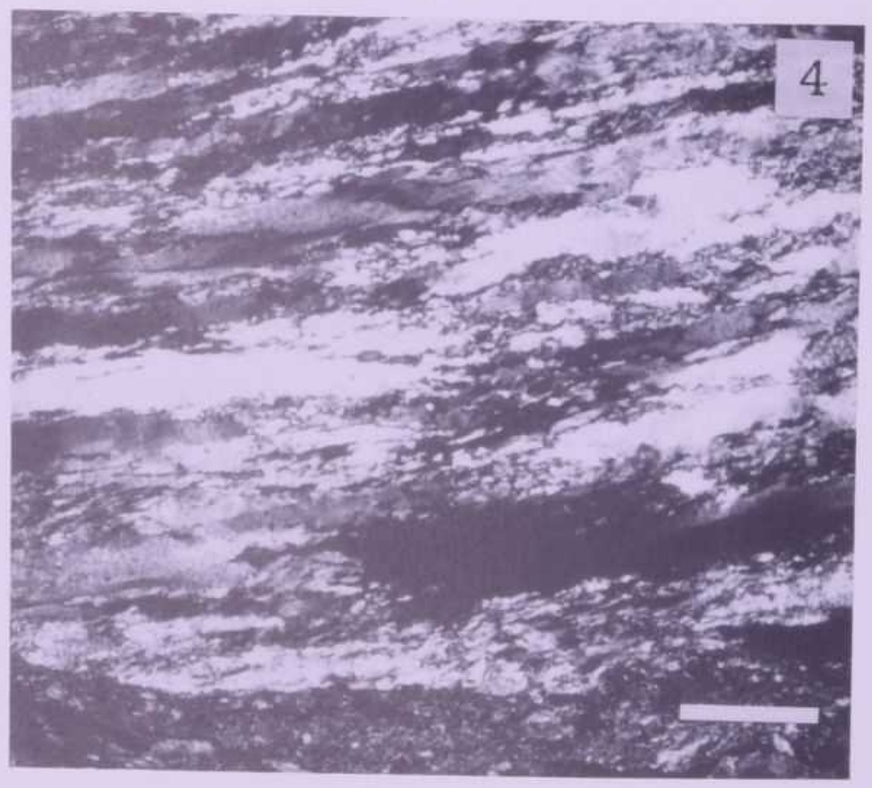
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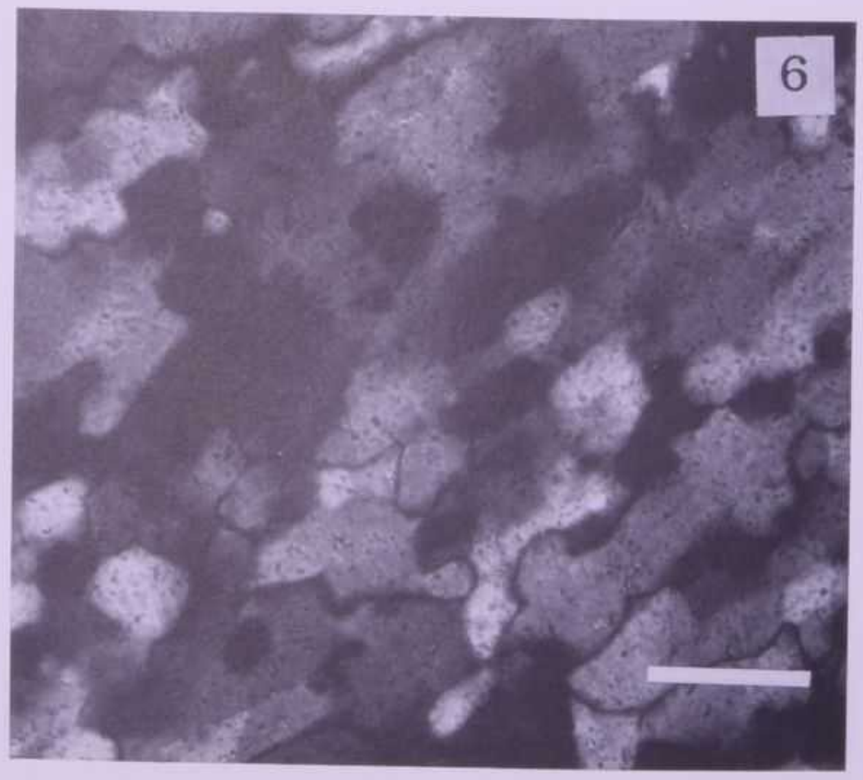
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EXPLANATION OF PLATE VI

Lower greenschist mylonites

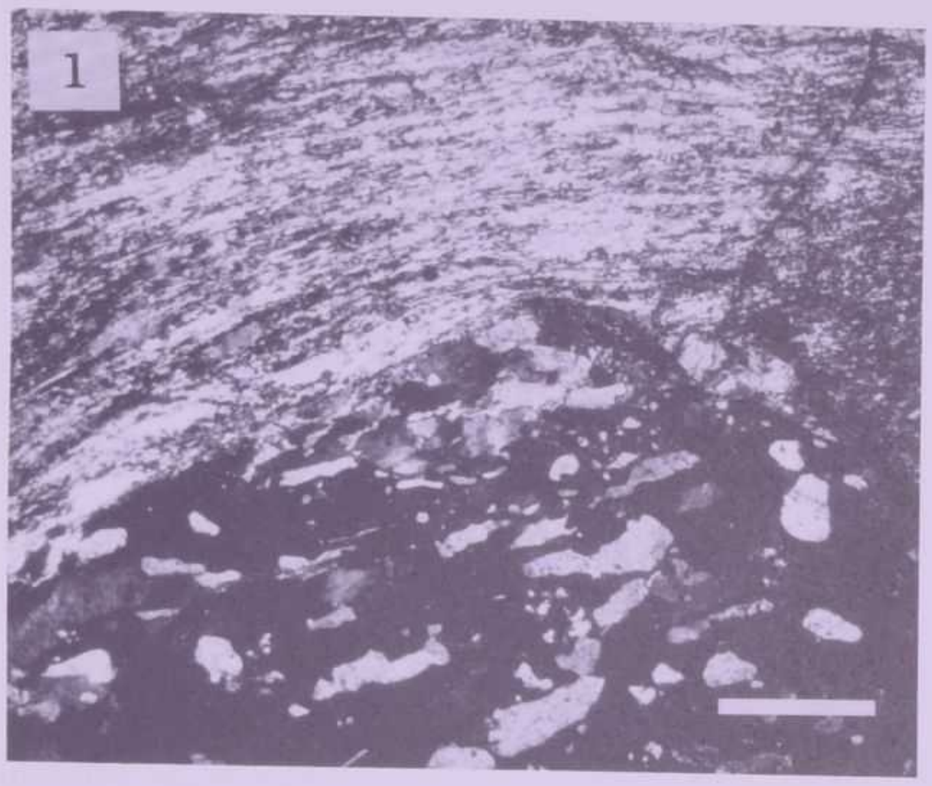
FIG. 1 - Feldspar clast in lower greenschist mylonite with exsolution structures typical for the clasts of the fault zone. Camaggiore profile; Sample VG 93; thin section cut parallel to the XZ plane which dips 20° to the E (left), X nichols, scale bar = 0.5 mm.

FIG. 2 - Myrmekitic structures in feldspars. The myrmekites could be stress-induced. NW of Acquaseria; Sample Gi 314; thin section cut parallel to the XZ plane which dips 34° to the ENE (right). X nichols, scale bar = 0.25 mm.

FIG. 3 - Microcline twinned K-feldspar typical for lower greenschist mylonites. NW of Acquaseria; Sample Gi 314; thin section cut parallel to the XZ plane which dips 34° to the ENE (right), X nichols, scale bar = 1 mm.

FIG. 4 - Sheared aggregate derived from intermediate plagioclase breaking down into sericite, chlorite and epidote. Note the length of the shear zone. Camaggiore profile; Sample VG 100; thin section cut parallel to the XZ plane which dips 44° to the E (right), // nichols, mm-grid in the lower part of the picture.

FIG. 5 - Blow-up of Fig. 4. Scale bar = 0.5 mm.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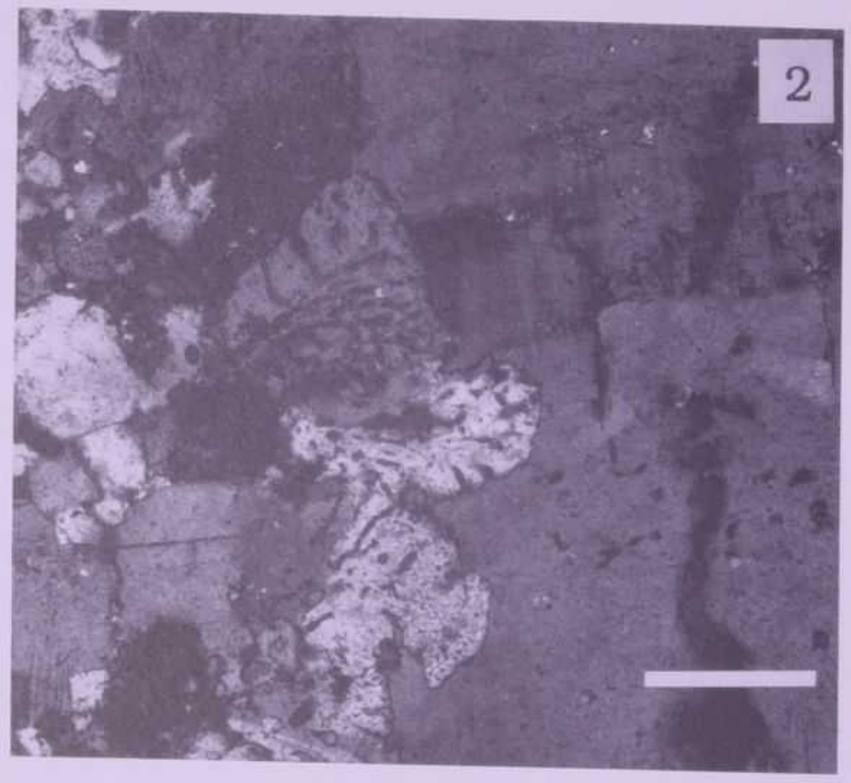
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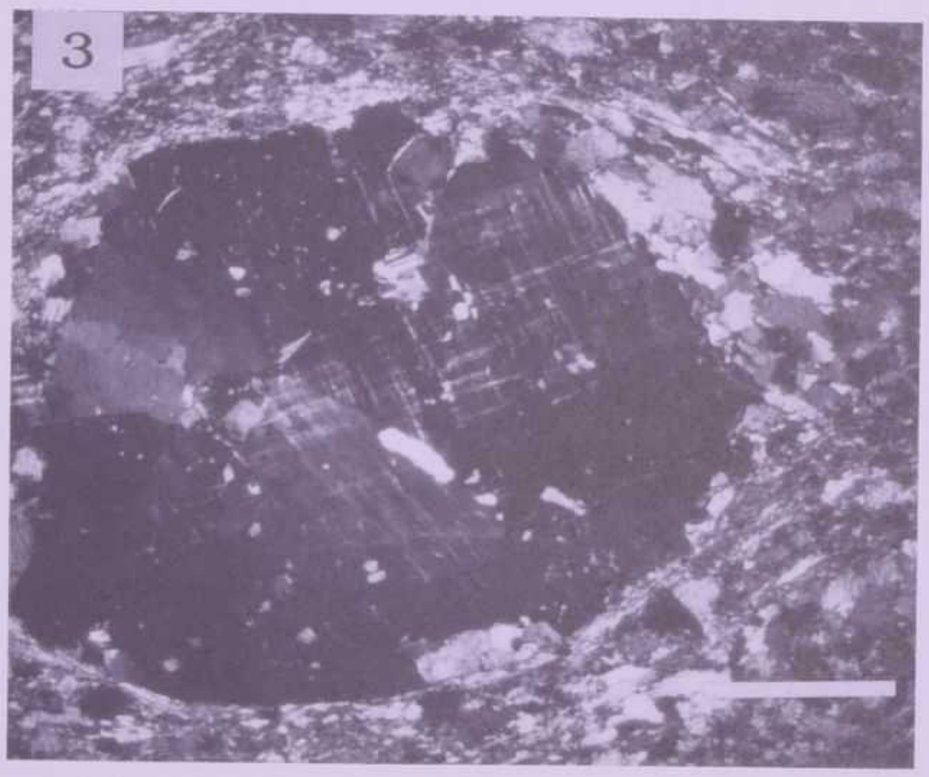
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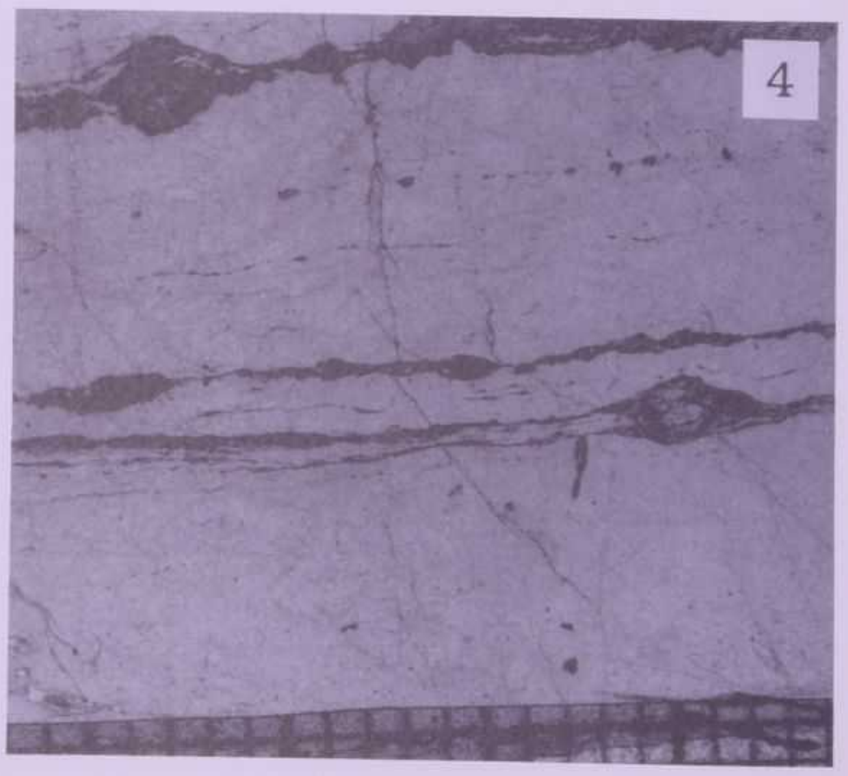
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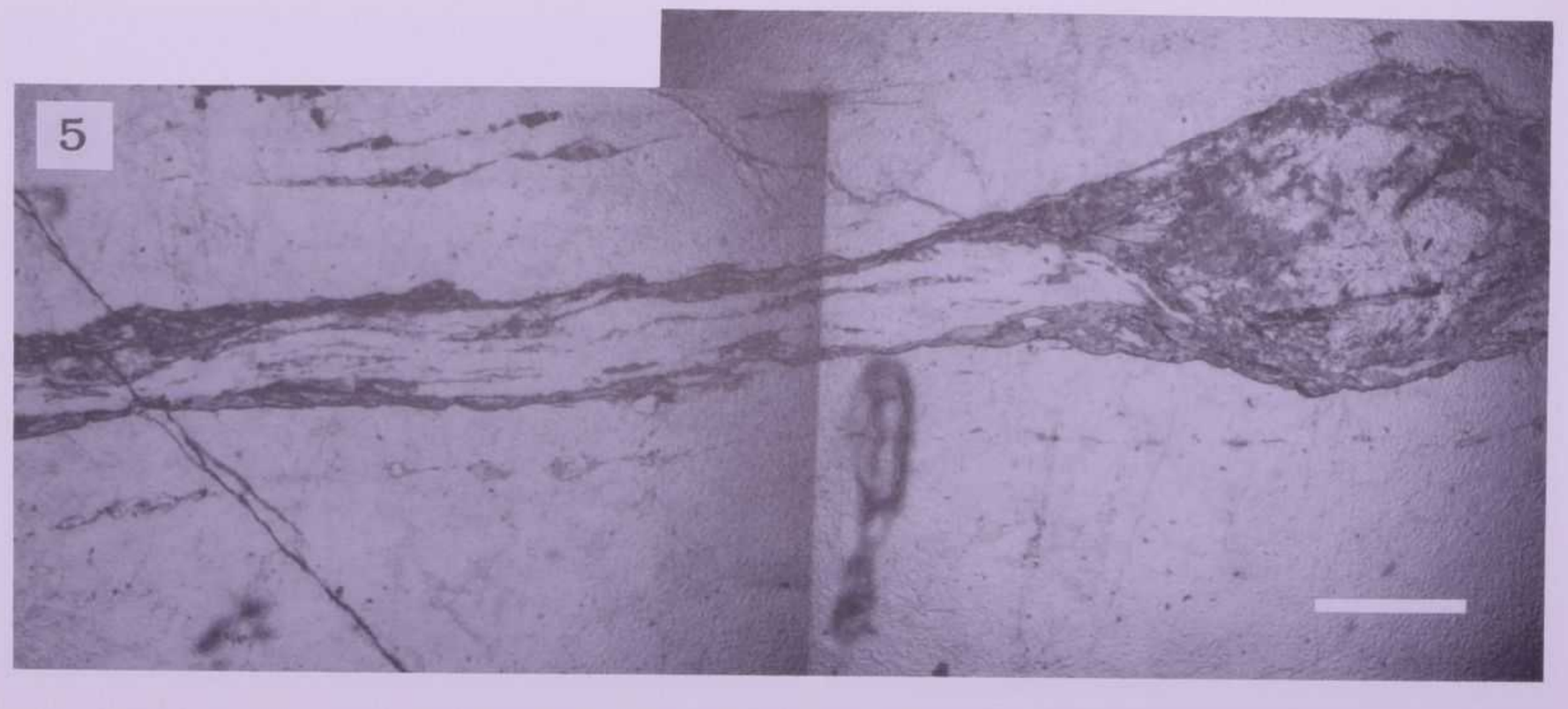
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EXPLANATION OF PLATE VII

Ultramylonites

FIG. 1 - Finely laminated dark schists made up of ultramylonites. M. Grona profile, S. Amate chapel.

FIG. 2 - Mylonitic quartz ribbons alternate with fine-grained ultramylonitic horizons. Sinistral shear bands marked by similar ultramylonitic horizons disrupt and fold the mylonitic quartz layers. Camaggiore profile; Sample VG 12; thin section cut parallel to the XZ plane which dips 24° to the ENE (right). X nichols, scale bar = 0.5 mm.

FIG. 3 - SEM back-scatter images of the groundmass of an ultramylonite: elongated crystals (probably sericite, chlorite and epidote) in a matrix of subrounded felsic minerals (albite and quartz).

FIG. 4 - Typical aspect of a well developed ultramylonite in thin section: mylonitic quartz bands are almost absent, only small-sized clasts of muscovite and feldspars are left. Note, as in the lower greenschist mylonites, dextral shear bands (sinistral in the field). Camaggiore profile, Sample VG 92; thin section cut parallel to the XZ plane which dips 32° to the ENE (left). X nichols, scale bar = 0.5 mm.

FIG. 5 - Micrograph of the fine-grained groundmass of the ultramylonites. Needle-shaped minerals are epidote; chlorite is also present. N of Acquaseria, Sample Gi 307. // nichols, scale bar = 0.05 mm.

FIG. 6 - Relic plagioclase almost completely replaced by a sericite and chlorite. These aggregates are the source for the ultramylonitic matrix. Camaggiore profile, Sample VG 92. Thin section cut parallel to the XZ plane which dips 32° to the ENE (left). X nichols, scale bar = 0.25 mm.


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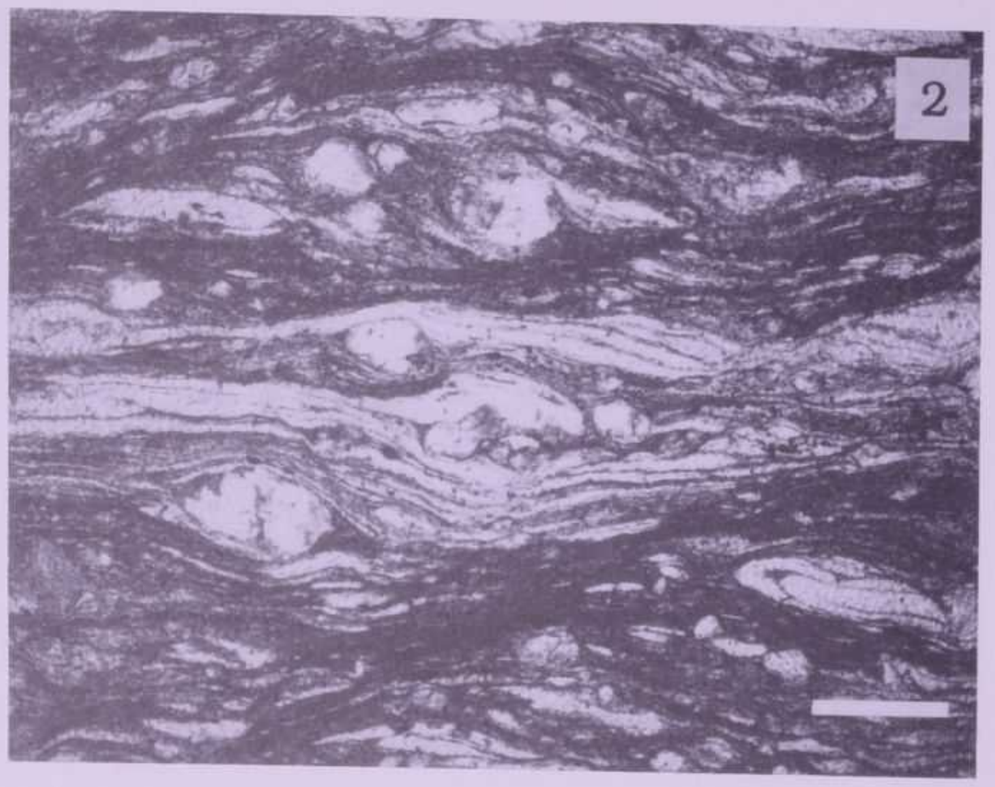
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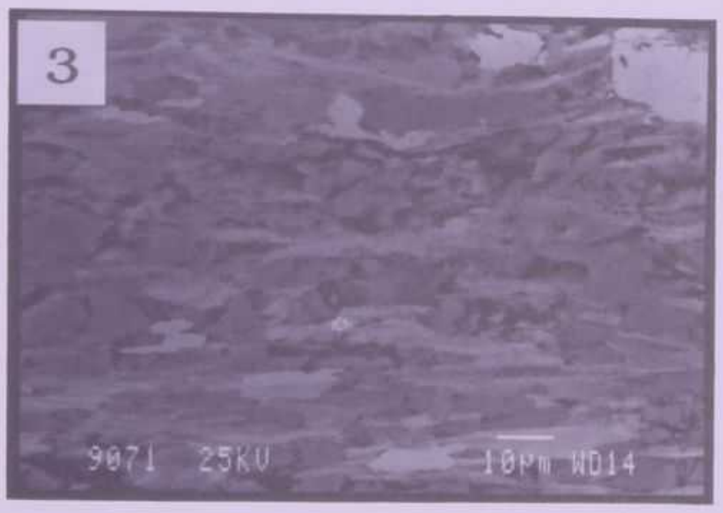
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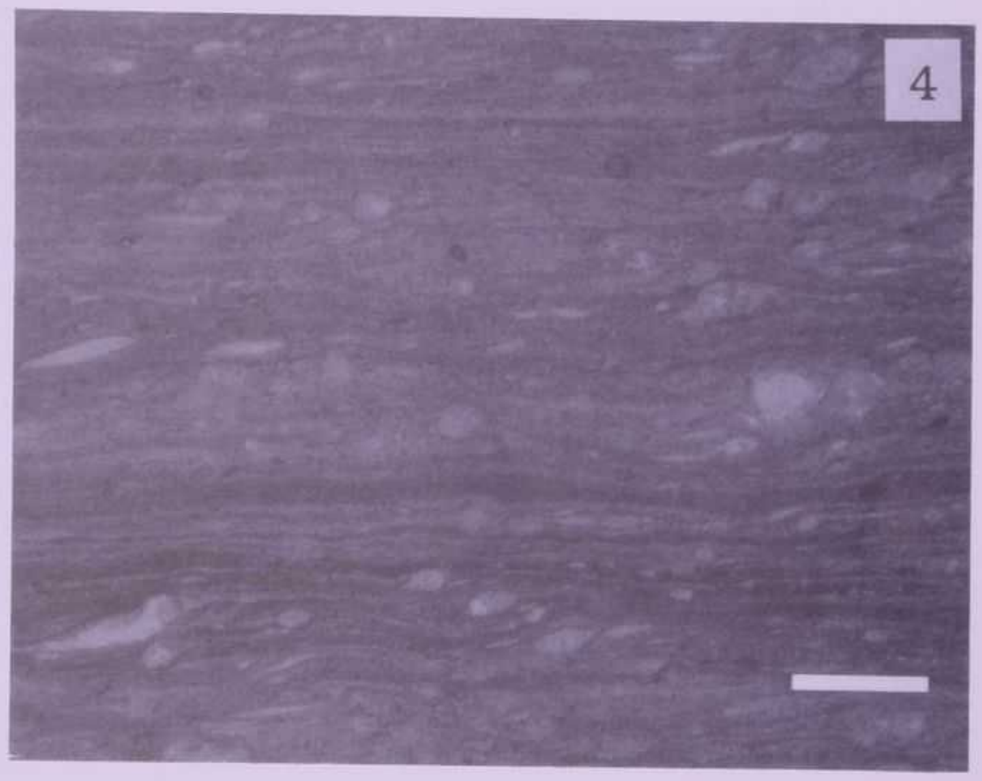
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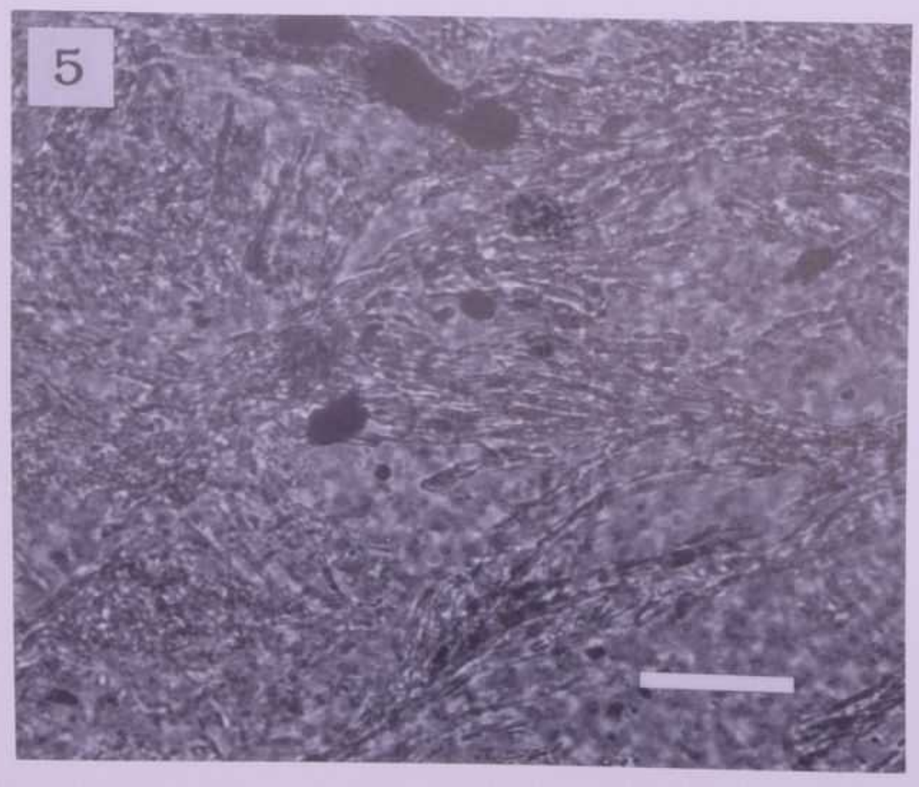
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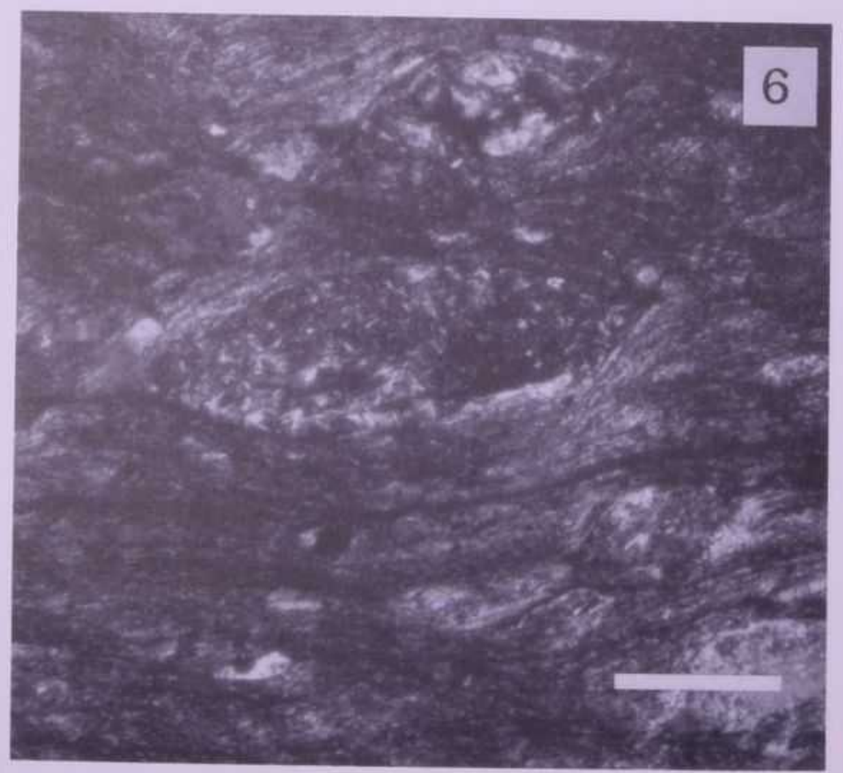
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EXPLANATION OF PLATE VIII

FIG. 1 - Cataclastic deformation disrupts mylonitic quartz layers (light-coloured). Camaggiore profile, Sample VG 98. // nichols, scale bar = 1 mm.

FIG. 2 - Pseudotachylite intruding a mylonitic quartz vein (white arrow). Camaggiore profile, Sample VG 98. X nichols, scale bar = 1 mm.

FIG. 3 - Well foliated gneiss consisting of Variscan upper greenschist mylonite. Similar mylonites are often found preserved within the Monte Grona/Val Grande fault zone. M. Grona profile, sample Gi 270.

FIG. 4 - Typical aspect of a Variscan upper greenschist mylonite under the microscope. Note ribbons of sub-rectangular quartz grains with short sides perpendicular to the foliation. M. Grona profile, Sample Gi 270. Thin section parallel to the XZ plane which dips 40° to the W (left). X nichols, scale bar = 0.25 mm.

FIG. 5 - Quartz overgrows the foliation in Variscan upper greenschist mylonites. Camaggiore, sample Gi 341. Thin section cut parallel to the XZ plane which dips 40° to the W (left). X nichols, scale bar = 0.5 mm.

FIG. 6 - Leucocratic gneiss cut by underformed veins filled with blue chlorite (dark) and calcite. Val Sanagra, Sample Gi 252.


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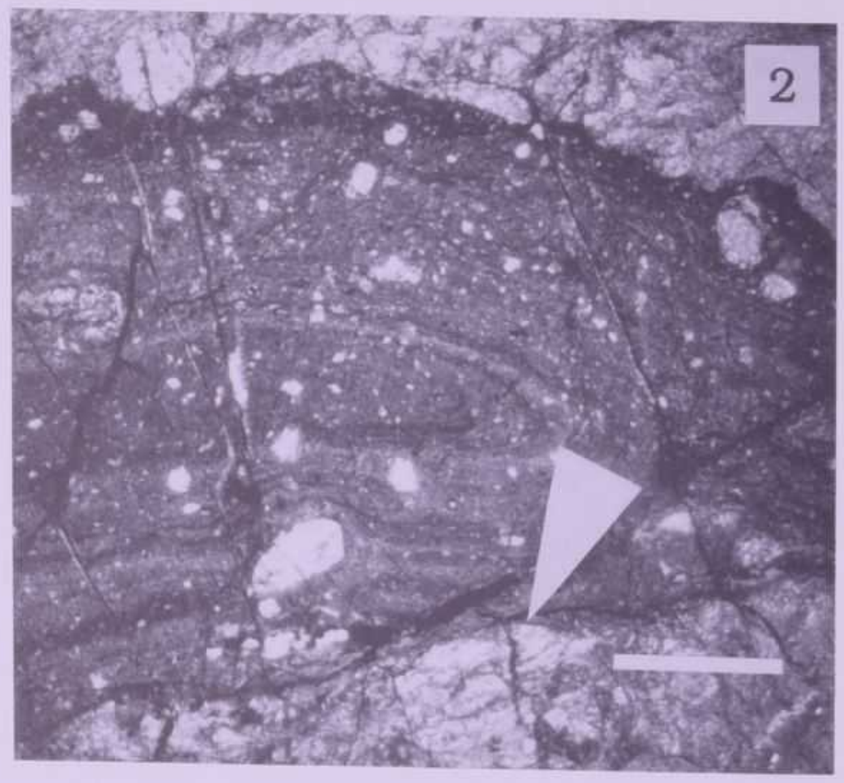
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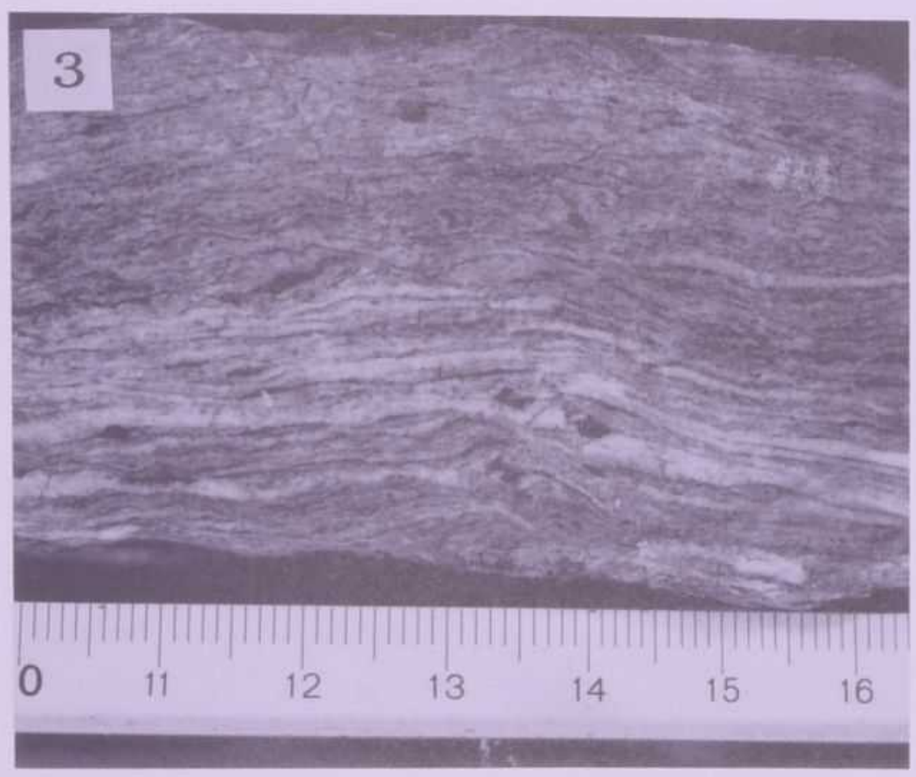
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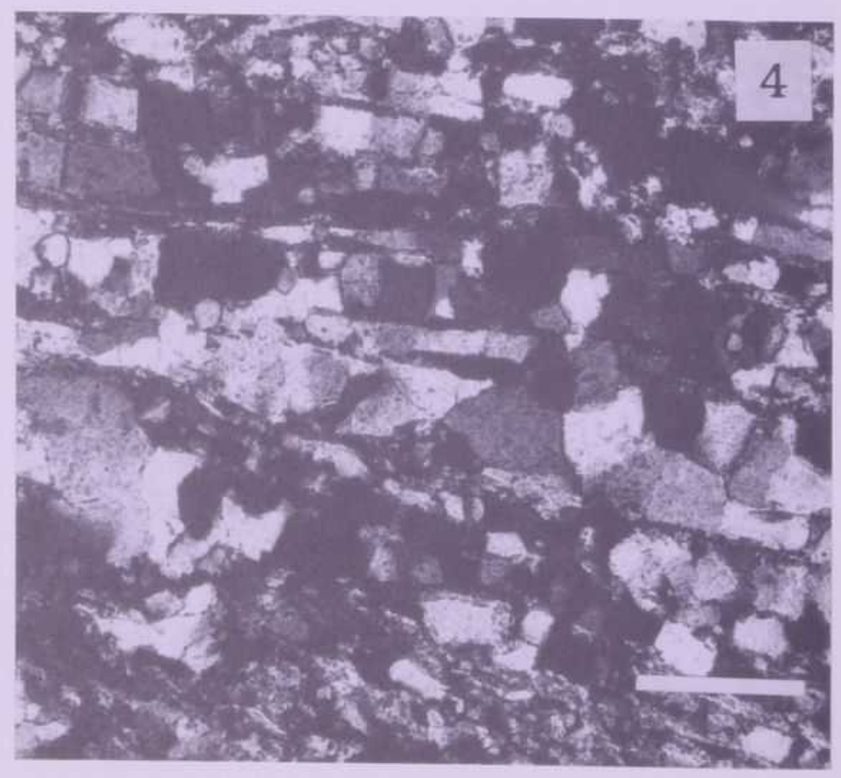
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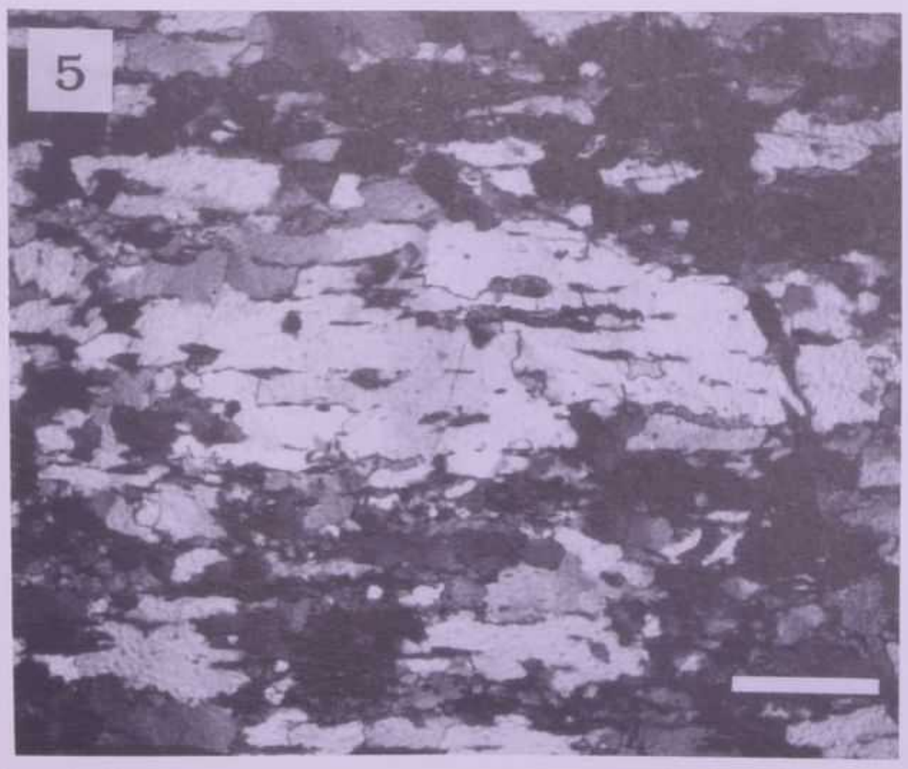
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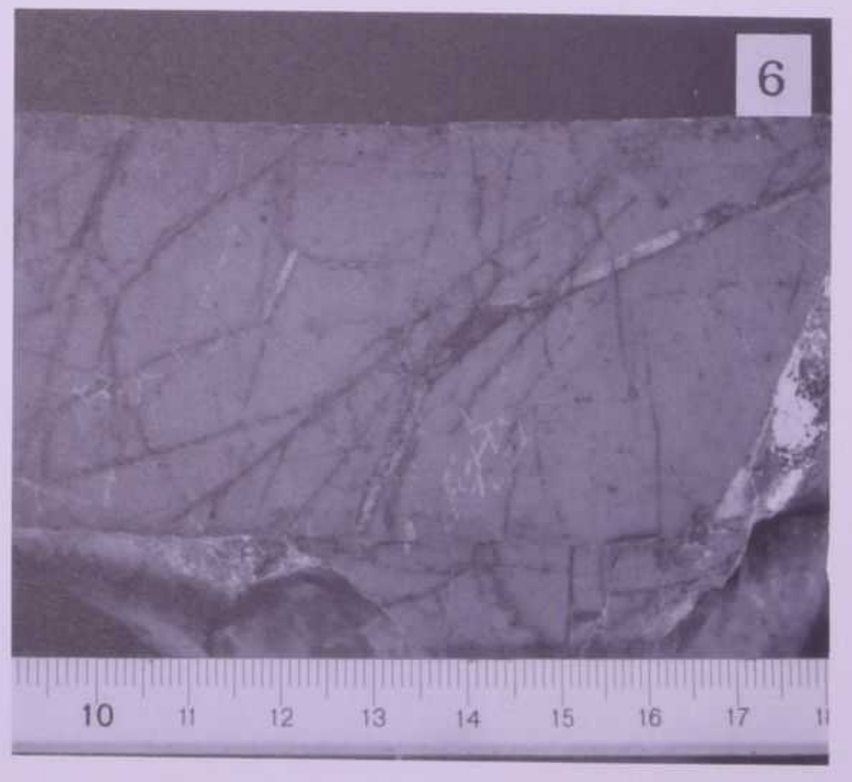
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EXPLANATION OF PLATE IX

Structures within the Monte Grona/Val Grande fault zone

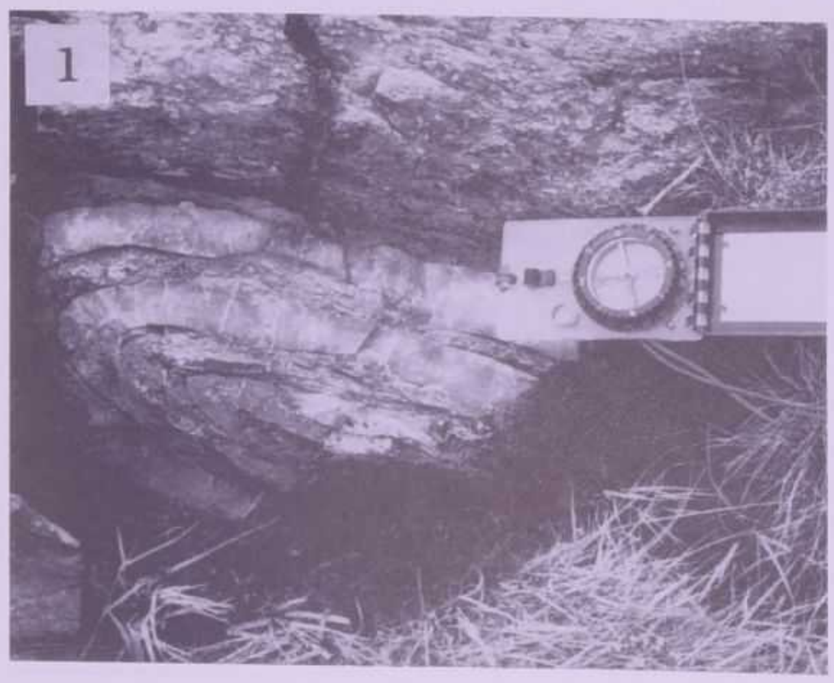
FIG. 1 - Quartz vein folded by syn-mylonitic fold. Camaggiore.

FIG. 2 - Micrograph of a mylonitic quartz vein folded by a syn-mylonitic fold. Grains on the two limbs are sub-parallel; no rotation of quartz crystals around the hinge is observed (see text for discussion). Camaggiore profile, sample Gi 335. Thin section cut perpendicular to the fold axis. X nichols, mm-grid shown.

FIG. 3 - Ductile post-mylonitic fold. Camaggiore.

FIG. 4 - Semi-brittle shear zone. The mylonitic foliation is folded into the shear zone which itself is marked by a thin horizon of ultracataclasites. Camaggiore profile.

FIG. 5 - Meter-thick, cataclastic shear zone at the base of the Gneiss Chiaro W of Acquaseria. E is to the left. A top-to-the-E sense of movement is deduced by internal thrusts and foliation (mainly due to pressure solution). Outcrop is some 5 meters high.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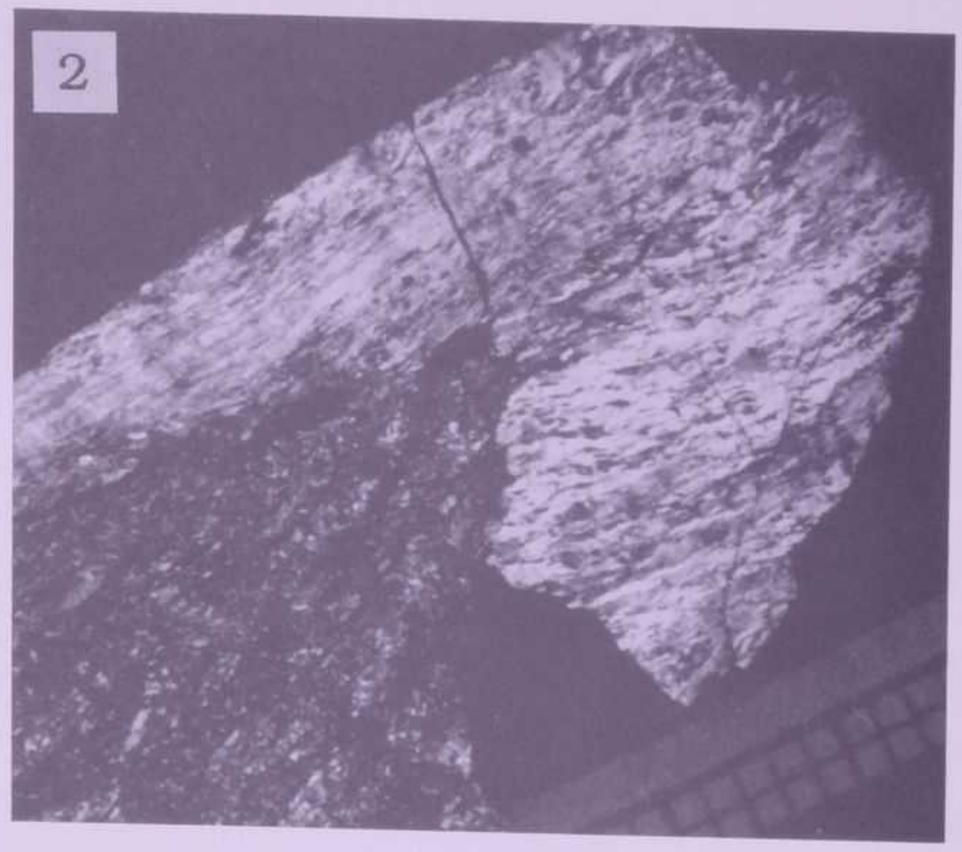
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EXPLANATION OF PLATE X

Fig. 1 - Macroscopic aspect of the Sasso di Cusino breccia (see text). Sample Gi 192.

Fig. 2 - Thin section photograph of the Sasso di Cusino breccia. Clasts are themselves fragments of breccias (A). Sparite-filled veins preceding the formation of the breccia are also observed (C). Gi 192. // nichols, mm-grid shown.

Fig. 3 - Cataclastic breccia at the base of Dolomia Principale. Upper Val Rezzo. Sample Gi 193.

Figs. 4 and 5 - Clasts in the Monte Legnone Gneiss strongly resemble those found in the lower greenschist mylonites of the Monte Grona/Val Grande fault zone. Both sections are taken from El Tahlawi's collection.

Fig. 4 - Microcline-twinned K-feldspar. Scale bar = 0.5 mm.

Fig. 5 - Feldspar with pronounced exsolution structures. Scale bar = 0.5 mm.


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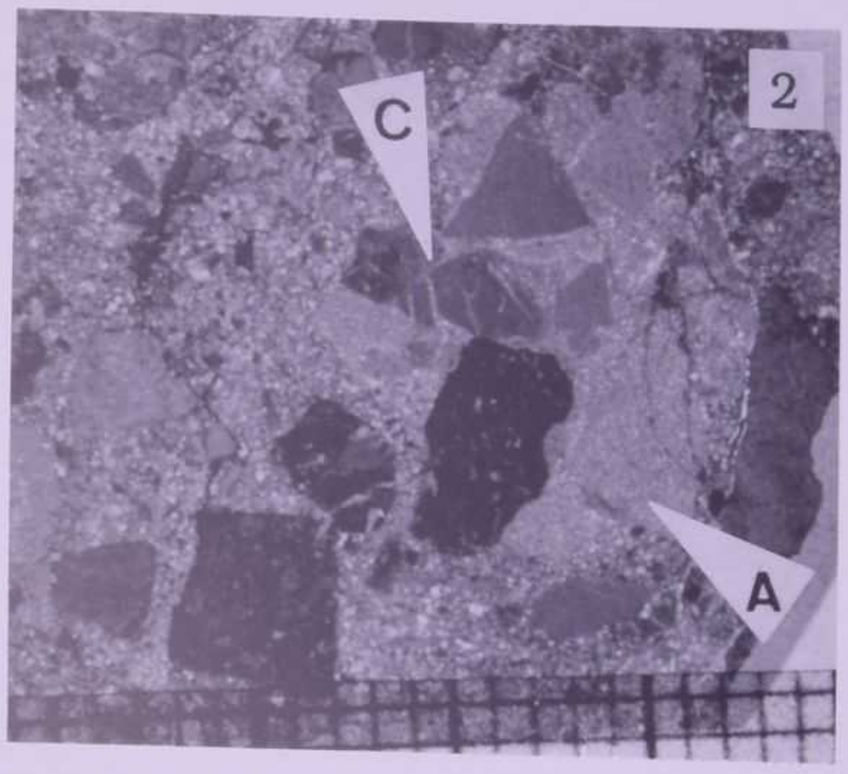
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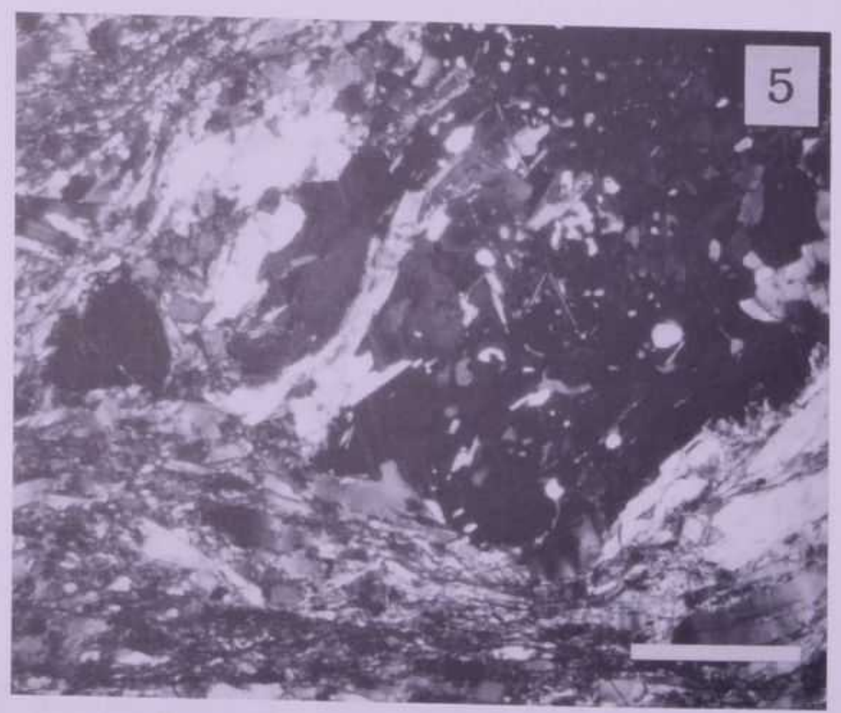
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