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FOURTH INTERNATIONAL CONFERENCE ON GEOMORPHOLOGY - Italy 1997

**Guide for the excursion**

# GEOMORPHOLOGY OF THE CENTRAL AND SOUTHERN ALPS

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## 1. INTRODUCTION

The geomorphology of the central sector of the Italian Alps has many interesting aspects. Some of these are: the relationship between the morphology of the territory and the geological structure; the morphology modelled by glaciers; the morphological evidences of neo-tectonics and the great landslides, which are typical of an alpine environment.

The itinerary of this field trip will allow us to observe several of these aspects, travelling by coach, ferry, chair-lift and walking. During the first day we will be travelling from Milano to Como and will cross the end-moraine system of Lake Como; then we will be travelling by ferry on Lake Como from Como to Colico. After this we will follow the Valtellina, one of the main valleys of the Italian Alps. During the second day, travelling by jeep and walking, we will visit some of the glaciers of the Ortles-Cevedale Group and their Late Glacial and Holocenic moraine systems. During the third and fourth days, moving from Valtellina to Valcamonica, we will examine the glacial geomorphology of the tonalitic massif of the Adamello. The fifth day will be devoted to the great rock-avalanches of the Sarca and Adige valleys in the Trentino region. On the sixth day, travelling by boat, we will observe the raised beaches and the lake levels of Lake Garda. Finally, during the seventh day we will cross the end-moraine system of Lake Garda and the Po plain and, passing through Mantova, we will arrive in Bologna for the Conference (fig. 1).

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FIG. 1 - Pre Congress Excursion A1: The itinerary.

## 2. GENERAL STRUCTURAL SETTING OF THE CENTRAL ALPS
(A. Zanchi)

According to the generally accepted plate tectonics models of the Mediterranean region, the history of the Alpine belt begins during Late Carboniferous after the end of the Hercinian orogeny. The evolution of the Alps is strictly related to the opening of the Neotethys ocean, splitting the Pangea supercontinent into the Laurasian and Gondwanan plates. The sedimentary successions deposited in the Alpine region record several different episods of rifting from Permian to Jurassic with predominant deposition of marine carbonatic successions. During Jurassic two main sinistral E-W trending fracture zones splitted the Europa and Africa plates forming in between small microplates (Iberia, Adria-Apulia, Pelagonids-Menderes) separated by small oceans (i.e. Ligure-Piemontese ocean).

At the end of the Jurassic the paleogeographic domains of the area now forming the Alps were, from north to south, as follows:

– The Helvetic-Delphines Domain: shallow sea successions deposited on the continental crust of the southern European margin.

– The Pennidic Domain: represented by oceanic crust (now represented by the ophiolitic units of the Zermatt-Saas Zone in the Western Alps, and Malenco-Forno Nappe in the Central Alps) and sedimentary successions deposited on the extended continental crust bordering the oceanic domain.

– The Austroalpine Domain: represented by marine successions deposited on the continental crust of the northern margin of the Adria microplate (a spur of the African plate).

– The Southern Alps: represented by marine succession deposited in the Adria microplate south of the Austroalpine domain.

At the end of the Jurassic, the opening of the Atlantic Ocean caused the inversion of the relative motion between Africa and Europa, leading to the closure of the interposed oceans (Ligure-Piemontese ocean) and to the beginning of the Alpine orogeny. The formation of an accretionary prism between Europa and Adria, probably due to south-vergent subduction, is well documented by the blue-schist and eclogitic metamorphism (Eoalpine phase) developed in the Pennidic and Austroalpine units of the Western Alps from Lower Cretaceous to Paleocene. Thrust-embrication was also active in the external domains of the Southern Alps and was accompanied by foredeep development with Late Cretaceous flysch deposition in the Lombardian sector.

During Eocene, the Central Alps were affected by intensive regional metamorphism (Mesoalpine phase) related to the thermal Lepontine dome, caused by intensive crustal thickening following continental collision. The Helvetian and Pennidic units affected by this metamorphic event are sharply truncated southward by the Insubric Line, an important E-W trending fault system now separating the metamorphic units of the Alps (Pennidic and Austroalpine) from the Southern Alps, where south-vergent thrust-stacking was accompanied by very low-grade metamorphism only in the deepest portions of the thrust-fold belt. South-vergent back-thrusting of the Central Alps with formation of antiformal structures (Cressim antiform in the Adula Nappe) and subsequent dextral strike-slip motion at shallow crustal levels during Miocene (Neoalpine phase) are generally recognized all along the Insubric Line.

During Oligocene, large plutons intruded north (Masino-Bregaglia) and south (Adamello-Presanella) of the Insubric Line.

At the end of the Oligocene, in the southern part of the Lombardian sector, a foredeep basin, roughly corresponding with the area now occupied by the Northern Po Plain, was infilled by coarse turbiditic deposits of the Gonfolite Lombarda Group, testifying high rate of uplifting of the Central Alps.

Thrust embrication in the Southern Alps and dextral motion along the Insubric Line were active up to the end of Tortonian. During this stage also the Oligo-Miocene foredeep successions were stacked southward leading to the formation of the Milano thrust belt, now buried below the Plio-Quaternary sediments of the Po Plain (the so-called pedealpine monocline).

During Messinian, due to the drying up of the Mediterranean basin, the Alpine region was affected by deep erosion leading to the excavation of the major alpine valleys. At the beginning of Pliocene, with the opening of Gibraltar and filling of the Mediterranean, the sea transgressed the lower part of the Southern Alps foothills, filling the Messinian canyons. The incoming of glaciations, probably starting from Late Pliocene, and the continuous uplift of the region caused a definitive regression of the sea from the Lombardian plane, which was successively covered by continental fluvial and fluvioglacial deposits. At present, low seismicity, continuous uplifting, and isolated neotectonic evidences suggest a possible prosecution of tectonic activity at low strain rates (fig. 2).

### 2.1 Southern Alps

Most of the field-trip crosses the units of the Southern Alps, including its crystalline basement and its sedimentary cover. The Southern Alps are bounded to the north by the Insubric Line which follows the lower Valtellina between the beginning of Val Chiavenna and Sondrio. East of Sondrio the fault zone continues eastward passing just north of the Aprica and Tonale Passes and bounding the northern part of the Presanella intrusive body along the Val di Sole up to Malé where it joins the Giudicarie Line.

The sedimentary cover of the Southern Alps includes thick carbonatic and terrigenous successions spanning in time from Late Carboniferous to Tortonian (Late Miocene). The Mesozoic paleogeography is dominated by the presence of roughly N-S trending subsident basins filled by carbonatic and terrigenous successions and structural heights with development of carbonatic platforms. In particular, in the Como area, the typical Triassic and Jurassic formations of the Lombardian Basin are exposed, whereas from Malé to Brescia outcrop the successions of the so-called Garda escarpment which make transition from the dee-

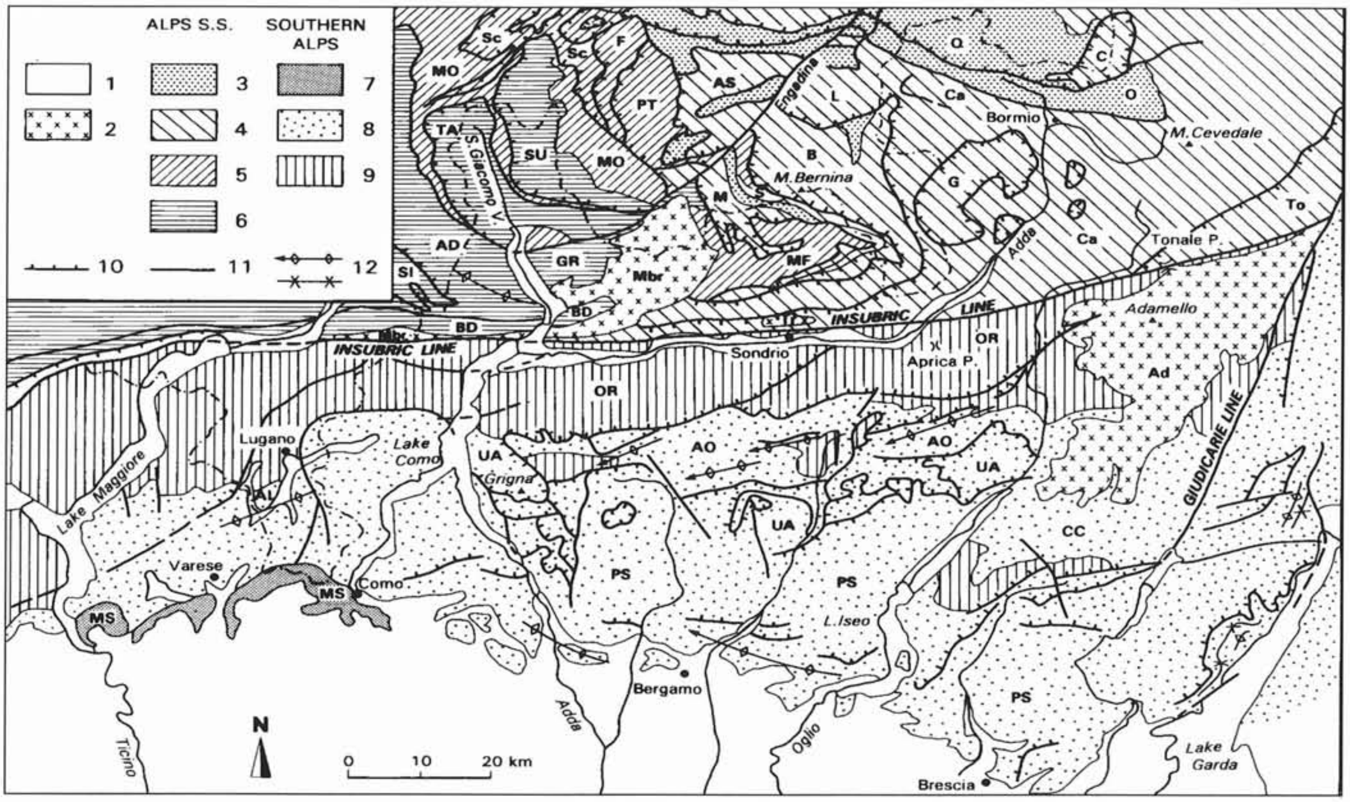



FIG. 2 - Simplified structural maps of the Alps; redrawn from Forcella & Moltrasio in Guide Geologiche Regionali. Alpi e Prealpi Lombarde, Soc. Geol. It., 1, 1990, Bema , Milano). 1) Late and post-orogenic units. 2) Cenozoic intrusives (40-25 MA); Mbr: Masino-Bregaglia, Ad: Adamello, Tr: Triangia. Alpine units north of the Insubric Line: 3) Austroalpine triassic sedimentary covers; Q: Quattervals Nappe, C: Chavalatsch Nappe, O: Ortles Nappe; 4) Crystalline basement of the Austroalpine units; B: Bernina Nappe, S: Sella Nappe, M: Margna Nappe, G: Grosina Nappe, L: Languard Nappe, Ca: Campo Nappe, TO: Tonale Nappe; 5) Mesozoic sedimentary covers of the Pennidic Units often associated with Ophiolitic units: MO: Mesozoic covers and/or associated with ophiolites, MF: Malenco-Forno Ophiolitic Complex, SC: Schams Nappe, PT Platta Nappe; 6) Alpine and Pre-Alpine Pennidic metamorphic basement: AD: Adula Nappe, TA: Tambò Nappe, GR: Gruf migmatitic gneiss, SU: Suretta Nappe, BD Bellinzona-Dascio Zone. Southern Alps: 7) Oligo-Miocene foredeep sediments: MS: Gonfolite; 8) Undifferentiated Permo-Mesozoic and Paleocene sedimentary cover: AL: Luganese Anticline, AO: Orobic Anticlines, CC Camune Anticline (Culminazione Camuna), UA: Alloctonous carbonatic units, PS: Parautoctonous units and pedealpine flexure; 9) Crystalline basement: OR: Orobic Thrust, CC: Camune Anticline; 10) Main overthrusts; 11) faults; 12 fold axes.

per Lombardian Basin to the less subsident Trento Plateau. The Garda escarpment, developed during Mesozoic, was successively reactivated at several times during the alpine compression and still separates structural sectors with different cinematic behaviour.

The crystalline basement consists of low grade metapelites and metapsammites which crop out in the northern part of the Orobic Alps; few fossil findings suggest an Early Paleozoic age of the protholits. Metagranitoids, amphibolites and quartzites also occur in many localities. A complex polyphasic metamorphic evolution with successive retrograde conditions has been recognized for these complexes, but no important metamorphism occurred during the Alpine orogeny.

The structure of the Southern Alps in the Lombardian sector is dominated by the presence of polyphasic southvergent thrust systems, whose age of motion and composition is generally younger southward. Also the crystalline basement and the Oligo-Miocene molassic successions are involved in thrust motions. In the easternmost sector of the Orobic Alps, fold axes and thrust structures rotate anticlockwise assuming a NNE-SSW trend in association with the Giudicarie Line, an important fault system superposed on the boundary between the Lombardian Basin and the Trento Plateau. East of the Giudicarie Line, ESE-vergent thrusts and folds (Tremosine-Tignale thrust and Tremosine-Tignale Gargnano folds) of the easternmost structural zone of the Lombardian Basin make transition to NNE-SSW high-angle reverse to sinistral strike-slip faults (Ballino-Garda, Arco and Sarca Lines) and faulted ramp folds with the same trend (M. Baldo and M. Altissimo anticlines) of the Venetian Platform structural zone. Important neotectonic activity has been recognized along some of these structures.

## 2.2 AUSTROALPINE DOMAIN

The Austroalpine units form the uppermost part of the Alpine belt, covering the Pennidic complexes. The tectonic boundary between the Austroalpine and Pennidic nappes, which is almost horizontal along the watershed of the Alps, is steeply folded southward toward the Insubric Line giving origin to the so-called Austroalpine root zone. The Austroalpine of the Central Alps is formed by several polymetamorphic crystalline nappes with a pre-alpine imprint, which did not undergo the HP-LT alpine metamorphism present in the Pennidic Nappes. In the northernmost part of upper Valtellina (north of Bormio) the Austroalpine includes a complex system of thick Mesozoic carbonatic successions. The Austroalpine is generally divided in three main nappe complexes.

The lower Austroalpine units consist of the Margna (pre-Permian basement and Permo-Mesozoic cover), Sella (late-Hercinian granites of 270-300 Ma) and Bernina (late-hercyinian intrusives with paragneiss and a discontinuous sedimentary cover) Nappes, cropping out along the Swiss border.

The middle Austroalpine units are represented by the Campo-Ortler Nappe, which comprises all the units located between the lower and the upper Austroalpine. This nappe includes marbles- and prasinite-bearing phyllites (Filladi di Bormio) and a lower paragneissic unit with augen-gneisses and amphibolites. Late-Hercinian granitic to granodioritic bodies are intruded into this unit as well as the stratified gabbric intrusions present around Sondalo (Gabbro di Sondalo). The sedimentary units of the Middle Austroalpine Nappe form the Ortles-Quattervals nappe, stacked above the Filladi di Bormio along the north-dipping Zebrù Line. At present it is not clear whether it is an independent nappe or it represents the sedimentary cover of the Campo Unit. The Ortles-Quattervals Nappe is formed by several folded thrust sheets including a thick Permo-Mesozoic succession which forms the «dolomitic chain» present north of Bormio.

The highest structural element is the Tonale-Grosina Nappe (Upper Austroalpine) forming the highest part of the Grosina Valley (Cima Piazzi). Migmatites, orthogneisses, garnet-biotite paragneiss and sillimanite paragneiss are the main lithologies of this unit, which comprises the so-called Austroalpine root zone developed along the Insubric Line from Valtellina to the Tonale Pass.

### 2.3 Pennidic domain

The Pennidic units are developed north of the Austroalpine, generally quite far from the region crossed during the field-trip. A natural section across the Pennidic of the Central Alps, from the lowermost units, consisting of continental crust, up to the ophiolite bearing complexes of the Pennidic ocean, corresponds to the Mera Valley (from lower Val Chiavenna to the Spluga Pass). All these units experienced the HP-LT Eo-alpine metamorphism and the successive Meso-alpine Lepontine event. The lowermost unit is the Adula Nappe which is strongly steepened southward (Cressim Antiform) against the Austroalpine Tonale Zone, occurring just north of the Insubric Line. The Tambo and Suretta Nappes, separated by the so-called Mesocco and Spluga Synclines, follow upward. Ophiolite-bearing calcschists of the Schams Nappe and other ophiolites are present further north and east in Switzerland.

The largest ophiolitic complex of the Pennidic domain of the Central Alps (Malenco-Forno Ophiolitic Nappe) occurs east of the Masino-Bregaglia pluton in the upper Val Malenco, forming the M. Disgrazia massif just north of the town of Sondrio.

### 2.4 Tertiary intrusives

Two main alpine post-collisional intrusive bodies outcrop in the Central Alps: the Adamello and the Masino-Bregaglia plutons.

The Adamello pluton is located between the Insubric and the Giudicarie Lines. The pluton is entirely intruded into the crystalline basement of the Southern Alps in its northern part, whereas the southern portion of the magmatic body is intruded in the Permo-Mesozoic succession of the same structural domain. Spectacular contact metamorphism developed especially in the southern sector. The pluton includes three main bodies: the Re di Castello (42-40 MA), the Adamello (36-32) and Presanella (32-29), whose composition varies from granodiorites to tonalites through quartz-diorites. These intrusions post-date early alpine folds and are affected by successive deformations.

The Masino-Bregaglia pluton entirely lays north of the Insubric Line among Val Chiavenna, lower Valtellina, and the top of M.Disgrazia; it is intruded into the Austroalpine and Pennidic units. Radiometric ages span between 36 and 24 MA, with the exception of the two-micas San Fedelino body which is 18 MA old. The intrusion of the Masino-Bregaglia complex post-dates the Mesoalpine Lepontine phase, whereas the southern part of the pluton is stretched along the Insubric Line by dextral shearing. Two main facies are present: the Serizzo diorite and the Ghiandone granodiorite with characteristic large K-feldspar crystals.

## 3. GEOMORPHOLOGICAL ASPECTS
(G. Orombelli)

The central sector of the Italian Alps has typical aspects of alpine relief, characterised by high relief energy and glacial erosion morphology. Only the southernmost strip of the Lombardian Prealps, which was only partially reached by the glaciers, shows wide areas characterised by fluvial morphology. In addition, many carbonatic reliefs are present in the Prealps, which exhibit a karst morphology. Finally, a characteristic of the central sector of the Italian Alps is the presence of the great pre-alpine lakes.

Among the most evident structural features in this alpine sector is the rectilinear valley of Valtellina, oriented E-W and conditioned by a first order geologic structure, the «Insubric Line». Along this line the two main geological-structural blocks that form the alpine construction come into contact. These are the Southern Alps to the south and the Alps to the north. Other evident structural linear elements, oriented NNE-SSW, are the Garda lake basin, the

Chiese valley and the Valcamonica. The first two are connected to the «Giudicarie Line».

The higher elevations are along the principal alpine water divide, which has numerous segments ranging between 3000 and 4000 m in altitude (Mount Bernina 4050 m a.s.l.), or in some mountain groups immediately to the south, which are elevated for lithological or structural reasons, such as Mount Disgrazia (3678 m a.s.l.), Pizzo Scalino (3323 m a.s.l.), Cima Piazzi (3439 m a.s.l.), San Matteo (3678 m a.s.l.) and in the quartz dioritic Tertiary pluton of the Adamello (3554 m a.s.l.). From the water divide of the Alpi Orobie the elevations regularly decrease towards the south and the Prealps, with the exceptions of some local peacks represented by carbonatic massifs such as Grigne (2410 m a.s.l.), Arera (2512 m a.s.l.) and Presolana (2521 m a.s.l.). The contact with the plain is rapid, without a gradual passage through hills. The present uplift rate is estimated to be about 1.5 mm/year in the axial alpine area and it decreases along the southern slopes and towards the east, where the eastern Po plain is undergoing active subsidence.

### 3.1 Glacial morphology

During the last great glacial expansion of between 20 000 and 15 000 years ago, the Alps were almost totally buried underneath a glacial cover, which in the main valleys could reach a thickness of 2 km (fig. 3). In the Italian Central Alps only the most elevated crests and peaks emerged from the ice, as well as a few isolated mountain groups and a great portion of the Prealps. These were surrounded by great valley glaciers that occupied the present day prealpine lakes. Between 15 000 and 10 000 years ago, the glacial cover progressively retreated, although with temporarily advance phases (the Late Glacial stages), until they were reduced to the present situation where only small residual glaciers are present on the higher massifs.

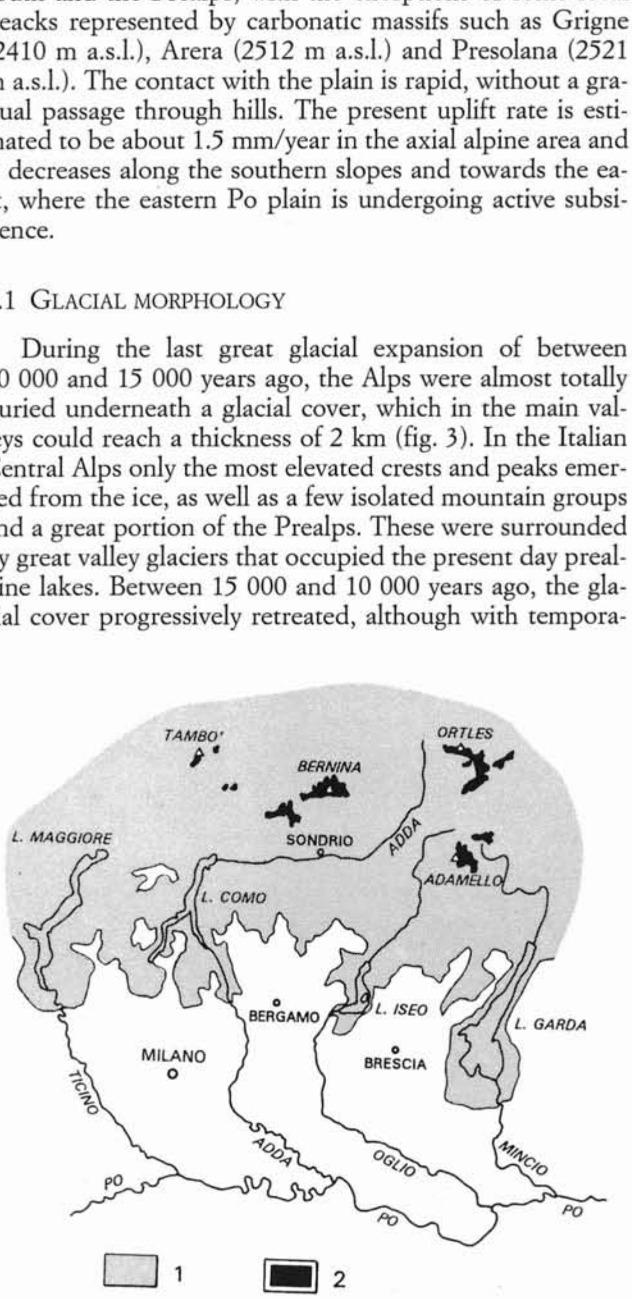



FIG. 3 - Late Glacial Maximum (1) and present (2) glacier extent in the Central Italian Alps.

In the past, the Alps experienced many other glaciations during the Quaternary and the glaciers were probably already present during the Pliocene. Therefore, it is apparent that the alpine territory has been deeply and almost totally modelled by glacial action. This differentiates their morphology from the one of the nearby Appennines, which were only slightly touched by glacial action in their most elevated areas. During the glaciations the glaciers expanded to form an ice field with a system of coalescent valley ice tongues, and reached the plain in the form of wide and flat piedmont glaciers. The Italian Central Alps have been dominated by some main glacial systems, which take their names from the lakes that formed after their retreat (fig. 3).

The glacial system of the Lake Maggiore collected the ice flows coming from a wide mountain area, ranging from the eastern side of the Monte Rosa to the high valleys of the Ticino, and outlet in the plain to form a broad piedmont glacier, about 30 km wide. A part of the glaciers of the easternmost valley of the Ticino, through the Monte Ceneri pass, flowed towards the present Lake Lugano to form the small end moraine system of Porto Ceresio. In addition, some of the glacier ice of the Adda and Como systems overflowed, into the main branch of the present Lake Lugano, through the Menaggio saddle. The Como glacier collected the glacial flows of Val Chiavenna, Val Bregaglia and Valtellina. Towards the south it split into many branches, and finally formed the piedmont glaciers of Como, Brianza (Lambro) and Lecco. The evidence for this is in the concentric hill arcs of the end moraine systems between the Adda and the Ticino rivers. Along the Valcamonica another important glacial system descended and occupied the position of the present Lake Iseo. From this position a small piedmont glacier expanded to the south, another lobe climbed the Val Borlezza up to the present depression of Clusone, and another entered the Val Cavallina up to and past the present Endine Lake. Finally, at the eastern margin of Lombardy, the glaciers of the Idro and Garda lakes descended from the Chiese and Sarca-Adige valleys respectively. The Garda glacier expanded in the plain of Brescia and Verona, forming a piedmont glacier about 30 km wide. Minor glacial systems, entirely confined inside the valleys, were present in the Brembana and Seriana valleys.

In the central sector of the Italian Alps there are, at present, over 200 glaciers, with a surface area of about 100 $km^2$. The greatest glaciers are located in the mountain groups of the Bernina-Disgrazia, Ortles-Cevedale and Adamello. By contrast, the glaciers of the northern side of the Orobie have very modest dimensions. The extent of glaciers is constant in time. In the present century glaciers are undergoing a marked retreat phase, but during the first half of the last century, and in the previous centuries, glaciers were more extended. For example, in 1860 the Forni Glacier was over 2 km longer, and the Fellaria Glaciers in Val Malenco reached the position at present occupied by the reservoir of Alpe Gera.

4. 1st Day. Itinerary: Milano - Como - Lake Como - Colico - Sondrio - Bormio

# THE END-MORAINE SYSTEMS OF LAKE COMO AND THE ORIGIN OF THE PREALPINE LAKES (A. Bini)

## 4.1 THE END-MORAINE SYSTEMS OF LAKE COMO

During the main glacial expansions all the alpine area was covered by an extensive ice field, with outlet glaciers ending in the piedmont plain with several ice lobes. The Adda Glacier, flowed in Valtellina and Val Chiavenna and in the valley now occupied by Lake Como. In the middle of the present lake it split into different tongues. One flowed in the direction of Val Menaggio and Lake Lugano and others flowed in the two branches of Lake Como and in Valsassina. At the foot of the mountains they formed coalescent piedmont lobes, namely the Faloppio, Como, Lambro and Lecco piedmont glaciers. Thirteen glacial drifts have been so far recognized south of the Alps, the two oldest referred to Late Pliocene. In the Lake Como moraine system only eight glacial drifts have been identified. Several $^{14}C$ dates are available for the last glacial maximum moraine system.

*The last glacial maximum (Lgm)* - The Lgm glacial drift is always identifiable from its geological and geomorphological characteristics and by $^{14}C$ datings. The Adda Glacier (BINI, 1987) reached Fino Mornasco and Cantù in the Como piedimont lobe; Lambrugo and Nibionno in the Lambro lobe, and Villa d'Adda in the Lecco lobe. Each lobe had at least one big meltwater stream, with the exclusion of the Faloppio that had a sub-glacial drainage.

During the glacial retreat, vast and shallow ice-contact and pro-glacial lakes formed in the Lambro, Faloppio and Como end-moraine systems. (ROSSI & *alii*, 1991)

## 4.2 ORIGIN OF THE VALLEYS NOW OCCUPIED BY THE ALPINE LAKES

The problem of the origin of the prealpine lakes, and mainly that of the valleys now occupied by the lakes (whose bottom is well below the sea level), has been discussed since the middle of the last century. Several theories have been put forward with a long debate on a tectonic versus a glacial erosion origin of the lakes. Finally the glacial origin of the lake basins was accepted by all the authors, as summarized by GABERT (1962) and NANGERONI (1956).

More recently, FINCKH (1978), BINI & *alii* (1978), FINCKH & *alii* (1984), CITA & *alii* (1990), CITA (1991), have proposed a different genesis for the alpine valleys. This is that the valleys were formed by river erosion following the drop in sea level of the Mediterranean Sea during the Messinian desiccation (HSU & *alii*, 1973). Afterwards the valleys were filled by marine and alluvial sediments, whilst the Plio-Quaternary glaciers modelled the valleys sides and partially eroded the sediments at the bottom.

*The Messinian model* - During the Messinian, the Mediterranean Sea remained isolated from the Atlantic Ocean because the connection between the two closed and the Straits of Gibraltar had not yet formed. Isolated, the Mediterranean dried out completely because the rate of evaporation was greater than the water supplied from the rivers[1]. At the bottom of the dry basin gypsum and salt were deposited and all the rivers outletting in the Mediterranean excavated long and deep canyons linking the river plains to the bottom of the basin (BARBER, 1981; CLAUSON, 1978, 1979, 1982, 1988). For example, the Nile excavated a canyon about 300 m deep that extended as far as Sudan with a total length of about 1200 km. Also the rivers of the southern side of the Alps, such as the Ticino, Olona, Adda, Brembo, Serio, Oglio, and Adige excavated deep canyons.

The River Adda excavated a deep canyon where the present Lake Como is located. The data presented in the following table are impressive (FINCKH, 1978).

| | Measures referred to the lake level | Measures referred to sea level |
|---|---|---|
| Present level of Lake Como | 199 m a.s.l. | |
| Present depth of Lake Como | 410 m | 211 m b.s.l. |
| Depth of the Messinian canyon | from 967 to 1097 m | from 756 to 886 m b.s.l. |

As can be seen, there is about 500-600 m of sediments between the bottom of the lake and the bedrock. The tectonic movements that affected the area of Lake Como in the Messinian forced the River Adda to excavate two valleys. These valleys are now occupied by the Lecco and Como branches of Lake Como. When the Straits of Gibraltar formed at the end of the Messinian, the Mediterranean basin was once again filled with sea water. The sea penetrated the canyons (up to the Sudan for the Nile) transforming them into rias. At the end of the Pliocene the glaciations started. The glaciers modelled the valleys, eroded part of the sediments and partially filled the canyons, but did not excavate the valleys deeper.

This model is supported by seismic reflection sections carried out across the valleys, the lakes and the continuation of the canyons to the south. One of the main objections that was made to the model of the Messinian origin of the valleys now occupied by the lakes is that the profiles of the valley floors of the lake tributaries are not cut down to the old canyon bottom level, but only to the present bottom of the lake. This was made after only superficial observation. In reality, the bottom of the tributary valleys of Lake Como can be related to the bottom of the old canyons. The Messinian valleys were filled with conglomerates (talus debris, alluvial fans, fan-delta) that were successively re-excavated, often in a slightly different location.

[1] The rate of evaporation in the Mediterranean is still greater than the supply of water from rivers. However, this deficiency is compensated by a great supply of water from the Atlantic Ocean through the Straits of Gibraltar.

5. 2nd day. Itinerary: Bormio - S.Caterina Valfurva - Val Cedèch - Valle dei Forni - Bormio

## HOLOCENE AND LATE GLACIAL HISTORY OF VALFURVA
(G. Orombelli & M. Pelfini)

### 5.1 INTRODUCTION

One of the environmental problems that has developed within recent decades are the environmental changes connected to the XX° century warming phase. The temperature increase during the present century has left strong evidence in the Alpine environment, especially at high elevations. The reduction in the surface area of the glaciers on the Southern side of the Ortles-Cevedale Group between the Little Ice Age and the present, has been calculated to be about 46-47% (PELFINI, 1992); whilst the rise in the equilibrium line altitude (Ela) has been of the order of 100 m. This can be related to an increase in the mean annual temperature of about 0.5 °C.

Valfurva (fig. 4) is one of the tributary valleys of Valtellina, which is located in the Province of Sondrio. To give some idea of the climate in the valley, the following meteorological data were recorded at the Forni meteorological station, which is located at an elevation of 2150 m a.s.l., next to a small dam:
Mean annual temperature: 1.5 °C; Mean temperature in January: –6.8 °C; Mean temperature in July: 11.1 °C; Days without thawing: 81; Days of frost: 25; Days without frost: 160; Mean annual precipitation: 807 mm; Days with rain 83.

### *The Upper Valfurva and the Forni Glacier*

The Ortles-Cevedale Group, of which the Valle dei Forni is a part, is the most strongly glacierized mountain group in the Italian Alps (DESIO & *alii*, 1967). From data

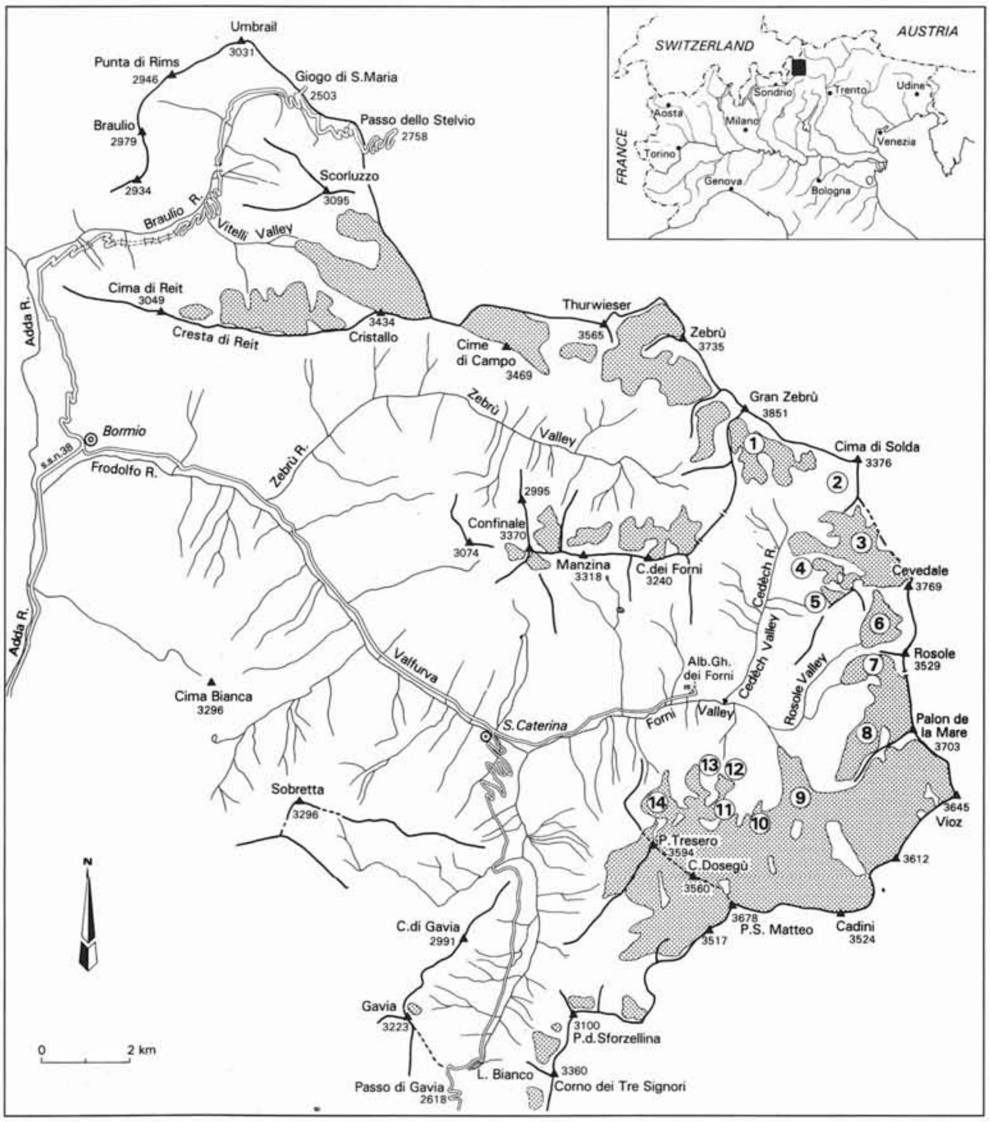



FIG. 4 - The Valfurva and the Ortles-Cevedale Group Glaciers (Lombardian sector). The numbers indicate the Forni valley glaciers and the tributary valleys of Rosole and Cedèch: 1) Val Zebrù; 2) Cima di Solda NW; 3) Cedèch; 4) Pasquale N; 5) Pasquale S; 6) Rosole; 7) Col di Lamare; 8) Palon della Mare; 9) Forni; 10) Isola Persa; 11) San Giacomo S; 12) San Giacomo E; 13) San Giacomo W; 14) Cerena.

collected during the first half of the 1980s by the Comitato Glaciologico Italiano (Cgi) for the World Glacier Inventory, in Valfurva twelve glaciers of various kinds and four glacierets are present. They cover a surface area of about 23 $km^2$ corresponding to about 35% of the total surface area of all Italian glaciers. Among the glaciers ten are small glaciers of various shapes, located in cirques or niches, such as the Gran Zebrù, Pasquale and San Giacomo glaciers. By contrast, the Cedèch Glacier is a valley glacier, whilst the Forni glacier is a composite valley glacier. The latter is formed from three accumulation basins and by several ice flows that gather in a single tongue, with the terminus at an altitude of 2420 m a.s.l. The Forni Glacier is considered the greatest valley glacier in the Italian Alps, with an area of approximately 13 $km^2$.

### *The Glaciers of Valfurva in the Late Glacial*

In Valfurva, long portions of Late Glacial lateral moraines deposited by the trunk valley glacier are preserved, as well as small moraine systems deposited by minor lateral glaciers. The frontal positions reached by the trunk glacier of Valfurva and the dating of the Late Glacial phases are uncertain. It is probable that the glacier of Valfurva underwent three or four stages during which the front was located on the Bormio plain, near S.Antonio Valfurva, near Santa Caterina Valfurva and at a inner position.

The last Late Glacial stage in the Alps is known as the Egesen stage and it is correlated to the Younger Dryas climatic event (11,000-10,000 years $^{14}C$ B.P.). At the end of this stage all the glaciers in the Alps retreated close to their present dimensions.

At an elevation of about 2400-2300 m a.s.l. in the Forni valley there are well preserved sub-parallel, moraine ridges, vegetated and with well-developed soils. These moraines are from a phase when the glaciers from Valle di Rosole and the Val Cedèch were tributaries of the Forni Glacier.

### *The Glaciers of Valfurva in the Holocene*

During the first half of the Holocene the extent of the glaciers reduced to minimum values, although with some uncertainty regarding the first two thousand years. In the second half of the Holocene the glaciers underwent a new period of expansion, which is called the «neoglaciation». During the general expansion of this period it is possible to identify many individual phases of advance and retreat. The best known advance phase is the Little Ice Age, which in the Italian Alps generally reached its maximum in the first half of the 19° century. In Valfurva the evidence of the Holocene glacial fluctuations is well preserved, including that of the maximum glacial extent during the Little Ice Age. Small moraines deposited in more ancient phases are locally preserved in lateral areas not reached by subsequent advances.

Inside of the moraines deposited during the maximum of the Little Ice Age, evidence of the front positions of the glaciers since the last century until the present are preserved. The last phase of advance ceased in the 1980s leaving small moraine arcs, which often still have an ice core.

## 5.2 The history of the Val Cedèch Glaciers

The Val Cedèch, orientated N-S, is about 6 km long and is on average 3 km wide. Glaciers are only present at the head of the valley and on the eastern side. At present there are four separate glaciers in the valley, which contrasts with maps from the 19th century that show a continuous cover of ice at the top of the valley with many lobes and tongues.

It is possible to distinguish two moraine systems in Val Cedèch (fig. 5). The first is the glacial deposits left by the two glacial flows of the Cedèch Glacier (A in fig. 5.2), the second is the moraine ridges deposited by the Gran Zebrù Glacier (B in fig. 5). The external end moraine of the Cedèch Glacier is from a phase when the glacier occupying the Val Cedèch was a single body, characterised by a well-developed tongue, into which the Gran Zebrù glacier also flowed. There is a grassland area in the upper part of the Val Cedèch that is completely surrounded by Holocene

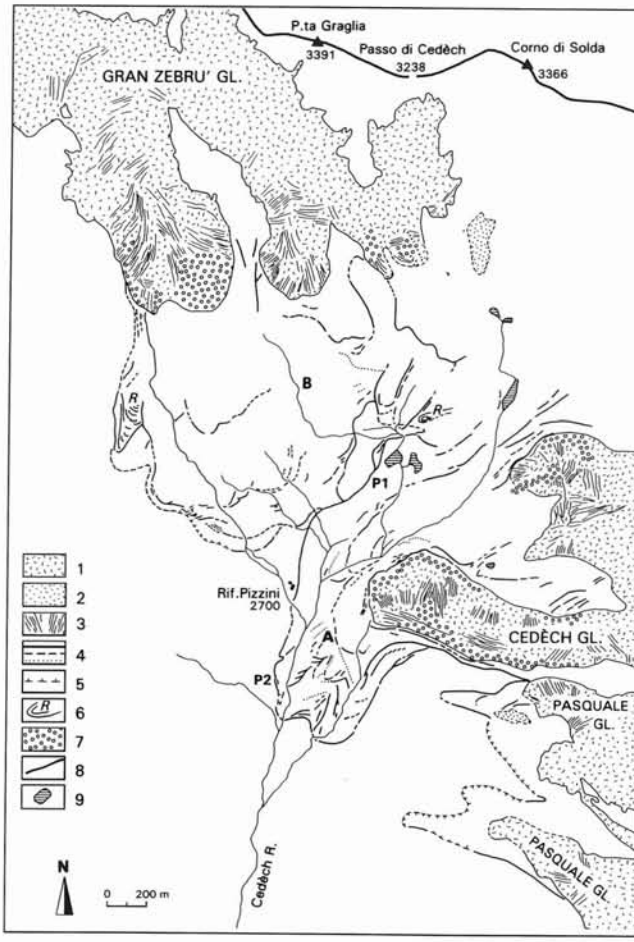



FIG. 5 - The glaciers and the moraines of Val Cedèch. A - moraine system deposited by the Cedèch Glacier; B - moraine system deposited by the Gran Zebrù Glacier. P1 and P2 - profiles opened into the lateral moraines, which will be seen in the stops 1 and 2. Legend: 1-glacier; 2-snow (nèvè); 3-crevasses; 4-Holocenie moraines (full line: very evident; dashed line: only present in fragments; dotted line: badly preserved); 5-limit reached by the Holocene glacial deposits; 6-rock glacier; 7-supraglacial debris; 8-main holocene moraines; 9-small lakes.

moraines. Based on the development of the soils and the lichen cover, it is possible to infer that from the Late Glacial onwards this area has remained free of ice.

Although Val Cedèch is cut into the «Bormio Phillades», a metamorphic rock formation, the glacial deposits of the western flow and, in part, those of the central flow of the Gran Zebrù Glacier are composed of carbonatic rocks. This is because the glacier erodes the sedimentary rock unit that forms the surrounding peaks.

The glaciers of Val Cedèch underwent the following stages of advance:

– In the early Holocene, or possibly at the end of the Late Glacial, the glacier had a tongue which was very similar to that of the Little Ice Age, only slightly wider and shorter. Slightly downhill of the Pizzini refuge a short length of the lateral moraine of the Cedèch Glacier is flanked by a partially buried, much older moraine, which was probably deposited earlier than 3920±85 $^{14}$C years B.P. (stop n° 2). Inside this moraine system it is possible to identify a number of small arcs that define the morphology of the glacier fronts during the numerous phases of advance.

– A phase of advance with an importance comparable to that of the Little Ice Age occurred around 2600-2800 $^{14}$C years B.P. and it is documented by buried soils.

– The maximum expansion of the Holocene occurred during the Little Ice Age. This advance probably took place during the first half of the last century.

– It is not possible to date the successive advances with certainty but the glacial deposits highlight at least two phases. The first phase could have occurred at the same time as the one generally registered by the Alpine glaciers at the end of the last century. The second phase is probably attributable to the 1920s.

– The last advance phase is the one recorded in the 1980s. This is evidenced by recent moraines bordering the present day glacier fronts, which sometimes have an ice core.

### 5.3 The Valle dei Forni and the recent history of the Forni Glacier

The Holocene moraine system is well preserved in the Valle dei Forni (fig. 6), and it is possible to distinguish three main moraine groups that were deposited in progressively more recent times. They are different from each other and the discriminating factors are soil development, degree of grass cover, lichen cover, and stoniness. The outer moraine system (A in fig. 6) suggests a thick glacial tongue that reached the plain which is located downhill of the Forni car park. This tongue was about 100 m thick at the location of the Branca refuge.

The internal moraine systems are very evident; one of them runs parallel to the Cedèch Torrent (C in fig. 6). The most internal moraine system, in a frontal position, is characterised by many small sub-parallel ridges (D in fig. 6).

The maximum Holocene expansion of this glacier is documented by a short frontal moraine ridge, located in the vicinity of the Forni refuge. During this advance the Forni Glacier dammed a small *roches moutonnées* basin, which was successively filled by peat deposits. The $^{14}$C dating of a bottom sample of these deposits, shows that the maximum expansion of the glacier occurred immediately

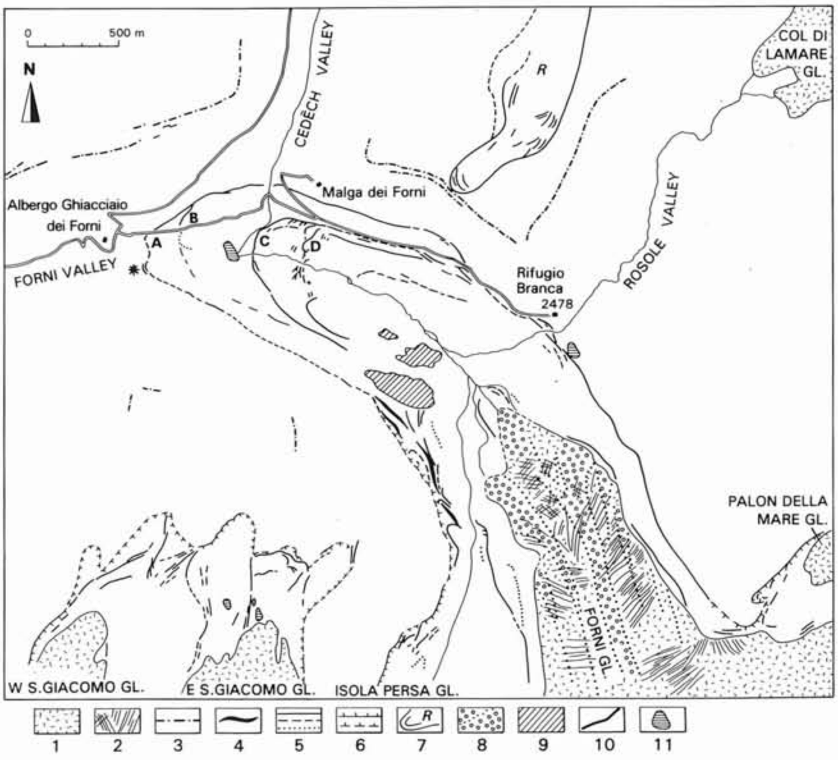



FIG. 6 - Geomorphological sketch of the moraines deposited by the Forni Glacier during the Holocene. A-moraine deposited at the maximum of the Little Ice Age; B-moraine deposited around the 1890s; C-moraine deposited in the first 15 years of this century; D- moraines deposited during the 1920s advance (1926); *-moraine deposited before 2670 ± 130 $^{14}$C yr B.P. Legend: 1-glacier; 2-crevasses; 3-Late Glacial moraines; 4-moraines deposited during the early phases of the Little Ice Age; 5-moraines of the Little Ice Age (full line: very evident; dashed line: only present in fragments; dotted line: badly preserved); 6-limit reached by the Holocene glacial deposits; 7-rock glacier; 8-supraglacial debris; 9-*roches moutonnées*; 10-main Holocene moraines; 11-small lakes.

before 2670 ± 130 $^{14}$C years B.P., corresponding to a calibrated age of 930-710 B.C. (OROMBELLI & PELFINI, 1985) (* in figs. 6 and 7). The extent of this advance is comparable to the following one during the Little Ice Age.

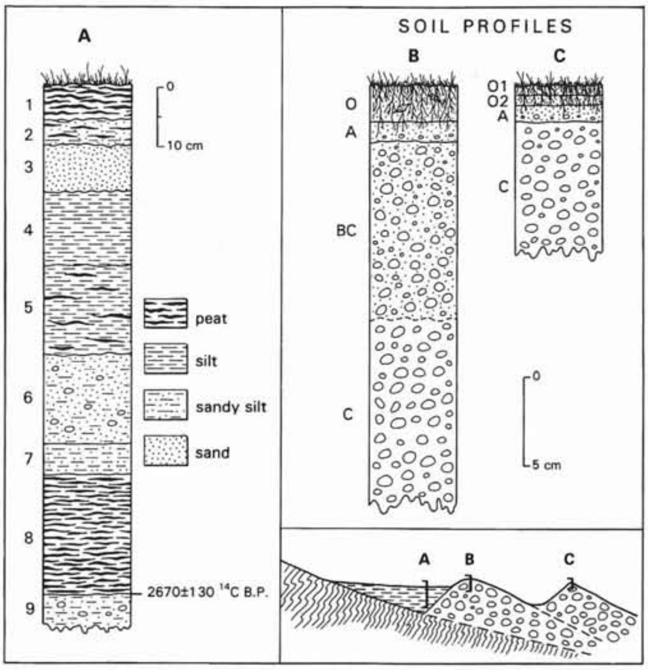



FIG. 7 - Section cut in the depression dammed by the most external frontal moraine of the Forni Glacier. The peat bottom layer has returned an age of 2670 ± 130 $^{14}$C yr B.P., (calibrated age of 930-710 B.C.).

The only evidence for the initial phases of the Little Ice Age is small segments of lateral moraines that are attributed, by lichenometric analyses, to the second half of the 17th century and to the 1820s. The maximim advance of the Little Ice Age for the Forni Glacier is represented by a complete moraine arc, which is recognisable along the whole Forni Valley (A in fig. 6). The dating of this moraine is not certain, as the historical documents are not precise. However, it seems more than likely that the maximum expansion during the Little Ice Age took place in the early years of the second half of the 19th century (PELFINI, 1988; 1992).

On the basis of the historical documents and field surveys, a time/distance curve (fig. 8) was constructed. This curve summarises the history of the Forni Glacier, the main phases of which are as follows.

– In 1833 (topographic map of the Regno Lombardo Veneto, 1833) the Forni Glacier was retreating. This is shown by the fact that the Cedèch torrent flowed into the Frodolfo torrent.

– In 1844 (Mappa Catastale del Comune Censuario di Valfurva), during an advance phase, the Forni Glacier blocked the Cedèch valley outlet. This phase may have ended with the deposition of the most external moraine.

– In 1864, when the glacier was undergoing a retreat phase (STOPPANI, 1865; 1880), a glacial outburst occurred.

– In 1867 (Originalkarte der Sudlichen Ortler-Alpen, by J. Payer) the front of the Forni Glacier still reached the Cedèch torrent.

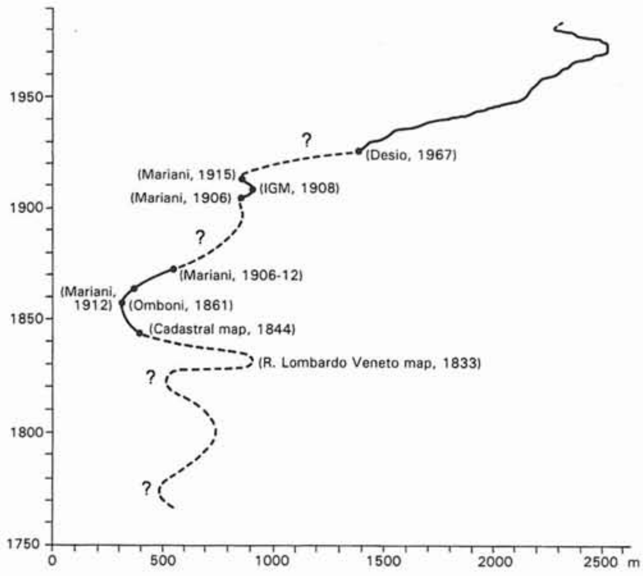



FIG. 8 - Time/distance curve for the Forni Glacier.

– In 1885 (Maps of the Istituto Geografico Militare, scale 1:25 000) the glacier was higher, thicker and once again dammed the Cedèch valley, confirming a new advance stage.

– Over time, there is more documentation of the history of the glacier. Around 1891-94 the glacier terminus looks thinner again, but the glacier tongue is still characterised by a notable thickness.

– Between 1898 and 1904 the retreat process ceased and the glacier advanced slightly. On the 9th August 1911 a second glacial outburst occurred, followed by a small advance in 1913-14 (MARIANI, 1915). The two advances of 1904 and 1913-14 probably reached about the same position. The moraine system located in the vicinity of the Cedèch torrent outlet was certainly deposited at the beginning of this century. The illustrations of the glacier during the first decades of this century start to show a flattening of the central part of the front and the evident presence of medial moraines.

– Between 1915 and 1925 the history of the Forni Glacier is not documented, but the glacier must have undergone a strong retreat, followed by a new advance phase.

– In 1925-26 DESIO (1967) witnessed an advance of the glacier that deposited the most internal moraine system. Starting in 1925-26, measurements of the frontal variations have been carried out yearly by the CGI (Italian Glaciological Committee).

– After the advance phase of the 1920s, the history of the Forni Glacier is characterised by a continuous retreat. In the 1960s the front is located above the rock cliff that is clearly visible from the Branca refuge. However, after 1970 the retreat phase ended and from 1974 to 1981 Forni Glacier advanced over 200 m, descending the rock cliff and stretching out at its base with a thin tongue.

– As with most of the Alpine glaciers, the Forni Glacier since the 1980s has also experienced a drastic retreat, which has taken the front back to the position it had in the 1960s.

*STOP 1 - The Little Ice Age in Val Cedèch* - At the head of the valley a grassland area is visible; this is bordered by the Holocene moraines that were deposited by the two glaciers. The development of the soil, the weathering of the exposed boulders and the intense lichen cover on the boulders, suggests an early Holocene age for the soil. A number of sections opened in the moraines that border the grassland area showed buried soils (P1 in fig. 5) which have been radiocarbon dated. The age obtained indicate that

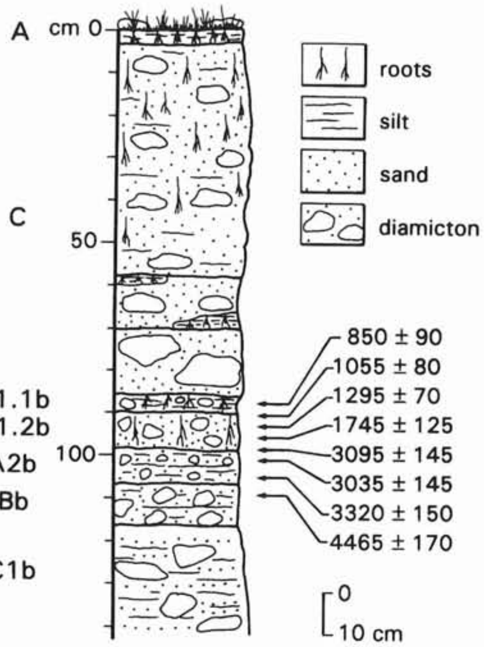



FIG. 9 - Profile cut at the edge of the area which has not been reached by the Val Cedèch glaciers during the Holocene.

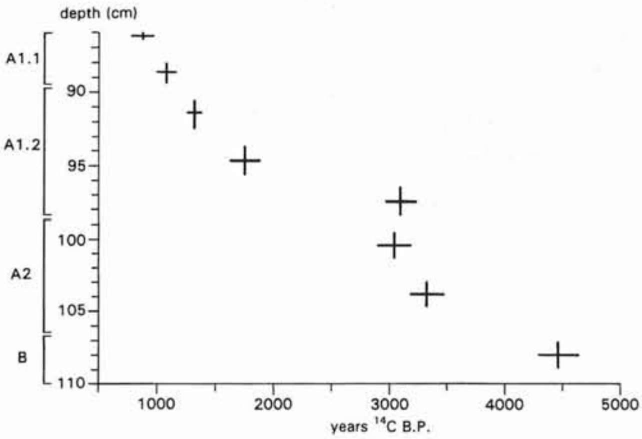



FIG. 10 - Age/depth gradient in the profile of FIG. 5.6.

the moraines belong to the Little Ice Age. Thin layers of soil at increasing depth were dated. The $^{14}$C datings showed high age/depth gradients, (figs. 9 and 10), which are about 120 years cm$^{-1}$ if only A horizon is considered, and about 176 years cm$^{-1}$ if the upper centimetre of B horizon is also considered. The buried organic layer started to develop from at least 4465±170 $^{14}$C years BP. Hence, the soil is older than that date.

The moraine that buried the soil was deposited by the Cedèch Glacier after 850 ± 90 $^{14}$C years BP and after 540±80 $^{14}$C BP, the youngest age obtained for the buried soil in a nearby profile. This advance occurred during the Little Ice Age, but its age could be much younger than the $^{14}$C dates reported above.

*STOP 2 - A pre-Little Ice Age glacial advance phase* - Downhill of the Pizzini refuge the maximum Holocene expansion moraine is flanked, for a short length, by a more external moraine. This is different from the previous moraine in both the morphological and lithological aspects (P2 in figs. 5 and 11). A first dating was performed on the crest of the older moraine by sampling the basal portion of the A horizon of the top soil. This provided a date of 890 ± 90 $^{14}$C yr B.P. A second profile was opened on the lateral moraine superposed to the previous one. Four samples were taken, which gave the dates shown in fig. 11. In this profile an age/depth gradient equal to 108 years cm-1 was obtained for the buried soil, or 160 years cm$^{-1}$ if the second buried horizon is also taken into account (fig. 12).

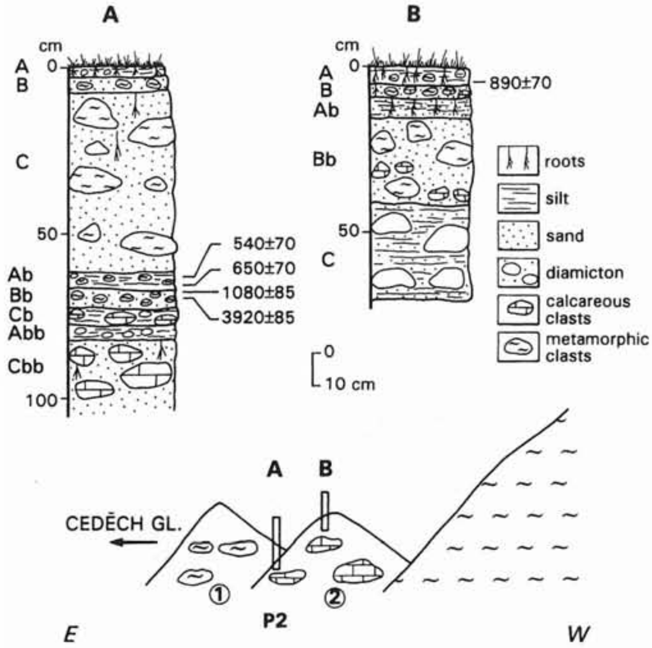



FIG. 11 - Sketch of the moraines P2 of FIG. 5.2 and relative cross sections.

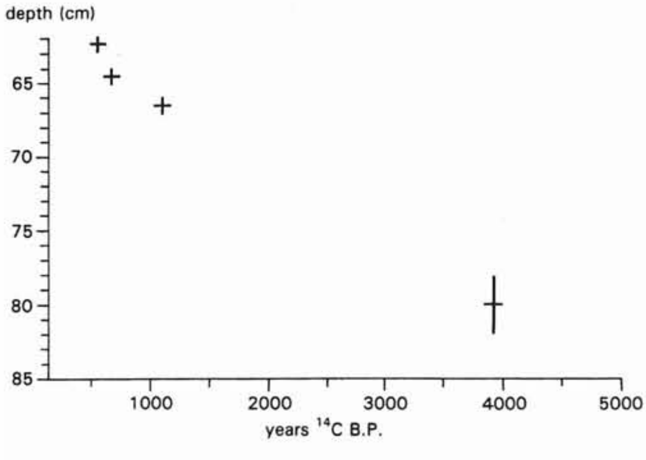



FIG. 12 - Age/depth gradient of the section P2 of FIG. 5.8.

The $^{14}C$ ages show that the deposition of the western lateral moraine of the Cedèch Glacier cannot be older than $540 \pm 70$ $^{14}C$ yr B.P., which confirms that it took place during the Little Ice Age. It was probably deposited in the middle of the last century, as also confirmed by lichenometric analyses. The first dating of the most external moraine showed that this had to be older than $890 \pm 70$ $^{14}C$ yr B.P. Also taking into account the information obtained from the second profile, it was inferred that the most external moraine was deposited earlier than $3920 \pm 85$ $^{14}C$ yr B.P. This implies that a glacial advance phase took place previously during the Holocene or, more likely, in the Late Glacial, as suggested by the different lithological composition, following dynamic conditions of the glacier different from the present. The dominant feature of the Cedèch Glacier was probably the glacial flow coming from the western side of the valley, which was responsible for the transport of the limestone clasts. By contrast, in the Little Ice Age the main flow was the one coming from the eastern side, which carried metamorphic material.

*STOP 3 - Forni lateral moraine deposited during the Lia maximum.* - After leaving Val Cedèch we enter the Forni valley, were the glacial deposits of the last and present centuries are very evident. We will stop where the road going to the Branca refuge crosses the most external Holocene lateral moraine of the Forni Glacier. Although the moraine is of small dimensions it is continuous along the northern side of the valley. It is characterised by a heap of big blocks and is fairly extensively covered by grass. Its chronological attribution to the early second half of the 19th century is based on historical and cartographic documents, and on lichenometric measurements. The contrast between this moraine and the impressive Late Glacial moraine located uphill allows one to appreciate the altitude reached by the glacier in the final phases of the last glaciation.

The lateral moraine is easily followed to where it is interrupted by the gorge of the Cedèch torrent. The moraine continues on the other side of the gorge indicating that during this phase the Cedèch torrent was forced to excavate its own course underneath the Forni Glacier.

Along the road that leads up to the Branca refuge, surfaces subject to glacial erosion in the final phases of the last glaciation can be observed. Weathering processes during the Holocene almost completely erased the glacial striations and scratches, leaving rough surfaces now extensively covered by crustose lichens. The soils that formed in the deglaciated areas from the end of the Pleistocene also show well-developed profiles. This situation contrasts with that in the areas glaciated during the Little Ice Age, which are characterised by glacial erosional surfaces with well preserved striations and incipient soils.

*STOP 4 - The Forni Glacier* - The Branca refuge is a perfect viewpoint for observing the Forni Glacier's front, its ablation area and parts of the high accumulation basins. The lateral moraine that starts at an elevation of 2700 m can be seen. It is also possible to observe the valley bottom which was progressively deglaciated during the last century and the lateral moraine that, near to the refuge, creates a small dammed lake at an elevation of 2452 m. This lake is progressively reducing its size due to the accumulation of alluvial deposits and the erosion of the damming moraine.

It is possible to observe the internal structure of the lateral moraine, which has been partly demolished by landslide events. It is characterised by a rough stratification due to the alignement of the coarser clasts along dipping surfaces.

The area of the present front that undergoes rapid evolution is of particular interest. During the last small advance (1970-1986) the glacier passed the steep rock slope and formed a short tongue on the valley plain. At the bottom of the cliff it is possible to observe a small but regular lateral-frontal moraine ridge, which was deposited during the last advance. The front has retreated back up to the edge of the rock cliff now, to the position held in the 1960s.

At a higher elevation it is possible to observe the late glacial moraine that consists of many parallel ridges. At the confluence with Val Cedèch the lateral Late Glacial moraine of the Forni Valley joins at a right angle the lateral moraines of the Val Cedèch. The Late Glacial moraines form a sort of terrace hanging above the Forni valley and they continue past the Cedèch valley trough, descending gently towards Santa Caterina Valfurva.

*STOP 5 - Confluence with the Val Cedèch* - On the valley floor the lateral moraine that was deposited in the first years of the 20th century is visible. This moraine ridge that flanks the road to the Branca refuge was deposited in a period that is not possible to date with precision. However, on the basis of historical and cartographic documents, it can be attributed to the first 15 years of this century. It is different from the most external moraine of the Little Ice Age because of a less extensive cover of grass and of the smaller diameter of lichen. When this moraine was deposited the Cedèch torrent was free to flow into the Frodolfo torrent. It is also possible to observe along the valley bottom the most internal moraine arc, which was deposited in 1926 (DESIO, 1967) at the end of the advance of the 1920s. This moraine ridge has no grass cover and very small lichens.

The point where the bridge is in the Cedèch valley bottom has never been reached by the glaciers coming from the head of the Cedèch valley during the Holocene. At a higher elevation a lateral-frontal moraine is present, which can be attributed to the early Holocene or to the final phases of the Late Glacial. By contrast, the outlet of the Val Cedèch into the Frodolfo torrent has been repeatedly blocked or diverted by the fluctuations of the Forni Glacier's tongue until the first half of the 20th century.

*STOP 6 - Holocene frontal moraines* - During its maximum Holocene expansion, the Forni Glacier deposited two parallel small moraine arcs. The most external moraine was dated $2670 \pm 130$ $^{14}C$ yr B.P. The age was obtained by dating the deepest layer of peat found inside a small depression that had been dammed by the moraine (* in figs. 6 and 7). This is a minimum age and agrees with the age of an important glacial advance phase, which has been recognised in many mountain chains in both hemispheres. The most internal small arc represents the maximum expansion of the Little Ice Age and was deposited around 1859.

On some rock surfaces in the vicinity of the moraine it is possible to see an evident trim line separating two areas with different weathering and lichen cover. The boundary is coincident with the limit of the glacial front's expansion during the Holocene. On a big *roche moutonnée* located in the valley floor glacial striation and scratches are still preserved as well as pothole carved by subglacial water erosion. The location of these last features suggests their origin was linked to subglacial melting water, produced by a pressure increase uphill of an obstacle.

In the fluvioglacial plain gravel and rounded boulders can be observed, which are frequently disposed as longitudinal bars. One of these, about 2 m high and composed of big rounded blocks up to 1 m diameter, can be attributed to a glacial outburst. An event of this kind occurred in August 1911.

## 6. 3rd Day. Itinerary: Bormio - Aprica - Passo del Tonale - Presena - Val di Genova - Pinzolo

## GEOMORPHOLOGY OF THE ADAMELLO GROUP
(C. Baroni & A. Carton)

### 6.1 INTRODUCTION

Valcamonica takes its name from the prehistoric people named Camuni, who were brought under Roman domination by the Emperor Augustus in the year 16 B.C. Valcamonica is one of the widest Alpine valleys and has an area of about 1450 $km^2$. The valley is cut by the River Oglio, which flows for about 80 km from Passo Gavia (at the southern border of the Stelvio National Park) and Passo del Tonale to the Lake Iseo. The valley runs through the mountain groups of the Adamello, the Alpi Orobie and the southernmost edges of the Ortles Cevedale Massif, and shows very strong evidence of glacial morphology. The highest peak of the Valcamonica basin is Monte Adamello at 3539 m a.s.l., which represents the maximum height of a massif composed of numerous peaks, all over 3000 m high. The first human settlements date back to the Late Palaeolithic age, although the valley was intensely populated between the Neolithic and the Iron Age. The magnificent stone carvings sculptured on the *roches moutonnées* witness the early human presence in the valley. There are more than 40 000 of these carvings and they are particularly concentrated around the area of Capo di Ponte.

From a geological point of view, a great part of the upper valley that leads to Passo del Tonale shows once again one of the main tectonic features of the Alps, namely the Insubric Line (fig. 2). In this segment of the line, the Southern Alps and the Austroalpine units come into contact. The slopes to the right and the valley floor uphill of Monno-Incudine are made up of metamorphic rocks of the Austroalpine. These rocks are associated with small magmatic bodies, the intrusion of which is dated towards the end of the Palaeozoic. The middle-lower parts of the left side of the valley and the valley floor are made up of metamorphic rocks of the Southern Alps crystalline basement, whilst the upper part of the left side of the valley is composed of intrusive rocks of the great tertiary pluton of the Adamello.

As far as the Quaternary deposits are concerned, those particularly worth mentioning are the glacial deposits dated to Late Glacial phases; it is possible to observe them on the hydrographic left side of the valley. End-moraine systems, in a series of concentric arcs, are very evident at the outlet of the tributary valleys of the Oglio river: Val Paghera (near Vezza), Val di Vallaro, Valle dell'Avio, Val Seria and Val Narcanello. Particularly evident are those near Temù and Ponte di Legno, dated to the Gschnitz stadium (CASTIGLIONI, 1961), but the ones in the area of the Passo del Tonale are also fairly well developed.

### 6.2 THE ADAMELLO MASSIF

The intrusive massif of the Adamello is the greatest pluton of Alpine age in the Alps. Given the great extent of the outcrops of its magmatic rocks (about 670 $km^2$, CALLEGARI, 1983 and 1985) it is considered to be a batholith. The magmatic mass is enclosed in a crustal structural wedge bounded to the north by the Linea del Tonale, the local name for the Insubric Line, which separates the Austroalpine from the Southern Alps, and to the west by the «Linea delle Giudicarie». Only locally do these tectonic elements directly affect the intrusive rocks. The presence of faults that cross the intrusive body and of regular sets of joints suggest that the area where the intrusive massif is located has been affected by recent deformations. Sets of fractures are mainly orientated NW-SE, WNW-ESE, NE-SW across the massif, frequently affecting the morphology of the valleys next to the rock threshold, some of the top surfaces, and the ridges with jigsaw profiles (arétes)). The presence of these fractures is also highlighted locally by the particular morphology of the *roches moutonnées*.

The setting of the Adamello intrusive rocks took place at different stages during the Alpine orogeny, between late Eocene and late Oligocene. Different lithologies can be distinguished (BIANCHI & *alii*, 1970; CALLEGARI & DAL PIAZ, 1973), as the Adamello batholith is made up of many plutons that are more or less differentiated. The intrusion of these bodies took place at different times in a period between 42 and 30 MA ago, according to Rb/Sr and K/Ar radiometric determinations on micas and amphiboles of the igneous rocks (DEL MORO & *alii*, 1985). The oldest ages refer to the lithologies outcropping in the southernmost area. The main outcropping rocks are tonalite, quartz diorite, granodiorite, etc. (BIANCHI & *alii*, 1970; CALLEGARI & DAL PIAZ, 1973):

- North-Eastern Presanella Tonalite: it has the same chemical composition as the Western Adamello Tonalite.
- Lago d'Arno-Lago Boazzo (Monte Bruffione) Biotitic Granodiorite: a granite-like rock with porphiric structure.

– Val Fredda leucoquartzdiorite: a granite-like rock, medium to coarse grained.
– Central Adamello Biotitic Quartz-diorite.
– Biolitic Quartz-diorite Val d'Avio type.
– Lower Val di Genova Biotitic Quartz-diorite.

Aplitic and pegmatitic dykes and sills are widespread all over the massif where they cut through the «hosting» rocks. The magmatic rocks intrude and deform the metamorphic rocks of the crystalline basement (Scisti di Edolo, gneiss, schists, slates) and the Permian-Mesozoic formations of the Southern Alps, showing uncomformity intrusive contacts (CALLEGARI, 1983). At the edges of the pluton a contact rim is present, highlighted by numerous typical minerals (MOTTANA & SCHIAVINATO, 1973).

*STOP 1 - Passo del Tonale (1883 m a.s.l.)* - The Passo del Tonale separates the two mountain groups of Adamello-Presanella to the South and of Ortles-Cevedale to the North. The pass is crossed by the water divide between the basins of River Po (Oglio) and River Adige (Noce), which represent the first and the second most important rivers in Northern Italy, respectively. The pass is 1-2 km wide, fairly plain and about 4 km long, and its presence is due to the previously mentioned structural feature of alpine importance, the Insubric Line, that in this segment is called «Linea del Tonale». During the last glacial maximum, the Tonale area was covered by ice up to an elevation of 2400 m, and the two basins of the Oglio and the Noce rivers were divided by the glacier (PENCK, 1909; CASTIGLIONI, 1961). Moraines and kame deposits dated to the oldest retreat stages are present on both sides of the valley. According to CASTIGLIONI (1961), the Late Glacial (Gschnitz and Daun stadiums) glaciers were confined inside the Presanella and Ortles-Cevedale massifs and the main valley was completely ice free.

Some sections in the Passo del Tonale area and at Pian Venezia (Ortles-Cevedale group) provided new $^{14}C$ dating that allows a better characterisation of the palaeoenvironmental evolution of the area (BARONI & *alii*, 1994; 1996). In particular, one section dug near the Tonale pass (fig. 6.1) gave a minimum age for the glacial deposits outcropping there, according to which these deposits are older than 12,275 ± 115 $^{14}C$ yr B.P. (GX-21040 AMS). At this time, the sedimentation of organic matter (moss peat with dwarf willows) started on the top of a layer formed by alluvial gravel and debris flow sediments, interpreted as an alluvial fan deposit (with material coming from the north). The organic sedimentation stops at around 10,170 ± 85 $^{14}C$ yr B.P. (GX-21040 AMS), because of the deposition of debris flow material. Peat deposits with sfagnum follow and successively peats with macroscopic remnant and trunks in natural position. In another nearby section (fig. 13), the organic sedimentation was dated at 9895 ± 170 $^{14}C$ yr B.P. (GX-19712); at the same time a forest soil representing the

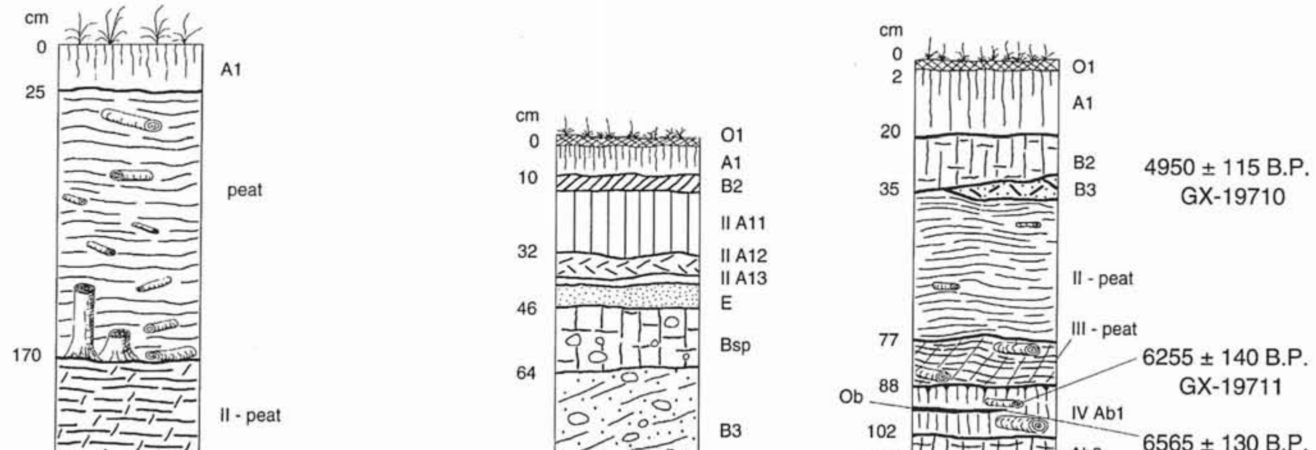



FIG. 13 - Passo del Tonale (1883 m), detailed stratigraphic profiles.

start of the Holocene formed in this area BARONI *et alii*, 1994). The corrected age is slightly older if the residence time of the analysed soil is taken into account.

If the oldest organic sedimentation described above is assigned to the Allerod interstadial, its starting time would be slightly earlier than that reported in literature. The Allerod interstadial, together with the oldest part of the Younger Dryas would be represented by peats with moss and dwarf willows (RAVAZZI, personal communication). By contrast, the upper part of the Younger Dryas would show the presence of slope instability phenomena with the development of a debris flow fan.

From the start of the Holocene (9895±170 $^{14}C$ yr B.P.) marsh areas are documented at the Passo del Tonale. A podsol developed at the borders of the oldest marsh areas and was successively buried by a peat layer referred to the Atlantic period (fig. 13).

The pollen analyses (RAVAZZI in BARONI & *alii*, 1994, 1996) suggests that, starting from the early Holocene, a widespread forest of *cembra* pine and birch covered the area of the Tonale pass. After 9895 ± 170 $^{14}C$ yr B.P., changes in environmental conditions lead to the formation of peat deposits which progressively extended uphill and buried the podsol, which in turn was set on top of the glacial deposits.

Moreover, at Pian Venezia and in Val di Pejo (SPERANZA & *alii*, 1996) it is confirmed that in the Preboreal an extensive forest (*cembra* pine and birch) covered the Central Alps, reaching 2300 m in altitude at the end of the Preboreal period. The data collected in the upper Val di Pejo also confirms the rapid rise of the tree line in the Italian Alps at the beginning of the Holocene. This contrasts with the data known so far for the Southern Alps.

The abundant macroscopic remnants of spruce and larch, which characterise the central portion of the peat deposit, disappear in the upper part at a date earlier than 4950 ± 115 B.P. $^{14}C$ yr B.P. (GX- 19710). This suggests that two different evolution stages of the peat deposit took place. The widespread presence of coal in the upper part of the deposit suggests a degradation of the forest, possibly due to human causes, as early as the Atlantic period. The peat deposit ceases at the end of the Atlantic or slightly afterwards.

*From the start of the cable car to the end of the chair lift*

The Paradiso cable car allows us to go up about 700 m in elevation, from Passo del Tonale to the Soldanella refuge (2587 m a.s.l.). After, a chair lift takes us up to the base of a wide glacial cirque reaching the Capanna Presena, just beneath the Western Presena Glacier, on which summer skiing takes place. Using appropriate transport it will be possible to reach the Passo Maroccaro, a splendid viewpoint on the greatest glacier of the Italian Alps: the Adamello Glacier.

After taking the chair lift to the Capanna Presena, passing two small lakes, it is possible to observe the limit of the maximum Holocene glacial expansion of the Vedretta del Presena (part of the main glacier). The limit is detectable at the transition between the debris deposits of glacial origin, and the bare outcropping rock.

*STOP 2 - Presena Glacier* - The Western Presena Glacier (65 hectares) fills a North-facing cirque and is surrounded by peaks with elevations higher than 3000 m. The fairly constant slope and the fact that there are very few crevasses favours the skiing activity in this area, even in the summer, although sometimes at limit conditions. Until a few decades ago this glacier and the Vedretta del Presena, located to the East, formed a continuous body of ice, which had an extent of 400 hectares in about 1850. This glacier is highly affected by human activity, not only by skiing but also by excursions and ski-mountaineering, being a passage to the Conca del Mandrone and the Val di Genova. For these reasons the Western Presena Glacier can no longer be considered to be only a product of the natural environmental conditions. The preparation of the ski routes involves the movement of snow from the accumulation area to the ablation area and an artificial snow making plant has been used for a few years. Artificial snow making may partially counterbalance the intense ablation of the Vedretta del Presena.

Historical curiosity: during World War One this glacier hosted the first battle ever fought on a glacier.

*From Capanna Presena to Passo Maroccaro*

Travelling up towards Passo Maroccaro the Vedretta del Presena and the Vedretta della Busazza can be observed in the distance. Their very evident end moraines can be seen at their ablation fronts. They were deposited during the maximum expansion of the Little Ice Age, which occurred during the second half of the last century.

*STOP 3 - Passo del Maroccaro* - Looking down from the Passo del Maroccaro towards the Conca del Mandrone, there is a wonderful view on what is considered to be the heart of the Adamello massif. The Monte Adamello peak (3539 m a.s.l.) is surrounded by a wide, glacialised land. Different glaciers flow more or less radially towards the neighbouring valleys.

The Adamello Glacier is a plateau glacier and it is formed by several individual parts. As a whole, the Adamello Glacier is the greatest of the whole Italian Alps (1813 hectares) and can be classified as a summit glacier of Scandinavian type (MARSON, 1906; SERVIZIO GLACIOLOGICO LOMBARDO, 1995). On the basis of detailed maps, some dated positions of the front of the Ghiacciaio del Mandrone have been reconstructed. These have also been used for checking the time-distance curve obtained from glaciological surveys. The various maps used show that the glacier's front has progressively retreated from the Malga Venezia plain, going back in successive stages along the valley deeply cut into the steep slope located downhill of the Acquapendente threshold. In the 1920s the glacier stabilised with a saw-toothed front uphill of the same edge. The total retreat, inferred from the cartographic documents of the period between 1864 and 1983, has been estimated to be about 1800 m (fig. 14).

The first precise data on the frontal variations measured during glaciological campaigns was between the end of the last century and the first years of this century. Through analysis of the data it is possible to detect, within a genera-

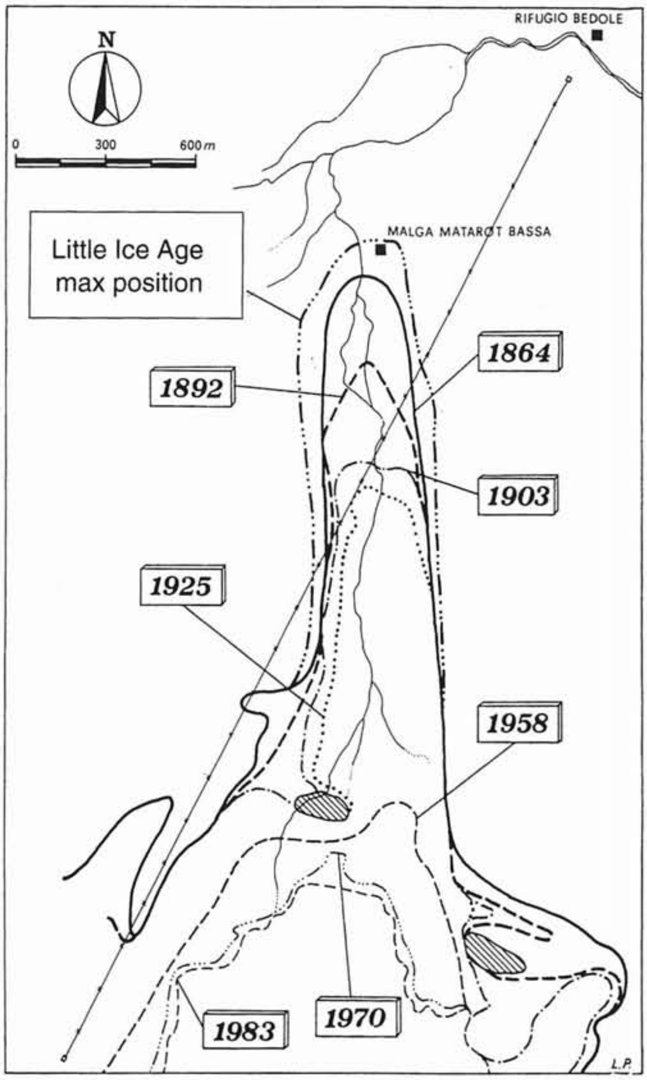



FIG. 14. - The Lobbia Glacier positions since 1864 A.D. as reconstructed from carthographic documents. 1864: Originalkarte der Adamello-Presanella Alpen (PAYER, 1865). 1892: Specialkarte von Oesterreich-Ungarn, zone 21 (K.U.K. Militaer-Geographischen Institutes). 1903: Karte der Adamello und Presanella Gruppe (Deutsch un Oesterreich Alpen Verain). 1925: Tavoletta IGM, Cima presanella, F° 20 IV SE, 1931 edition. 1958: Catasto dei Ghiacciai Italiani (Comitato Glaciologico, C.N.R., 1961). 1970: Tavoletta IGM, Cima presanella, F° 20 IV SE, 1973 edition. 1983: Carta Topografica della Provincia Autonoma di Trento.

lised retreat phase, some short stages of slight advance or stability of the front: 1885-1895 (not measured); 1933-1941 (+47.5 m); 1952-1953 (+1 m); 1958-1961 (+9 m); and 1973-1980 (+24.5 m). (fig. 15)

The comparison of the time-distance curves (fig. 15) of the Lobbia and Mandrone glaciers highlights a similar behaviour during the period 1885-1895, which was followed by an unmeasured stability stage slightly before World War One. The remarkable advance of the Lobbia glacier of +42 m recorded in the period 1919-23 is in contrast to the Mandrone glacier's opposite behaviour. The advances of the 1930s (and early 1940s) and 1970s are basically recorded on both glaciers, even if the Mandrone glacier advance was more consistent. In the first period mentioned the Mandrone glacier started to advance six years before the Lobbia glacier, whilst in the second period the advance phase dies out in 1980 rather than in 1985, but is much more marked. By contrast, the advances of the 1950s and 1960s are completely out of phase.

The Mandrone glacier's present equilibrium line altitude (Ela) has been computed on the basis of the Carta Topografica Generale of the Provincia Autonoma di Trento to be 2994 +18/-23 m a.s.l. This value has been obtained by considering an accumulation area ratio (Aar) of 0.67 ± 0.05. By applying the same parameter to the whole surface of the Adamello Glacier, the Ela becomes 3014 +17/-18 m a.s.l. If an Aar of 0.6 ± 0.05 (PORTER, 1975; NESJE & DAHL, 1991) is used instead, which probably represents a more representative value for these kinds of glaciers, a significant rise of the Ela is obtained. In the case of the Adamello Glacier the Ela would be 3038 +18/-7 m a.s.l.

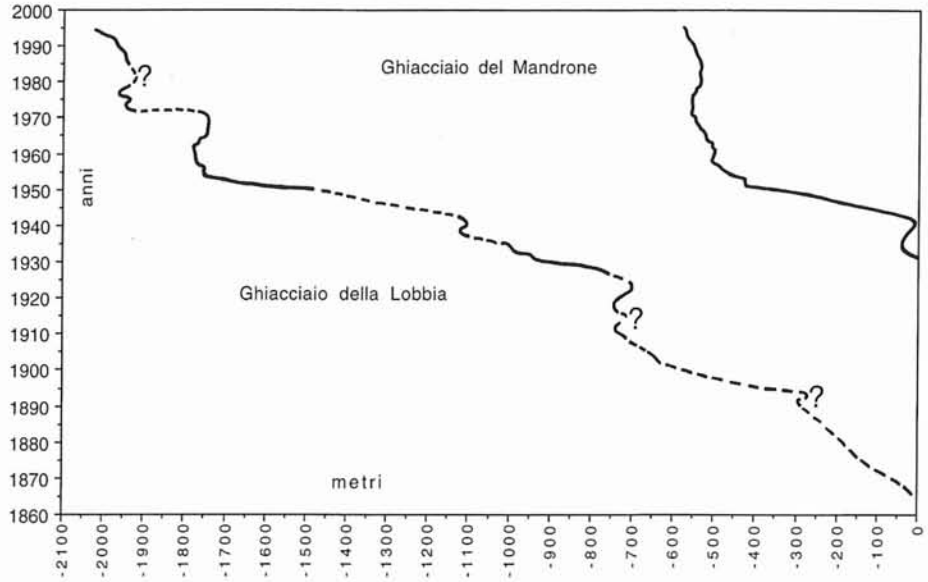
Therefore, it is possible to notice a difference between the Ela of the Lobbie and Adamello glaciers of 78 m, considering a Aar of 0.6 and of 70 m, considering a Aar of 0.67. It is important to emphasise that in the last few years the snowline on the Adamello massif has been progressively rising, uncovering in summer large areas of the Adamello glacier and almost all of the Lobbia-Fumo glacier. Using the above mentioned data, it seems that the Ela of the Adamello glacier is the most representative of the snowline, which in this massif is at present well over 3000 m a.s.l.

In the 1960s the Comitato Glaciologico Italiano (Cgi) carried out some geodetic and geophysical research in the area. As part of these studies a photogrammetric survey and geophysical measurements of the ice thickness at Pian di Neve were performed CARABELLI (1964). It was calculated that the maximum thickness of the glacier was about 260 m, the thickness in the area of the Passo di Adamè was 150 m, and that it was possible to reconstruct the shape of the rock substratum over a wide sector of the study area. Recent geophysical surveys (1991) have especially concentrated on the definition of the ice thickness and the shape of the rock substratum of the Lobbia and Adamello glaciers. This showed in particular that in the Passo Adamè area the ice thickness is about 100 m, whilst on the Lobbia tongue it is about 50 m.

On the opposite slope, the Vedretta della Lobbia and the Vedretta di Fumo are, strictly, connected and together form a summit glacier. An ice divide passing through the Passo della Val di Fumo (2990 m a.s.l.) separates the two diverging parts that flow into the basins of the Sarca-Mincio rivers (Val di Genova) and of the Chiese-Oglio rivers (Val di Fumo) respectively. The Passo della Lobbia Alta is the connection between the Vedretta della Lobbia and the Vedretta del Mandrone. At present the Vedretta della Lobbia has a surface area of 712 hectares and its two fronts flow downhill to an elevation of 2565 m a.s.l. to the north and an elevation of 2720 m a.s.l. to the south.

Comparisons between the situation depicted on recent maps and the one shown in the 1865 map (PAYER, 1865),

Fig. 15. Fluctuations of Lobbia and Mandrone glaciers tongues. 0 refers to the 1864 position for the Lobbia glacier (Payer, 1865) and to 1932 position for the Mandrone Glacier. The curves have been drawn on the basis of data supplied by annual glaciological surveys (continuous line), cartographic and iconographic documents and data from literature (dashed line), after Baroni & Carton, 1996.




permit some observations to be made about the glacier's variation in surface area and thickness. Some of the peaks that are part of the rocky crest that separates the Vedretta della Lobbia from the Vedretta del Mandrone (Cima Lobbia Alta), in 1865 seemed to be completely surrounded by the two neighbouring glaciers. Therefore, it is apparent that from 1864 until the present the Vedretta della Lobbia has reduced on one side, causing a lowering of the glacier's surface of over 200 m on its western edge. Similarly, at Passo della Lobbia Alta the lowering of the glacier's surface can be calculated to be 75 m.

Regarding frontal variations, a fairly continuous data series that starts from the first years of this century is available. On the basis of figures, drawings and photographs, the time/distance curve was calculated. From this curve it is possible to infer that from 1864 to the present the glacier has retreated 1950 m in total. The generalised retreat of the glacier is interrupted by advance stages in the years 1887-1895 (not measured), 1912 (not measured), 1919-1923 (42 m), 1939-1940 (15 m), 1962-1967 (36 m), 1973-1975 (12 m) and 1977-1985 (> 25 m).

From a cartographic comparison between the positions reached by the front in 1864 and 1983, it can be inferred that the retreat was 2030 m.

Using the general topographic map of the Provincia Autonoma di Trento and updating the boundaries of the Vedretta della Lobbia, the altitude of the present snow limit was computed. Using an Aar of 0.67 ± 0.05, the Ela is at 2917 m 17 m. This limit rises to 2944 m +13/-20 m, if the Lobbia and Fumo glaciers are considered as a whole. Using an AAR of 0.6 0.05 (Porter, 1975; Nesje and Dahl, 1991), which is probably a more representative value for these kinds of summit glaciers, a higher rise of the Ela to 2960 m +11/-9 m would be obtained.

In addition, the topography of the glacier during the maximum expansion period, coincident with the Little Ice Age, was reconstructed. Also, the altitude of the related snow limit was calculated. The value obtained for the Vedretta della Lobbia only is 2855 m +21/-38 m, with an Aar of 0.67 ± 0.05. This shows that there has been a rise of about 60 m from the maximum of the Little Ice Age until the present. If this rise were only related to an increase in temperature, it would represent an increase of about 0.37 °C from the Little Ice Age to the present.

*From Passo del Tonale to Pinzolo* - The return trip will take us to Pinzolo in Val Rendena. We will travel along the Val di Sole trough from the Passo del Tonale to the village of Dimaro that follows the previously mentioned Tonale Line. Near Dimaro the Tonale Line crosses the Giudicarie Line, which borders the intrusive massif of the Adamello to the east.

In the first part of the descent from the pass to the village of Fucine, it is possible to observe on the right-hand side a series of valley troughs of different dimensions, at the heads of which other glaciers of the Presanella group are located. In addition, in the main valley bottom, frontal moraine deposits are still evident and they are thought to belong to the Late Glacial phases of the Vedretta della Presanella, the biggest of the whole group. Also, from the left side of the valley glacial tongues used to descend to the main valley floor (Val Strino, Val Verniana, Val Saviana), but the related Late Glacial end moraine systems are located at much higher altitudes because the glaciers that deposited them were facing south.

After having left Val di Sole near Dimaro we will climb a valley to Passo Campo Carlomagno, which is the water divide between the Adige (Noce) and the Po (Sarca-Mincio) basins. From the pass it is possible to look west to the Adamello-Presanella massif, which we have just visited, and to the east to the Brenta group, orientated in a north-south direction. The latter is a spectacular dolomitic-limestone mountain range with a typically «Dolomitic» morphology. A number of small cirque glaciers have carved many depressions along the crests, and rock towers often rise inside these depressions. Very long terraces can also be observed that are due to the selective erosion of the sedimentary lithotypes. Surface and underground karst is also present.

Campo Carlomagno and the nearby little town of Madonna di Campiglio are two well known and very busy tourist centres, especially for winter sports and more than 30 000 people visit the area each season. Madonna di Campiglio was a rich pasture area around 1830 and it became a tourist centre as early as 1872. It was originally built around a fortified hospice run by monks, that was founded before 1188.

The intense human activity in the area covers the many moraine ridges originating from Val Nambino, which are mostly parallel to the main valley and are principally composed of clasts of tonalitic rocks. The traces of the glacier that used to flow in the valley of Campiglio, located at lower elevations, are attributed to würmian retreat phases. Large areas to the west of Madonna di Campiglio are characterised by glacial terraces, which are at times superimposed on each other. Some of these glacial terraces host peat deposits. Moving closer to Pinzolo the glacial deposits are evidently more and more influenced by the interaction between the Adamello and Brenta glaciers; the glacial tills become progressively richer in carbonatic components aside of the tonalitic components.

7. 4th day. Itinerary: Pinzolo - Val di Genova - Pinzolo - Riva del Garda

## HOLOCENE GLACIER VARIATIONS IN VAL DI GENOVA
(C. Baroni & A. Carton)

*From Pinzolo to the top of Val di Genova* - In today's field trip a series of stops at the top of Val di Genova are scheduled, with the aim of analysing and characterising the lateral and frontal moraine deposits of the Lobbia and Mandrone glaciers. The deep valley trough of Val di Genova separates the Adamello massif from the Presanella group, and runs from Carisolo (800 m a.s.l.) for over 20 km into the heart of the Adamello massif and up to the Adamello plateau glacier.

The water captures and the canals that are present along the valley bottom in the first part of the Val di Genova show that the whole Adamello massif is intensely exploited for the production of hydroelectric power. The water exploitation in this area dates back to the early years of this century. The morphology of the massif, particularly in the southern sector, is extremely suited to the construction of hydroelectric power plants.

The first part the valley is set along the contact between the tonalites on the left and the granites on the right. The valley is cut into steep mountain slopes and only one of the many glaciers that surround it is visible towards the top of the valley. The lateral tributaries, especially those to the right, are a series of steep sided cirques, two of which still host glaciers of fair dimensions (Vedrette di Lares and di Folgorida). At the front of these cirques a complex system of moraine arcs is present. However, glacial deposits are basically absent within the Val di Genova almost until the head of the valley is reached. Only around an altitude of 1380 m (Cascina Muta) have some previous authors observed a frontal moraine arc. By contrast, debris accumulations, alluvial fans and alluvial plains filled with sediments are abundant. Numerous avalanche and debris flow deposits are present, especially in the middle high portion of the valley. The valley also has a series of rock steps where the River Sarca cuts spectacular, deep incisions.

In the plain uphill of the Bedole refuge and near to Matarot, the head of the Val di Genova, there are a concentration of extensive glacial deposits. The majority of the deposits are shaped in fairly well preserved ridges, and be-

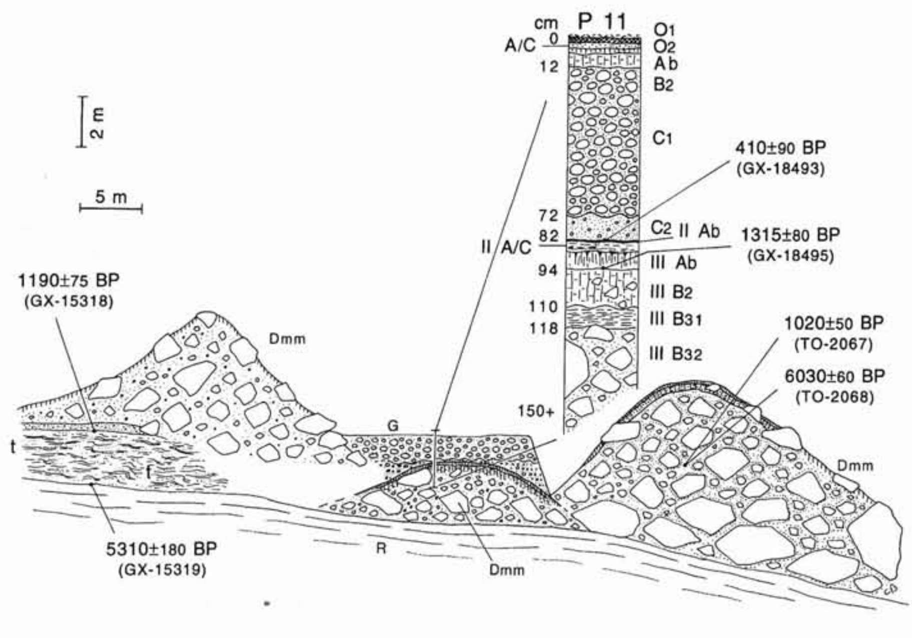



FIG. 16. - Val di Genova, stratigraphic sketch downstream the bridge near Mga Matarot. In evidence the pedological profile and the $^{14}$C dates so far obtained. G= fluvioglacial gravels; Dmm= massive diamicton matrix supported; t= peat and silty peat, deformed; R= bedrock (tonalite). (BARONI & CARTON, 1996).

long to the Lobbie and Adamello (Mandrone's tongue) glaciers.

Within the whole Adamello Group this area is probably the only one in which the glaciers' tongues have recently penetrated into an area of forest. For this reason it was possible to reconstruct the holocenic glacial history using absolute chronology techniques. On the basis of morphological, sedimentological, stratigraphic, lichenometric, and dendrochronological observations, and with the help of $^{14}C$ dating, the principal phases of the geomorphological evolution of the area were reconstructed. The Holocenic maximum extents of the Vedretta della Lobbia and the Vedretta del Mandrone were reconstructed, and some moraines associated with different events during the Little Ice Age were dated (BARONI & CARTON, 1991b, 1996). The most evident traces of a stationary position of the Mandrone and Lobbia Glaciers are preserved uphill of the Lago Nuovo step (or Acquapendente threshold) and uphill of the Malga di Matarot Bassa step, respectively. However, remnants of the Mandrone Glacier's moraines are also preserved downhill the Acquapendente sill, down to the altitude of the start of the chair-lift that goes to the Mandrone Glacier.

*STOP 1 - Holocene moraines of Mandrone Glacier* - The moraine system of the Vedretta della Lobbia is confined to an area uphill of the Malga di Matarot Bassa threshold and it is formed by many frontal moraines, some of which are associated to evident lateral moraines. A latero-frontal moraine to the north of Malga Matarot, located at an altitude of around 1780 m, marks the maximum Holocene expansion of the Vedretta della Lobbia.

Another moraine, located on the left-hand side of the river at a slightly higher elevation, is more recent, having been dated at 1189 ± 75 $^{14}C$ yr B.P. (GX-15318). This dating was made using the top of a set of organic sediments (peat and peaty silt), which have been deformed and buried by the moraine itself fig. 7.1. The bottom of this organic deposit was dated at 5310 ± 180 yr $^{14}C$ BP (GX-15319; BARONI & CARTON, 1992). Another moraine ridge located on the left of the river (no. 4 on the geomorphological map, BARONI & CARTON, 1996) provided two samples of organic material: a segment of soil dated at 6030 ± 60 $^{14}C$ yr B.P. (TO-2068), and a remnant of larch branch dated at 1020 ± 50 $^{14}C$ yr B.P. (TO-2067). It is more likely that this moraine ridge is a right lateral moraine of the Vedretta del Mandrone than a frontal moraine of the Vedretta della Lobbia. Either way, its age is more recent than 1020 ± 50 $^{14}C$ yr B.P. (TO-2067).

A layer of silty peat with vegetation remnants, located in between the two moraines mentioned above, provided a date of 410 ± 90 $^{14}C$ yr B.P. (GX-18493). This is an organic layer set at the top of a soil that formed on a glacial deposit, which in turn is dated as older than 1315 ± 80 $^{14}C$ yr B.P. (GX-18495). This soil has been buried by fluvioglacial coarse gravel, possibly related to the glacial advance phase that determined the position of the moraine behind. It is believed that these coarse deposits represent a depositional event, related to the glacial advance phase that caused the deposition of the moraine ridge that in fig. 16 covers the organic sediments dated between 1190 ± 75 $^{14}C$ yr B.P. and 5310 ± 80 $^{14}C$ yr B.P.. If the date of 410 ± 90 yr BP is calibrated according to STUIVER & REIMER (1993) an age of 1430/1635 A.D. is obtained. This allows a correlation between the advance phase described above with the first phases of the Little Ice Age.

*STOP 2* - Uphill (East) of Malga Matarot (1 on the geomorphological map, BARONI & CARTON, 1996) a push moraine is present (fig. 17). This moraine is locally made up of peat deposits over 2 m thick. A dating performed at the

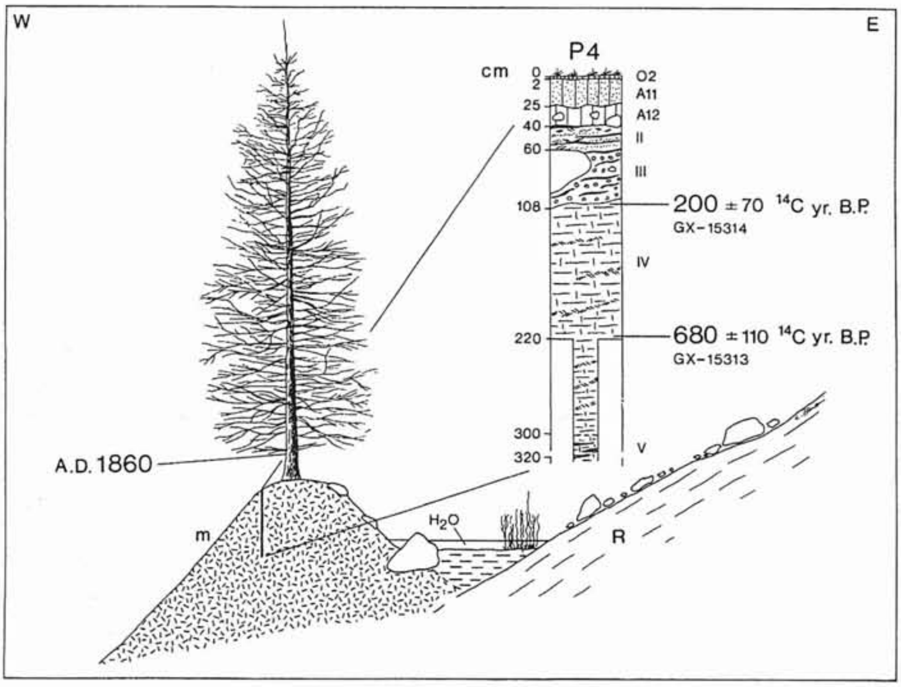



FIG. 17. - Val di Genova, stratigraphic sketch of the push moraine to the East of M.ga Matarot (BARONI & CARTON, 1991b).

top of the peat deposits returned an age of 200 ± 70 $^{14}$C yr B.P. (GX-15314). This age is related to a calibrated date ranging from between 305 years ago to the present. At a depth of 210 cm the same deposit provided a date of 680 ± 110 $^{14}$C yr B.P. (GX-15313). The moraine ridge is soil covered and colonised by a few larch trees more than 100 years old. A 137 years old (1860 A.D.) larch tree from this area was sampled. Considering that the settlement period for larch trees is around 15 years, this would give an age for the moraine of more than 152 years (1845 AD). These data and the calibrated age of the top of the peat layer, give this moraine an age ranging from between 1645 to 1845 A.D. The moraine can therefore be related to the Little Ice Age.

*STOP 3* - A well-marked ridge that is about 5 m high and located 100 m to the south of Malga Matarot Bassa, marks the position of the glacier front in 1864 (PAYER, 1865). This ridge can probably be related to an advance phase that occurred around the middle of the 19th century.

A double ridge formed by massive diamicton is located near a rocky step at about 1900 m a.s.l. This defines the position of the glacier in the first years of this century.

The lateral right and left moraines are well developed. They are locally doubled and are formed by accretion and superimposition moraines. They show sharp, knife-blade like profiles and in the uppermost portion are cut by avalanches and debris flow channels.

*STOP 4* - Until the end of the last century the Mandrone Glacier reached as far as the Acquapendente sill, creating a lobe that was stationary at around an elevation of 1700 m a.s.l.

The most advanced Holocene position of the glacier is shown by a left lateral ridge that develops from the bottom of the Acquapendente threshold up to an elevation of about 1750 m a.s.l. Further evidence of this position is given by a small segment of frontal ridge located near to the start of the cable-car for the Città di Trento refuge (1690 m a.s.l.). Downhill of the left lateral ridge silty peats from fluvioglacial deposits have been sampled, which can be correlated to the ridge itself. The samples have been taken from a natural section (fig. 18, P6), at the base of an artificial structure probably used for making charcoal. These samples provided an age of 770 ± 60 $^{14}$C yr B.P. (GX-15315) that indicates a minimum age for the moraine. Whether or not this moraine ridge represents the maximum expansion position of the Holocenic glacial front in the plain of Malga Venezia is not certain. This is because the area has been deeply modified by fluvioglacial processes that may have hidden or eroded other more advanced moraines.

*STOP 5* - Uphill of the Bedole refuge in the Malga Venezia plain at an elevation of 1675 m, a larch trunk over 12 m long was found (fig. 18, P8). This was uncovered through fluvial erosion on the left side of the River Sarca in 1987, and has been given an age of 3255 ± 140 $^{14}$C yr B.P. (GX-15317). This larch trunk was in a heap of fluvioglacial sediments 6 m thick at a depth of 2.4 m, and was orientated

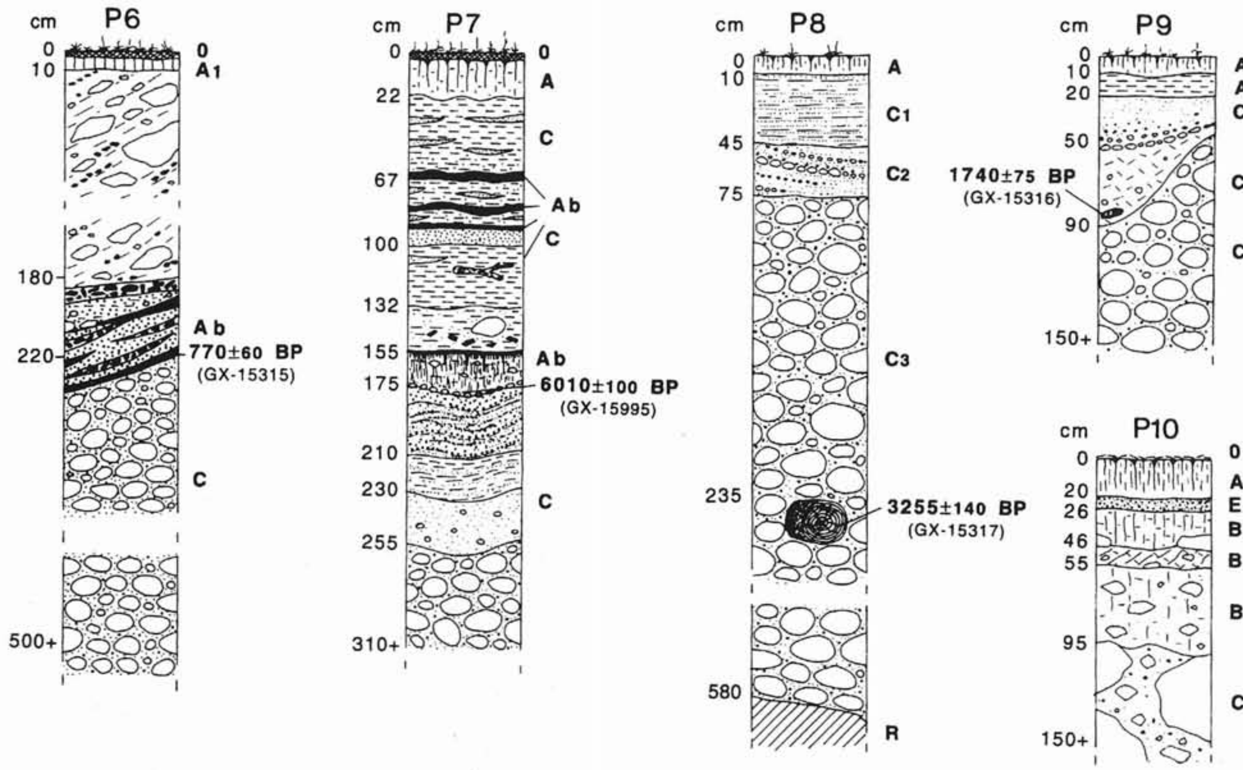



FIG.18. - Val di Genova, detailed stratigraphic profiles (BARONI & CARTON, 1996). In evidence the $^{14}$C dates obtained.

parallel to the water flow. Assuming that the age of this trunk is the same as that of the sediments it was found in would mean that the glacier has never gone further than this location since 3255 ± 140 $^{14}$C yr B.P.. A minimum age for the deposits containing the trunk has been obtained from a piece of wood found at a depth of about 70 cm from ground level in deposits that filled a channel inside the same fluvioglacial deposits (1740 ± 75 $^{14}$C yr B.P.; GX-15326; fig. 18, P9). Close to the Bedole refuge a section has been cut into alluvial deposits (1635 m a.s.l., fig. 18, P7). This showed that at this location fluvial sedimentation was occurring even earlier than 6010 ± 100 $^{14}$C yr B.P. (GX-15995). Later evidence of glacial sedimentation has not been found in this area.

*STOP 6* - Late Glacial moraines and deposits can be found on the *roches moutonnées* that border the Malga Venezia plain to the SE. This shows that here the maximum Holocene expansion is not as advanced. A sequence of peat deposits with a thickness of 1.5 m located between two moraine ridges at around an altitude of 1685 m (marked 6 on the geomorphological map, BARONI & CARTON, 1996) has an age of 7940 ± 90 $^{14}$C yr B.P. (TO-2066) at the bottom. This age suggests that the ridge that dammed the small pond in which the peat was deposited during Late Glacial time. On top of the peat there is a well-developed podsol with a thickness of approximately 1 m (fig. 18, P10).

The position of the Mandrone glacier around the middle of the last century (PAYER, 1965) is well defined by two lateral ridges and a segment of a frontal moraine. These ridges outline a fan shaped glacial flow that reached an elevation of 1700 m. On top of these ridges, which are also colonised by vegetation (larch and fir trees), a soil a few centimetres thick of type A/C has developed. Some of the larch trees have ages ranging from between 1877 and 1880 AD, assuming a settlement time for the larch of about 15 years (BARONI & CARTON, 1991b; BARONI & *alii*, 1993).

A small ridge at the bottom of the Acquapendente threshold at around an elevation of 1870-1920 m defines the position of the front in 1895 (MARSON, 1906). The most recent moraines of the Mandrone Glaciers are located on the Acquapendente step, where small isolated ridges are present. Other recent evidence can be observed at the margins of the present glacier tongue and in the Lago Nuovo depression. Progressively lower levels are evident that are also confirmed by the reducing of lichen diameters (*Rhizocarpon geographicum and Aspicilia cinerea*). Some of the stationary positions of the glacier are underlined by alignments of rock blocks.

The use of lichenometric analyses allows a relative chronological ordering of the different moraines and correlations to be made between different morphological units. The relative chronological curves of the different groups of lichens were constructed using lichens growing on a military construction of World War One as a basis. The curves cover the period between 1864 and 1915. The curves obtained (fig. 19) are very different from one another. The curve for the *Rhizocarpon geographicum* group is lower because of a lower growth rate, which is about 0.5 mm/year

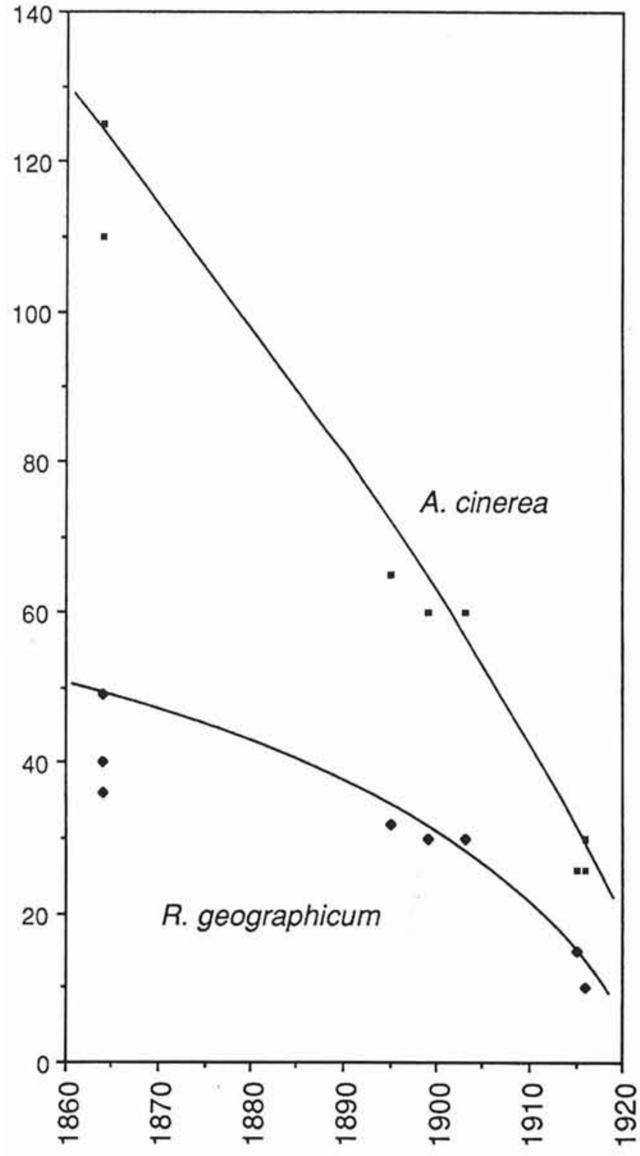



FIG. 19 - Lichenometric (*Rhizocarpon geographicum* and *Aspicilia cinerea*) growth curve for the Upper Val di Genova (BARONI & CARTON, 1991).

for the period considered. The curve for the *Aspicilia cinerea* shows a higher growth rate of about 1.7 mm/year for the same period. The correlation coefficients are quite high, being 0.93 and 0.99 for *Rhizocarpon geographicum* and *Aspicilia cinerea* respectively. As a verification of the curves' reliability, cross-checks were performed between the lichenometric, the dendrochronological and the iconographic data of the neighbouring areas.

Before leaving Val di Genova it is worth admiring the spectacular Nardis waterfalls, which are very popular for ice-climbing in winter. The Val di Nardis is the most important left tributary valley of Val di Genova and descends directly from the southern side of the Presanella peak. The moraines of the last century are quite well preserved, as well as the Late Glacial moraines at lower elevations. The water feeding the waterfalls come almost exclusively from the Western Vedretta del Nardis, which is a mountain type south-facing glacier with a surface area of 170.5 hectares.

## 8. 5th Day. Itinerary: Riva del Garda - Dro - Sarca Valley - S. Lorenzo in Banale - Riva del Garda
(A. Cavallin, M. Marchetti & G. Orombelli)

### 8.1 INTRODUCTION

The study area belongs to the Southern Alps and from the paleogeographical viewpoint it includes the eastern part of the Trento «plateau» and the western «Lombardy basin», whereas tectonically it belongs to the «Giudicarie System».

Rock types ranging from the Paleozoic to the Holocene crop out in this area. The most ancient units are the «Vulcaniti Atesine», overlain by the Permian Val Gardena Sandstones and Bellerophon Formation which are followed by the Mesozoic and Tertiary marine sedimentary units. The Triassic terms crop out as alternances of prevailing carbonate units with marlstones, sandstones and siltites. This sequence is closed at the top by the Dolomia Principale Formation which is widespread all over the area and is followed by a Jurassic-Cretaceous sequence made up of the following terms: shelf-facies rock types including Grey Limestones, St. Vigilio Oolithic Limestones and Misone Limestones, whereas basin-facies rock types are composed of a thick sequence of cherty pelagic carbonate sediments. In the transition facies between shelf and basin, units related to a synsedimentary tectonic phase are found. During the Paleogene the deposition of the «Scaglia», marly limestones, calcareous marls and clays took place. The Neogene units are rock types made up of Miocene calcarenites and greywakes and Pontic regressive conglomerates. The latter crop out only in the southern part of the area, near the shore of Lake Garda.

The study area contains tectonic alignments of regional importance, since the boundary between the «Lombardy basin» and the «Trento porphyric plateau» is here recognisable.

In the Garda region and in the southern Adige Valley, the geomorphological elements, valleys and ridges, are controlled by and parallel to the tectonic lines of the «Giudicarie System», with a NNE-SSW direction. The geological structures are characterised by west-dipping monocline ridges which are bordered to the east by tectonic scarps forming the right slopes of large asymmetrical valleys, sometimes split up by dip-upstream shear surfaces. The geological structure of the southern sector of the «Giudicarie System» is overlain by a Triassic-Eocene carbonate succession divided into NNE-SSW elongated blocks. These blocks dip toward WNW and overlap the main Giudicarie lines. They give rise to monocline morphostructures marked by asymmetric ridges, the western slopes of which correspond to beds, while the eastern ones are tectonic scarps. Therefore, the valleys between ridges correspond to depressions between tectonic scarps and dip slopes. This general morphological situation is very prone to the onset of vast mass movements which affect the slopes of the asymmetrical valleys, in correspondence with the monocline structures and the tectonic scarps.

From the morphological standpoint, the landforms resulting from gravitational processes assume a greater importance than glacial landforms, which deeply affected the Alpine valleys in the past, and which will be illustrated in greater detail in the following days. The large landslides are deeply linked to the marked asymmetry of the valleys, where western slopes are characterised by tectonic scarps as long as 20 km; there are differences in level of up to 500-1000 m and slope gradients of 30 to 50 degrees. Morphological and structural characteristics allow two main types of slope movements to be recognised: 1) rock falls from tectonic scarps which correspond to slopes facing ESE; 2) translational slides along bedding surfaces from slopes facing WNW. In the first case the detachment processes affect prevalently vertical rock bodies, whereas in the second case blocks of thin beds with wide lateral extension are affected. In the accumulation areas it is possible to distinguish heterometric detrital deposits overlapping matrix-prevailing *diamicton*, from rock bodies which in places still preserve some elements of the original structure.

The surface of the landslide deposits is irregularly undulating, with confined depressions and long, narrow rises. By means of air-photo analysis some arcuate, sometimes concentric structures can be recognised. They derive from both modest bank-like rises, small, narrow valleys and alignments of large boulders or fine-textured belts, which are often shown up by the vegetation. These landslide accumulation features are generally interpreted as flow structures (VOIGHT, 1978; PORTER & OROMBELLI, 1981; EISBACHER & CLAGUE, 1984).

Mass movements of this type, i.e. large bulks of mostly dry rock debris caused by the collapse of a slope or cliff moving at high velocity and for a long distance, are classified as rock avalanches (ANGELI & *alii,* 1996). Concerning the genesis of these phenomena, various hypotheses have been suggested.

For these landforms some Authors proposed a glacial origin rather than a gravitational origin (OMBONI, 1875). OMBONI (1878) maintained that rock avalanches were landslide deposits occurring in glacial conditions, whose accumulations must have been distributed by glacial processes. This interpretation cannot be accepted since the widest landslides (Molveno, Drò, Marco) are far more recent than the Late-Glacial Age. Many concordant chronological data tend to ascribe them to the Holocene, between 3000 and 1000 years B.P.

Studies on neotectonics carried out in Italy in the 1980s showed that these large landslides were induced by structural activities. Some of the landslide scarps are, in effect, connected with presumably active structures (CAVALLIN & *alii*, 1989 b). Among the wide tectonic scarps found in this area, it is possible to identify «fault scarp-walls» connected with the neotectonic evolution of the Giudicarie System

(CAVALLIN & *alii*, 1989 a). Morphologically these features can be classified as: a) discontinuous, i.e. made up by a sequence of triangular or trapezoidal facets (e.g. W of Lake Molveno, fig. 20; b) continuous, i.e. with linear development: they are made up of a wide linear wall with a straight upper edge and base (e.g. the western Drò fault scarp-wall in the lower River Sarca valley, fig. 20); c) ribbon-like, i.e. made up of a ribbon-like band, rather even in width, winding along the slope down the lateral small valleys and up the spurs.

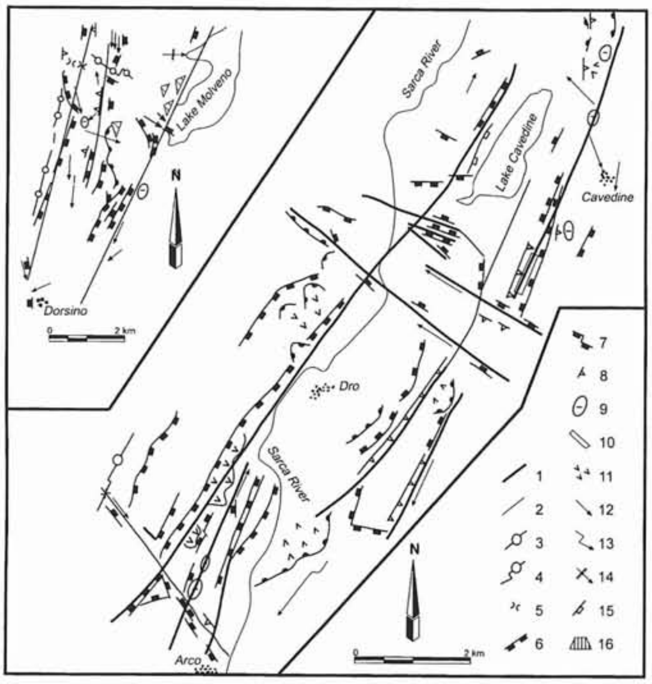



FIG. 20 - Neotectonic sketch of the Molveno and Drò areas. Legend: 1) lineament; 2) fracture; 3) linear ridge; 4) planar discontinuity of ridge; 5) saddle; 6) scarp; 7) diascontinuity of scarp; 8) reverse slope; 9) karst-epigenous and pseudodoline forms; 10) oblonged depression; 11) landslide; 12) gully; 13) double river bend; 14) step in the longitudinal profile of water course; 15) straight coast; 16) triangular and trapezoidal facet (modified after CAVALLIN & MARCHETTI, 1995).

## 8.2 THE ITINERARY

The itinerary continues across the village of Arco, with a view of a medieval stronghold on top of a rugged cliff dominating the valley, and to the village of Drò.

The valley has steep slopes cut in the Mesozoic carbonate units; the valley floor is wide and its margins are sharply connected with the slope toes and are partially covered by shallow belts of slope deposits. Some accumulations deriving from rock falls are at times visible.

The talus fans are often exploited for quarrying activities which are mainly concentrated on the slope deposits rather than on the valley floor, where production activities with higher economic value are mostly located.

Along the valley there are evident traces of the modelling action of the Garda glacier, shown also by the presence of *roches moutonnées*.

In the proximity of the village of Drò, the valley is dammed by the large accumulation of the so-called «marocche». This term is of dialect origin and means rock avalanche deposits. The rock avalanche stretches for a considerable length between the crown and the accumulation zone, covering a surface of about 10 km$^2$, with maximum length and width of about 5 and 3.5 km respectively and a total difference of altitude of 1250 m, between 1400 m at the landslide crown and 150 m at the toe (BARONI & *alii*, 1990).

The main landslide scarp affects a vast portion of the right-hand slope which corresponds to a reverse-fault tectonic scarp. The main scarp is rather segmented and complex and its geometrical characteristics are controlled by discontinuity surfaces, among which large subvertical slickensides are noticeable.

Proceeding in the itinerary, the national road 45-bis is left behind and a path is taken going uphill on the left hand slope of the lower Sarca valley. This leads to a point where there is a good view of the vast mass movement which dammed the valley and gave origin to Lake Cavedine (fig. 21). Along the right slope, to the north, other apparently older main scarps are visible.

The last hairpin bend, just before entering the village of Drena, offers a panoramic view of the large landslide zone of accumulation. Arcuate structures showing the same flow direction may be easily seen. The material comes from the zone of depletion located on the right hand slope and has partially run up the opposite slope for 200 m, up to the altitude of about 350 m, thus forming a reverse slope, confining a depression decametres wide on the left slope. Most of the zone of accumulation lacks continuous vegetation cover (apart from the northernmost portion and some areas SSW of Lake Cavedine, where a thick wood scrub has grown). Belts of shrubby vegetation often grow over concentrations of fine-particle materials.


FIG. 21 - Geomorphological map of the Cavedine valley. Legend: 1) alluvial fan; 2) fluvial deposits; 3) edge of kame terraces; 4) landslide accumulation; 5) ice contact and kame deposits due to the Sarca tongue; 6) ice contact and kame deposits due to the Sarca-Cavedine tongue. (after BOLLETTINARI, in PANIZZA & *alii*, 1996).

The excursion proceeds downhill towards the valley floor in order to observe the zone of accumulation where the internal structures of the landslide deposits are visible. In particular, in these materials reverse graded bedding and karst dissolution phenomena are quite evident. The latter have brought into evidence chert nodules which were truncated during the mass movement.

From this point, the excursion continues up the slope and through the Cavedine valley. This stretch of valley is 400 m higher than the adjacent Sarca valley which is separated by the Gaggio ridge, at a height of about 750-800 m. Along this valley on the SE slope various orders of deposits interpreted as *kame* terraces are found; they are ascribed to the deglaciation phases of the late Würmian glacial stage (fig. 21) (PANIZZA & *alii*, 1996). The most elevated ones are compatible with a single glacial tongue which occupied the whole Sarca valley and the Cavedine gorge, covering also the Gaggio ridge. On the other hand, the deposits located between 700 m and the valley floor are considered as ice contact and *kame* derived from the barrier across the Cavedine gorge formed by the Sarca glacier during its withdrawal (PANIZZA & *alii*, 1996).

Once across the Cavedine valley, the itinerary proceeds northwards as far as the 45-bis national road, near the Toblino Lake. Here there is a panoramic view of a large scarp-fault surface stretching along the eastern scarp of M. Paganella-M. Gazza, continuing southwards on the eastern slope of the M. Casale-Monte Brento - M. Biaina ridge. From this fault surface the previously described Drò landslide was detached (fig. 20). The scarp-fault surface of Mt. Paganella runs just above the base of the spectacular eastern slope of the M. Paganella - M. Gazza ridge and is shaped like a steep winding ribbon-like belt several metres high and some hundred metres long (CAVALLIN & *alii*, 1989 a).

Once again the excursion continues by leaving the 45-bis national road and going up the upper Sarca valley as far as Villa Banale, where a diversion on the left-hand slope of the valley is followed, towards Lake Molveno.

Similarly to the previously illustrated Lake Cavedine, also Lake Molveno was formed by a landslide barrier. Along the road heading to Lake Molveno, the main morphostructural features are well in evidence (fig. 20) (CAVALLIN & *alii*, 1989 c). In particular, the valley asymmetry is observable, with an extremely steep western slope affected by scarp-fault-surfaces, whereas the eastern one descends gradually along a structural slope, with slope gradient corresponding to the attitude of the stratified formations. The Molveno landslide gave origin to the lake located north of it by damming the valley floor with its displaced material. This landslide occupies a total surface of about 6 $km^2$, stretching for a maximum length of about 4 km, a width of 2.7 km, and a difference of altitude of about 1000 m, between 1600 m and 550 m (BARONI & *alii*, 1990). The material detached from the right-hand side of the valley which corresponds to a tectonic scarp which is made up of the Mesozoic carbonate sequence overlying Eocene limestones. Some hundreds of metres upstream of the landslide crown, near Soran (not reachable), the slope is displaced by a reverse slope scarp few metres high and 1 km long. Landslide crowns are present also on the left-hand side of the valley within a monocline structure. Nevertheless, their smaller dimensions make them much less evident from the geomorphological standpoint. The large landslide crown has a square contour and seems to be controlled by a NNE-SSW fracture system, in turn intersected by NW-SE oriented fractures. At its highest extremity N-S oriented open cracks up to some decimetres in width are present; they affect both the bedrock and the herbaceous soil cover and give rise to a step-like morphology with elongated depressions (cracks and extension trenches).

Beyond the St. Lorenzo in Banale tunnels a complete view of the accumulated landslide materials is offered. This deposit covers an area over 4 $km^2$ with thickness ranging between 550 and 1100 m. It is made up of blocks overlying massive diamicton deposits with finer matrix, as observed along the fluvial erosion segments. From a vertical viewpoint it is possible to observe arcuate concentric structures within the deposit showing different orientations. They seem to be related to different areas of provenance or could be from complex movements of the displaced materials during their deposition.

The surface of the zone of accumulation is covered by a mixed deciduous wood which has grown on a moderately developed soil. The large limestone boulders emerging from the vegetation show widespread and moderately developed karst microlandforms (PERNA & SAURO, 1979). The development of the soil and the karst microlandforms is in agreement with a protohistoric age (2908 ± 153 $^{14}C$ yr B.P.) of the landslide derived from the $^{14}C$ dating carried out by MARCHESONI (1958). A recent survey of the area has pointed out the presence of superficial karst landforms even below the normal level of the lake which is located at the altitude of 775 m (BORSATO, 1991).

From Lake Molveno it is possible to observe the Molveno bluff along the western shore of the lake itself. This feature is composed of a series of triangular facets and small slope faces (fig. 20), degraded by several rock falls which have emphasized the presence of slickensides in places preserving subhorizontal friction *striae*. On the whole, it is a single large fault scarp retrograded following the rock falls which took place during the Holocene. The degree of conservation of this scarp increases from north to south. In its southernmost portion the scarp forms a 10-15 m high subvertical and straight lacustrine shore.

The day-trip will end with the return to Riva del Garda followed by a guided tour of Lake Ledro and the lake-dwelling Museum of Ledro, which is a Section of the Trento Museum of Natural History.

## 9. 6th Day. Itinerary: Riva del Garda - Lake Garda - Riva del Garda - Lavini di Marco - Riva del Garda

### THE PALEOLEVELS OF LAKE GARDA AND THE «LAVINI DI MARCO» LANDSLIDE
(C. Baroni & M. Marchetti)

Lake Garda is the largest Italian lake, covering a surface of about 370 $km^2$ and a perimeter of 155 km. It is located in correspondence with a deep depression formed along an important tectonic alignment, called the Ballino-Garda line (CASTELLARIN, 1981), which was intensely reactivated after the Mesozoic. Between the Ballino line and the parallel Giudicarie line to the west, an area with extensive overthrusts of competent carbonate formations on top of Jurassic and Cretaceous rock types is observed (CASTELLARIN, 1981).

East of Lake Garda Mesozoic marine sedimentary rocks crop out in the Mt. Baldo and Mt. Pastel areas. From a structural viewpoint, this sector consists of a sequence of ridges extending from N to S. The ridges are westward or WSW inclined and are separated by faults with a «Giudicarie System» direction showing morphological features mainly in well-defined scarps (CAVALLIN & *alii*, 1989 a). The structural picture is completed by the asymmetrical anticline of M. Baldo and the syncline of Ferrara di M. Baldo which are joined by means of a thrust.

The lake stretches from the margin of the Po Plain into the Pre-Alpine reliefs, separating two different domains: the Lombardy basin to the west and the Trento high ridge to the east. Regarding the origin of Lake Garda, as for the other great Pre-Alpine lakes, various theories have been suggested. All the main lakes lying south of the Alps (Maggiore, Lugano, Como, Iseo and Garda) are cryptodepressions. For example, Lake Garda's deepest point (minus 346 m) is -281 m below sea level. The most commonly accepted theory is that the depressions occupied at present by the Pre-Alpine southern lakes are due to canyons which were deeply eroded by the watercourses during the Messinian evaporitic drawdown. The erosional phase can be dated at the late Miocene (Messinian), when the Mediterranean basin was isolated from the Atlantic Ocean (BINI & *alii*, 1978).

As for Lake Garda, the Messinian fluvial erosion took place in a pre-existing valley which was tectonically controlled by structures belonging to the Giudicarie System.

During the following glacial phases the Garda valley was affected by the action of glaciers, with consequent glacial remodelling. The Pre-Alpine margin along the Garda Valley was reached several times by the Alpine glaciers at least since the late Early Pleistocene. These glacial processes caused the deposition of a number of arcuate hills at the front of the lake; they make up the moraine amphitheatre of Lake Garda (fig. 22). During the last maximum glacial advancement (stage 2), the lake was filled up by a glacier which reached a thickness of 1000 m near Tremosine. The limit of Würm moraines is over 1000 m a.s.l. in the lake's upper part, about 650 m in Tremosine, 400 m at Gardone Riviera, 260 m in Salò and about 200 m circa in Solferino (minimum level), at the southernmost boundary of the amphitheatre.

Many Authors have studied this amphitheatre starting from the second half of the $19^{th}$ century (PAGLIA, 1861; SACCO, 1894, 1896; NICOLIS, 1899; PENCK & BRÜCKNER, 1909; COZZAGLIO, 1934; VENZO, 1957, 1961, 1965; FRAENZLE, 1965, 1969; HABBE, 1969; CHARDON, 1975; HENTKE, 1983; CREMASCHI, 1987). It is known that the chronostratigraphic attribution of the moraines of Lake Garda is based on classic Alpine stratigraphy, even if there is no unanimous agreement among the authors. Following CREMASCHI (1987), the moraines nearest to the lake are the result of Late Pleistocene maximum, whereas those at the western margin (Carpenedolo-Faita) belong to the Middle Pleistocene and the outmost ones (Ciliverghe) to the Early Pleistocene.

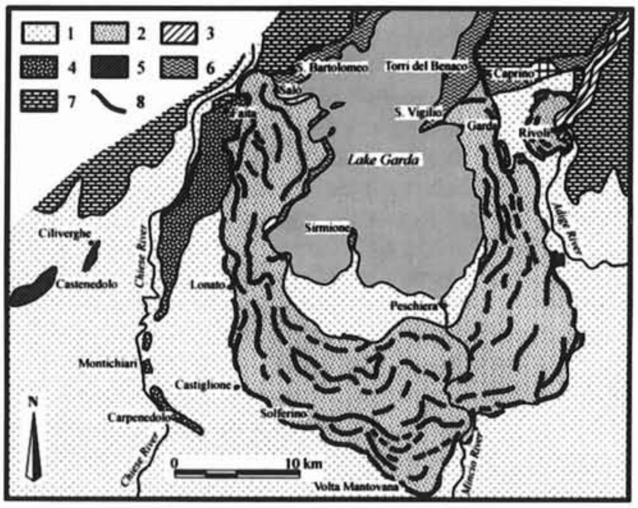



FIG. 22 - The Garda morain system. Legend: 1) fluvioglacial deposits generally named: «Po Plain Main Level» (Upper Pleistocene); 2) Upper Pleistocene moraines; 3) ancient lake in the Adige valley (Upper Pleistocene); 4) Middle Pleistocene moraines of the Carpenedolo phase; 5) Early Pleistocene moraines and related fluvioglacial terraces of the Castenedolo phase; 6) Pre-Quaternary rocks covered by glaciers; 7) Pre-Quaternary rocks never covered by glaciers; 8) Morain ridge (modified after PENCK & BRÜCKNER, 1909 and CREMASCHI, 1987).

The evolution of the Garda amphitheatre has been strongly conditioned by the tectonic events of the area. The Giudicarie line separates two regions with deeply different tectonic styles. In fact, its Pliocene and Pleistocene development is clearly conditioned by a more pronounced uplift of the western area (Brescia) with respect to the eastern one (Verona). Whilst the latter behaved like a plateau during the Pleistocene and was affected mainly by pedogenetic processes (ZANFERRARI, 1982), thick gravelly deposits were formed at the toes of the slopes at the Brescia margin, during the Pliocene and Early Pleistocene. The Quaternary uplift of S. Bartolomeo di Salò (W of the lake) is at least 600 m. The evolution of the amphitheatre has therefore undergone a development which in the most recent phases caused a progressive eastward rotation of the glacial tongues from the Early Pleistocene to the Late Pleistocene (fig. 23) (CREMASCHI, 1987).

### 9.1 THE PALEOLEVELS OF LAKE GARDA (C. Baroni)

The average Lake Garda level is around 0.9 m over hydrometric zero (64.027 m a.s.l. at Peschiera). The ordinary high-water level (month of April) is up to 1.3 m over hy-

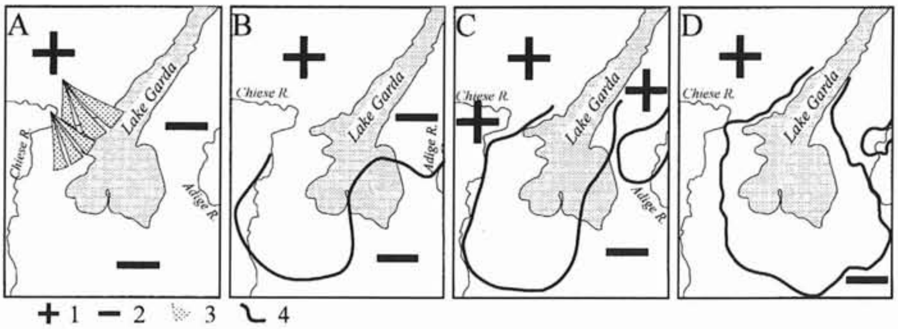



FIG. 23 - Evolution stages of the Garda moraine amphitheatre from Late Tertiary - Early Pleistocene to Upper Pleistocene. Legend: 1) uplifting area; 2) subsident area; 3) alluvial fan; 4) outer limit of frontal moraine; A) Preglacial period (Late Tertiary or Early Pleistocene); B) Castenedolo glacial stage; C) Carpenedolo-Montichiari glacial stage; D) Solferino glacial stage (modified after CREMASCHI, 1987).

drometric zero, whereas the low-water level goes down as far as 0.3 m. The seasonal fluctuation is therefore around 1 m, with maximum values much inferior to those of the other Italian Pre-alpine lakes (the assessment based on historical sequences gives values around 2 m). The volume of water stored within the lake is about 50,000 million m$^3$, with an average permanence time of about 28 years.

The Garda basin is now regulated by the dam across the River Mincio near Peschiera, with a maximum springtime limit of 140 cm in April (exceptionally 175 cm), an autumn limit around 80 cm and a minimum absolute limit at 15 cm over hydrometric zero (DA CASTO, 1986). Along the shores of Lake Garda, morphological and sedimentological evidence directly or indirectly linked to ancient lake levels was found (Holocene - Late Glacial). Many of the elements described are observed along the cliffs which to a large extent are still active; others were observed and described during digging works or were studied during archeological research and excavations.

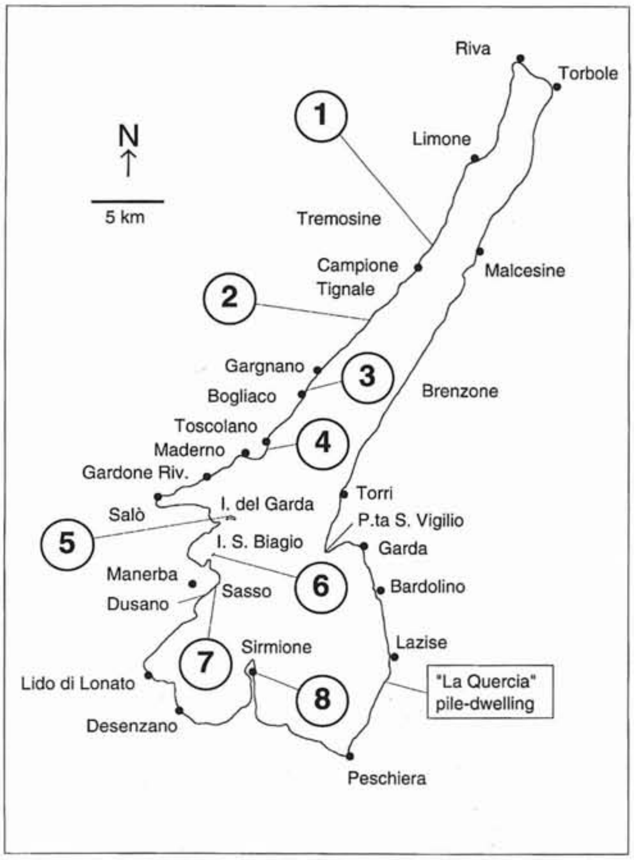



FIG. 24 - Lake Garda and location of the visited sites.

After the withdrawal of the Pleistocene glaciers, between the Lateglacial and 6000 years B.P., the lake level progressively diminished. Some kinds of morphological evidence (i.e. the apices of lacustrine deltas on the western shore and other evidence identified on the high cliffs of the Manerba-Sasso area) may witness the past existence of very high lake levels, over 50 m above the present level. Reliable findings are identifiable at various elevation between the height of 80 m circa a.s.l. (+15 m above the present lake) and the present lake level.

AMS $^{14}$C dating obtained from single fresh-water Gastropods shells, associated with relict shorelines suspended between 4.5 and 1.6 m above the present level (fig. 25), has shown ages ranging between 10,070 ± 70 (TO-4136) and 6140 ± 60 (TO- 4766) $^{14}$C yr B.P. (uncorrected ages). The age of the highest levels can therefore be dated to the Late Glacial (15,000?/10,000 years B.P.).

In the lower Neolithic (6000 B.P. ca, cal age), the lake level was at an elevation lower than 68 m (+3 m above the present level), as witnessed in Sasso di Manerba, where a settlement of fishermen was identified (BARFIELD 1978 and 1981) above the level of lacustrine beaches, dated at 10,290 ± 80 $^{14}$C yr B.P., TO-4902 (BARONI, ined.; fig.28). Other archeological horizons found at the top of these deposits and several prehistoric perilacustrine sites show that after the Neolithic the lake level has always remained below this altitude, since its variations have always been confined around the present level.

Fluctuations from some decimetres up to 1 m are witnessed during the ancient and middle-recent Bronze Age (4100-3500 B.P., cal age). During this period, Lake Garda underwent several variations of level. Settlement stages of pile-dwellings (Lazise, San Felice, Gabbiano di Manerba, Moniga, Cisano, Lazise, etc.) are recognizable by beach and peat deposits which are located below the present level and witness lower-than-present levels. In particular, at Lazise, the underwater excavation of «la Quercia» pile-dwelling revealed various phases of occupation and abandonment of the prehistoric site which are clearly related to

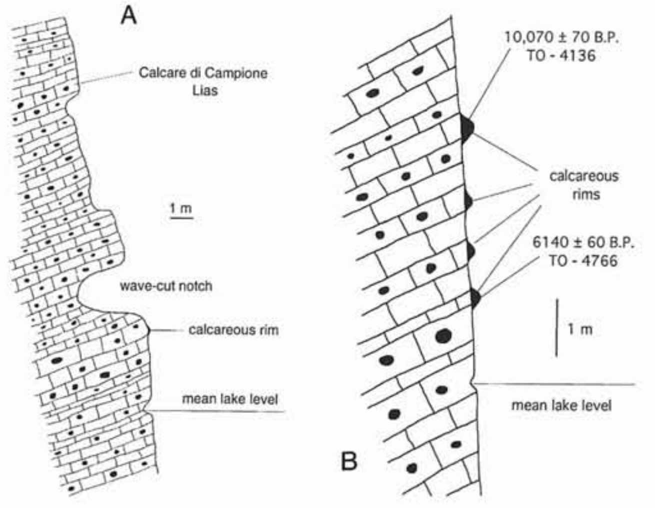



FIG. 25 - 1) North of Campione (Tremosine area): present wave-breaker notch and notch suspended over 5 m, eroded in the cherty limestones of the Campione Formation (Lias). 2) Schematic section with calcareous rims placed between 1.5 m to 4.5 m above the present lake level. these features develop orizzontally on the lacustrine cliff between Gargnano and Tremosine.

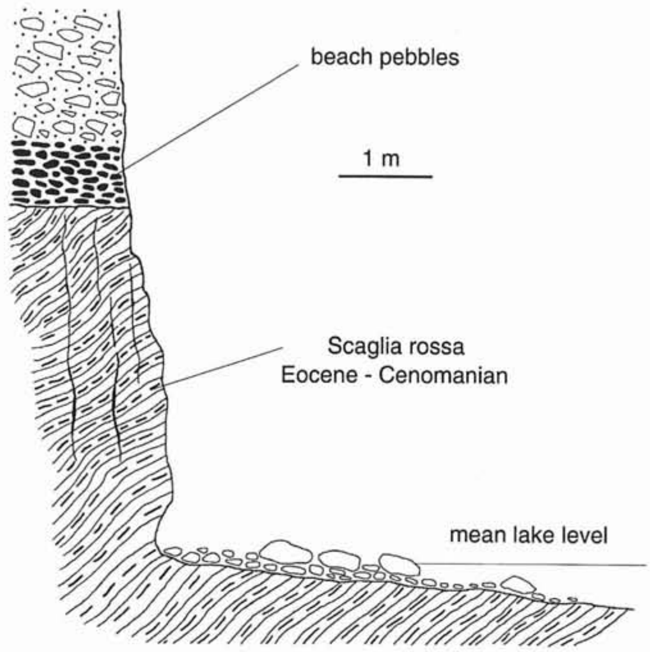



FIG. 26 - Toscolano, near San Giorgio: present abrasion platform and elevated beach pebbles covered by slope deposits; the bedrock is made up of «Scaglia Rossa».

lowerings and subsequent risings of the lake level (ASPES & *alii*, in press). The $^{14}C$ dating obtained from the relict shorelines and the elevated beaches, permits the identification of a relative uplift of the shore north of Salò of about 1 m compared with the Manerba area. This relative rise took place between 10,000 years B.P. (uncorrected age) and the present. If the reservoir effect of Lake Garda is taken into account, the above assessed $^{14}C$ age becomes about 9000 years B.P.; this age, calibrated according to STUIVER & REIMER (1993), would actually correspond to 8055/7845 cal yr B.P.. The uplift of the western Garda coast (North of Salò), with respect to the Manerba area, is therefore assessed at 12.5 cm/1000 years. The rising of the northern region, compared with the lake's lower portion, seems to be even more accentuated south of the Sirmione-Garda fault. It is not yet clear whether the uplift is to be related exclusively with areal movements or ascribed to activity phases of the faults situated on the floor of the lacustrine basin. It is also possible that the relative uplift be the result of a combination of the two processes. The main geomorphological and sedimentological features characterising the ancient shorelines and the present coast are here described.

*Lake cliffs* - Most of Lake Garda's rocky shores are subvertical and show large stretches of dead cliffs which preserve traces of ancient lake levels, represented by suspended abrasion platform, beaches above lake level and wave-cut notches. Active cliffs are prevalently found in the lake's upper portion, although they are found also in rocky outcrops of the middle-lower lake. In particular, the Sirmione cliff (near the Roman villa known as «Catullo's Grottoes») is actively eroded by wave breakers, to the extent that remedial measures such as reinforcement walls have been constructed in order to mitigate the coastline withdrawal and preserve this important archeological area. The cliffs of the lower lake are linked to abrasion platforms of varying extension (up to 100 m and over), whereas in the upper lake the subvertical rocky coasts dip continuously below the lake surface; only in places are incipient abrasion platforms to be found.

*Wave-cut notches* - The present wave-cut notch is relatively widespread along the rocky coasts. The depth, height and planimetric development of the present notch are conditioned by the strata attitude and the amount of wave motion (which, in turn, depends on the prevalent wind directions and dominant currents). Dimensions vary from some decimetres to over 2 m. On the Liassic limestones of the upper Lake Garda the notch is more evident and better typified, although its evolution takes place also at the expense of the bio-calcarenites and the limestones of the Manerba Formation (Early Miocene-Oligocene). The notch which has developed in the «Scaglia rossa» formation (Eocene-Cenomanian) shows less morphological evidence.

Suspended notches, with respect to the present lake level, are found near the 97 km point of the western Gardesana national road (+5 m above the present level; fig. 25) and in the Manerba area, where they are visible up to several dozens of metres above the present lake level (fig. 27).

*Abrasion platforms* - These are the maximum expression of the erosion action exerted by wave motion and are prevalently found in the lake's western sector, which is particularly affected by dominant winds. Present abrasion platforms, which come to the surface during low-water periods, are typical of the main islands and peninsulas. The most spectacular one borders the Sirmione peninsula, whe-

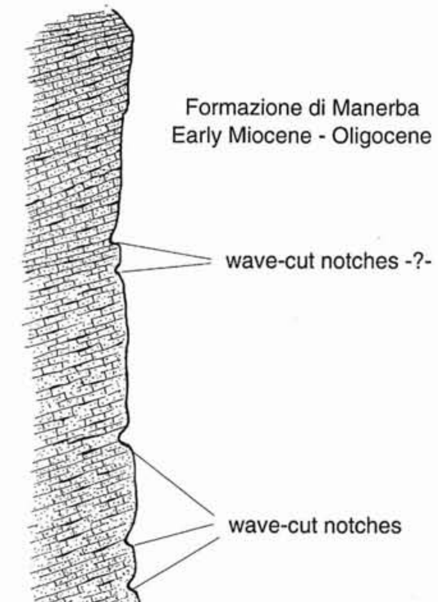



FIG. 27 - Manerba del Garda, loc. Sasso. Cliff with evidence of relict wave-cut notches.

re it is modelled in the Scaglia Rossa, stretching for over 100 m. Worthy of note are also the platforms of Punta Belvedere and Sasso di Manerba. Discontinuous strips and underdeveloped abrasion platforms are found also along the upper lake's shores.

Suspended abrasion platforms were identified in the Manerba area. The most evident and evoluted of which is located at about 3 m above the present lake level, is found on the island of San Biagio. Abrasion platforms at 8 and 12 m below the present lake level were identified at Punta San Vigilio.

*Lacustrine terraces* - In all the area of the lower lake, from Moniga to Bardolino, the coast shows various levels of terraces up to 15 m above the present lake level, which have been modelled in glacial and fluvioglacial deposits. They have been greatly altered by man's activities which have modified to various extents the original morphology. Among the most relevant lacustrine terraces those of Maguzzano-Lido di Lonato and Pieve di Manerba should be mentioned. Above these lacustrine terraces, levels of Holocene beaches were locally identified.

*Dissolution landforms* - Lacustrine *lapiez* of various nature and dimensions are found quite frequently on the carbonate rocks along the coast. In particular, some corrosion pits (*Kamenitza*) should be noticed: their dimensions range between 10 and 80 cm and they evolve on the abrasion platforms. Rounded cobbles are sometimes found inside them.

*Other erosion landforms* - Along the coast also other erosion forms are found, such as little natural bridges (extending for some metres) and small caves. The latter were also observed in correspondence with ancient levels near Rocca di Manerba and in the upper lake.

*Calcareous rims* - In some stretches of the rocky coasts, horizontally arranged concretionally levels are present which are overhanging the lake level. They are made up of algal stromatoliths and other organisms with shells of freshwater *Gasteropoda*. They are among the best indicators of lacustrine paleolevels and were described in marine environment as calcareous rims (PIRAZZOLI, 1991) or organic ledges or «corniche» (FAIRBRIDGE, 1968). In various cases the rock below the incrustation shows cell-like cavities and lacustrine *lapiez*, whereas the upper portion is undisturbed. The border between the two levels is outlined by a thin (some cm) relief bed on the cliff face. These features are either isolated or grouped and are located from 1 to 5.7 m above the present lake level (fig. 25). They show the upper limit of the zone which is regularly wet by the lake. $^{14}C$ AMS analyses carried out on single shells of fresh-water *Gasteropoda* (*Bithynia tentaculata* and *Theodoxus fluviatilis*) found within these concretion levels showed ages ranging from 10,070 ± 70 (TO-4136) to 6140 ± 60 (TO- 4766) $^{14}C$ yr B.P. (uncorrected ages).

*Beaches above lake level* - Remnants of beaches above lake level made up of rounded and embricated pebbles are found in various places. From the lithological viewpoint the composition of the beaches is conditioned by the nature of the bedrock. Beaches above lake level were identified at altitudes of 0.5 to 15 m circa over the present lake level and are related to separate lacustrine beds. The most evident ones crop out at Porto di Dusano and near San Giorgio di Toscolano (fig. 26). Some $^{14}C$ datings were obtained from fresh-water *Gasteropoda* (*Bithynia tentaculata* and *Theodoxus fluviatilis*).

*Lacustrine deltas and suspended fans* - Along the lake shore, lacustrine gravel deposits turning to deltaic deposits in proximity with the main watercourses are found. The great changes along the coast, induced by man's activities, have nearly completely obliterated natural deposits, whose characteristics may be well observed only occasionally, when artificial excavations are carried out. Besides the River Sarca delta, which is the tributary river of Lake Garda near Torbole, numerous active deltas deposit their gravelly-sandy sediments in the waters of Lake Garda. Along the western coast, north of Salò, these deltas are fed by streams running out of suspended valleys which have been cut into deep and spectacular gorges. The main deltas are found in Campione, Toscolano Maderno, Barbarano, Salò and Manerba, but many other fan-like deltas cover the steep rocky slopes of the upper lake coast, whereas several small deltas dismantle the innermost deposits of the moraine amphitheatre around the lake's lower portion.

The deltas of the watercourses are terraced in the upper part, where suspended deltaic deposits made up of gravels, sandy gravels and sands are found. The most evident forms were identified between Salò and Gargnano. The lowering of the lake level during the Lateglacial and the Lower Holocene induced an erosional stage which cut converging scarps, thus leading to the formation of new deltas. The suspended deltas show evident sedimentary structures, such as foreset beds, which are recognisable inside artificial excavations up to at least 30 m above the present lake level, whereas the apices of the suspended deltas are found as high as 100-120 m of elevation.

Bedded debris fans, eroded here and there by the cliff's withdrawal, were recognised along the western coast of the upper lake.

*STOP 1 - Wave-cut notch* - Along the lake's western coast, at the 97.150 km point along the western Gardesana national road, various remnants of shorelines are visible on a rocky spur made up of the cherty limestones of the Campione Formation (Lias). In particular, the wave-cut notch located at over 5 m above the present lake level should be noticed (fig. 25). Under the notch, a small calcareous rim made up of calcium carbonate concretion connected with a lower lake level is visible.

*STOP 2- Calcareous rims* - Between km 96 and km 89.5 of the western Gardesana national road, several levels of calcareous rims are visible; they are made up of calcium carbonate on the calcareous bedrock of various Mesozoic units. These elements, which developed above the lake shoreline, developed horizontally and stand out with respect to bedding. They are located at various altitudes ranging from about 1 to 5.7 m above the present lake level (fig. 25).

*STOP 3 - Delta deposit* - Near the little harbour of Bogliaco, a deltaic deposit cropping out at some metres above the present lake level was recently discovered during works for the extension of the harbour. Foreset bed-type structures are easily visible in rounded and selected gravelly deposits dipping towards the lake at a high angle.

*STOP 4 - The Toscolano Maderno delta* - The Toscolano Torrent cuts out the deep gorge of the Cartiere valley and flows into the lake after cutting through the last outcrop of Scaglia Rossa covered by glacial deposits. The large and impressive fan of the lacustrine delta stretches into the lake for over 1.5 km. Two converging symmetrical scarps border the watercourse which today feeds a centrally placed delta, although its profile tends slightly to SSW.

*STOP 5 - Present abrasion platform* - Near Isola del Garda, a large abrasion platform, modelled in the Manerba Formation bio-calcarenites (Early Miocene-Oligocene), is found. Along the eastern slope of the island two lacustrine terraces located at 1.7 and 4.7 m above the present lake level are observable. Similar platforms are also found at the next stops in Manerba and Sirmione.

*STOP 6 - Relict abrasion platforms* - The San Biagio Island is the morphological relict of two abrasion platforms located respectively at 3.35 and 7.8 m above the present lake level. The higher one is smaller and is surrounded by the lower one. The latter is also correlated to a similar platform located at Punta Belvedere peninsula. A large active platform surrounds the island. The directions of elongation of the island and of its active platform are parallel to the main tectonic features of the area and correspond to overthrusts affecting the Oligo-Miocene calcarenites of the bedrock.

*STOP 7 - Cliffs and evidence of ancient lake levels* - Between Porto di Dusano and Sasso there is a cliff, still active in places, modelled in the calcarenites of the Manerba Formation (figs. 27 and 28). At its base a relatively wide and continuous abrasion platform is found. Remnants of abrasion platforms above lake level locally crop out at 2 to 3.5 m above the present lake level; they are in places associated with small erosion caves and lacustrine gravel. At the top of one of these levels of lacustrine gravel, traces of a settlement of fishermen of the Early Neolithic (V millennium B.C., BARFIELD, 1978 and 1981) were found. These findings show a minimum age for the lake level which at this time was placed at about 3.5 m above the present one. A new $^{14}C$ AMS dating obtained from the above-described gravel (10,290 ± 80 $^{14}C$ yr B.P., TO-4902) allows this beach to be correlated to a calcareous rim found at 4.5 m above the present lake level in the Tremosine area. Also a relative uplift of the area north of Salò, compared with the southern lake portion, has been recognised. This uplift probably took place between about 8000 years B.P. (cal age) and the present.

A series of presumed notches is located some tens of metres above the present lake level and characterises the northern part of the rocky slab near Sasso. The altitude of some of these coastlines shows a correlation with the apices of the suspended deltas of the western shore.

*STOP 8 - Abrasion platform and cliff*

A wide abrasion platform modelled within the Scaglia Rossa Formation (Eocene-Cenomanian) borders the northern portion of the Sirmione peninsula and, on the eastern slope, is linked to an active cliff. Walls and protection works constructed for the conservation of the Roman villa's archeological area hinder the cliff's withdrawal. Remnants of an overhanging abrasion platform (+3.7 m) are identifiable

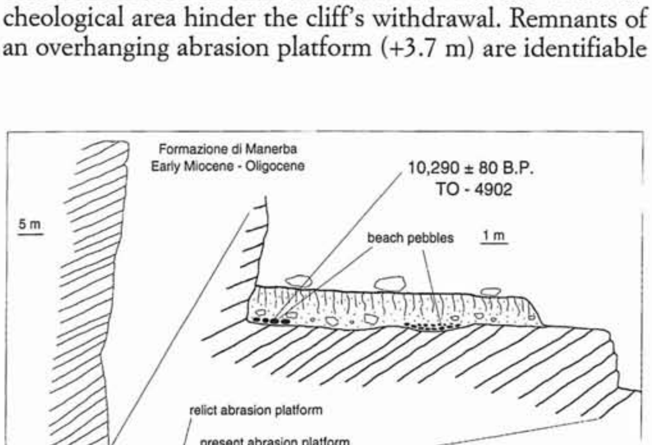



FIG. 28 - Manerba del Garda, loc. Sasso, archeological site (Calcolithic and Early Neolithic). A settlement of Early Neolithic fishermen was find above the level of beach pebbles resting on a relict abrasion platform. The Ams date in evidence was obtained from *Bithynia tentaculata* shells colected within the lacustrine gravel.

in places. Elsewhere along the peninsula, remnants of paleobeaches are also found, which are located some 3.3 to 4 m above the present lake level.

### 9.2 THE «LAVINI DI MARCO» LANDSLIDE (M. Marchetti)

Once arrived on the northern shore of Lake Garda, the route continues eastward along the national road 240 to the Adige Valley. At the San Giovanni di Loppio pass the road goes through the deposits of a large rock slide displaced from the southern valley slope, from a height of 500 m on Mt. Altissimo di Nago; this slide affected a surface of less than 1 $km^2$. In the Adige Valley, at the road junction of Rovereto Sud with the A22 Brenner motorway, the vast, impressive complex of «Lavini di Marco» is visible.

The name «Lavini di Marco» refers to an extensive landslide accumulation with various minor gravitational deposits. This landslide has been known for a very long time and its notoriety is mainly due to the quotation made by Dante Alighieri (the Father of the Italian Language) in his «Divina Commedia» (Inferno, XII, 4-9):

*Qual'è quella ruina che nel fianco*
*di qua da Trento l'Adice percosse,*
(6) *o per tremoto o per sostegno manco,*
*che da cima del monte, onde si mosse,*
*al piano è sí la roccia discoscesa,*
(9) *ch'alcuna via darebbe a chi sù fosse*

*As on Adige's flank this side of Trent*
*an earthquake or a subsidence of ground*
(6) *has wrought such devastation that the rocks,*
*which tumbled from the summit to the plain,*
*have made it possible to scramble down,*
(9) *such was the path descending that ravine*
(English translation by Kenneth MacKenzie - The Folio Society, London, 1979)

The main landslide occupies a surface of about 6 $km^2$, with maximum length of about 5 km, maximum width of 1800 m circa, and difference of altitude of 1030 m (BARONI & *alii*, 1990). The zone of detachment extends over a large portion of the left slope; it coincides with bedding planes and affects a modest thickness of the calcareous sequence (up to some tens of metres). Within the zone of accumulation it is possible to observe quite easily two different lobes. The major lobe shows a fan-like expansion on the valley floor for a distance of over 1 km, it is mostly covered by spontaneous vegetation and is characterised by several depressions on the flat, cultivated portion of the plain. The lesser lobe lies on the bulging part of the slope and is made up of large rock blocks and is mostly lacking in any vegetation cover. These two lobes could either have been generated in different times or be the outcome of a single event during which a resistant rocky ridge located in the failure surface (Costa Stenda) could have halved the displaced material during its movement (BARONI & *alii*, 1990).

On the basis of historical evidence, the Mori landslide (i.e. «Lavini di Marco») was ascribed by several Authors to the early Middle Ages (SACCO, 1940; FUGANTI, 1969; GORFER, 1975; EISBACHER & CLAGUE, 1984; OROMBELLI & SAURO, 1989). Nevertheless, this attribution is not sufficiently documented. The corrosion karst microlandforms show a degree of development comparable and in any case not superior to those of the Drò landslide.

The visit to the landslide will start with a panoramic view from the bottom of the right-hand slope of the Adige Valley. From this position both the detachment scarp and the accumulation zone are quite visible. Bedding surfaces which were stripped of the vegetation cover by the slide of the displaced material are recognisable. These surfaces which completely lack any pedogenetic cover are called «laste». The bedding surfaces are often subdivided by a network of metric patterns forming a mosaic-like floor of large polygonal slabs. These slabs are juxtaposed and sometimes partially superimposed along shear surfaces which are oblique with respect to bedding, with the upper slab partially overlapping the lower one. Fold structures affecting the slide surfaces are considered indications of the mechanical behaviour of the most superficial layers which are characterised by calcareous-marly intercalations. FUGANTI (1969) maintains that slow differential slide movements are still taking place in the uppermost levels, with consequent opening of extension fractures, slow descent of dismembered slabs, production of debris at the margins of each slab and formation of small anticline-shaped fold structures.

The itinerary continues with the visit to the rock avalanche deposits which are made up of large sharp-edged blocks and boulders often showing a parallel structure and totally without any pedogenetic or vegetational cover. The blocks which are in some cases hanging or in precarious equilibrium, lean on each other corner to corner, forming wide cavities. Among the lithological rock types, light-grey limestones and greenish and pinkish marly limestones prevail.

The trip will end with the return to Riva del Garda.

## 10. 7th day. Itinerary: Riva del Garda - Garda - Bardolino - Rocca di Garda - Affi - Mantova - Bologna

### RIVOLI MORAINIC AMPHITHEATRE (M. Marchetti)

The morning of August 28th will be dedicated to the excursion across the Garda moraine amphitheatre and to a description of the Rivoli Veronese amphitheatre, in the Adige Valley.

The trip will proceed along the western flank of the Mt. Baldo monocline, on the eastern shore of Lake Garda (national road 249), going through the villages of Torbole, Malcesine, Torri del Benaco, Garda as far as Bardolino. From here the trip continues towards the nearby Adige Valley (about 6 km to the E, on a direct line) reaching a panoramic position on the Rocca di Garda. This is a 295 m high relief consisting of Oligocene calcarenites overlapping Miocene arenaceous limestones; both formations are gently dipping towards Lake Garda. If the meteorological conditions are favourable, from this panoramic position it is possible to admire the entire Garda amphitheatre.

Looking NW, the tip of the St. Vigilio peninsula, bordered by a reverse fault named Sirmione-Garda fault, can be seen. The scarps on Mt. Luppia characterise the NE segment of this fault. Along the fault, three aligned thermal mineral springs are present inside the lake (CARRARO & *alii*, 1970).

Moving eastward, the junction between the Garda moraines belonging to the main Garda amphitheatre and the moraine deposits of a lesser amphitheatre can be seen. Moraine ridges of the Garda system face a smaller but well preserved system of frontal moraine ridges, close to the Rivoli village, deposited by the glaciers descending along the Adige Valley, where thick loess covers, red soils, and fluvioglacial terraces are also preserved. The moraine system consists of regular semicircular forms made up of several concentric ridges. Considering the good state of conservation of the moraines and their stratigraphic nature, Cremaschi (1990) dated the whole amphitheatre as Upper Pleistocene (fig. 29). It consists of three orders of concentric moraines progressively grading down toward Rivoli Veronese, in the inner part of the system. The distribution of soil types is more complex on the ridges of the moraines. Catenary sequences of soils consist here of lithosols at the very top of the moraines, of rendzinas and calcic soils along the slopes. Complex profiles, including colluvial cover and fersiallitic brown soil with Bt horizon (*Haploxeralfs*), are preserved in the lowermost part of the moraine ridges. Out of the Upper Pleistocene main moraine ridge, Middle Pleistocene eroded moraines and highly eroded and terraced glacial deposits pushed against the north reliefs of Caprino Veronese, are found. They are all ascribed to the Early Pleistocene.


FIG. 29 - Geological map of the Quaternary deposits of the Rivoli Veronese area. Legend: 1) alluvial deposits of the Tasso and Adige Rivers (Holocene); 2) alluvial and glaciolacustrine deposits of the Adige River (Late Glacial); 3) fluvioglacial deposits related to the retreat of the Garda and Rivoli glaciers (Upper Pleistocene); 4) fluvioglacial deposits related to the inner moraines (Upper Pleistocene); 5) inner morain ridges (Upper Pleistocene); 6) fluvioglacial deposits related to the Last Glacial Maximum (Upper Pleistocene); 7) moraine ridges of the Last Glacial Maximum (Upper Pleistocene); 8) widest loess outcrops (Upper Pleistocene); 9) eroded, very eroded or dissected moraine ridges (Early-Middle Pleistocene); 10) pre-Quaternary bedrock mainly limestones; 11) scree slope; 12) main scarp higher than 10 metres; 13) main scarp less than 10 metres; 14) alluvial fan; 15) moraine ridge.

The excursion continues downhill, toward the plain where the Torrent Tasso has cut through the Rivoli Veronese amphitheatre, proceeding along the eastern flank of Mt. Moscal. The latter is a ridge which reaches the maximum altitude of 427 m and is made up of Miocene limestones dipping toward Lake Garda. On top of it levels of Upper Pleistocene loess deposits are found, whereas its eastern slope is characterised by a distinct scarp over 100 m high, with a cover of Upper Pleistocene stabilised slope deposits at the base.

A short stop along the road rising from the south to the top of Mt. Moscal offers a panoramic view of the Rivoli Veronese amphitheatre.

Afterwards the trip continues downstream towards the village of Affi where the A22 motorway is followed heading south toward Mantova. After a few kilometres the

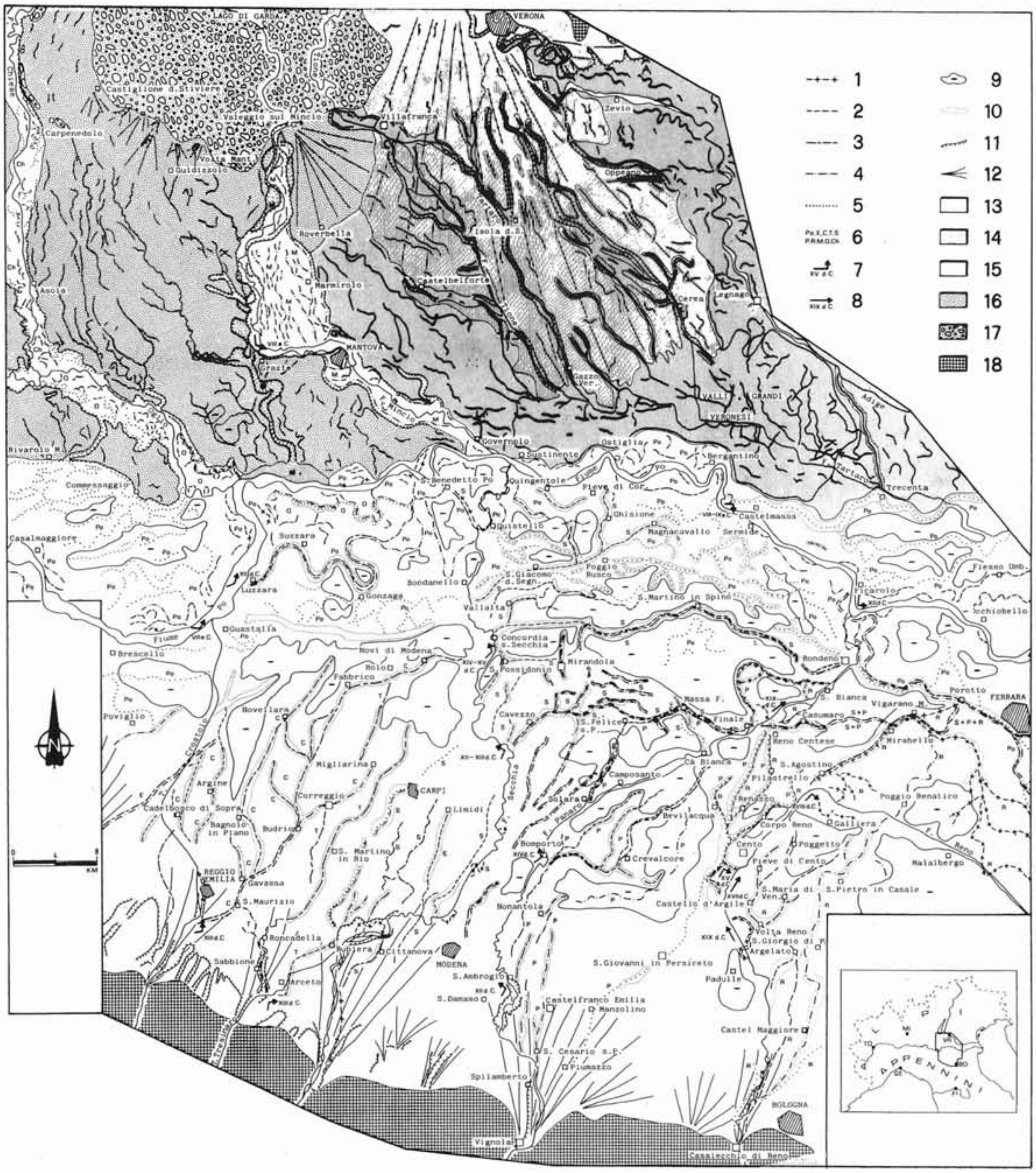



FIG. 30 - Geomorphological map of the Po Valley between Verona and Modena. Legend: 1) paleoriver of modern age; 2) paleoriver of late Middle Ages; 3) paleoriver of the early Middle Ages; 4) paleoriver of the Roman age or Iron age; 5) paleoriver of the Bronze age; 6) paleoriver domaine (PO for Po River; E for Enza River; C for Crostolo River; T for Tresinaro River; S for Secchia River; P for Panaro River; R for Reno River; M for Mincio River; O for Oglio River; Ch for Chiese River); 7) main fluvial deviation and its age; 8) main fluvial cut and its age; 9) depression in alluvial plain; 10) alluvial ridge; 11) scarp; 12) alluvial fan; 13) fluvial deposits (Holocene); 14) alluvial deposits (Sub-Boreal); 15) alluvial deposits (Late Glacial and Early Holocene); 16) fluvioglacial deposits (Upper Pleistocene); 17) glacial deposits (mainly Upper Pleistocene) 18) pre-Quaternary bedrock (after CASTALDINI, 1987).

landscape suddenly changes, the horizon widens as one enters the Po Valley, which is one of the main morphological units of the Italian peninsula (GUZZETTI & REICHENBACH, 1994) as well as the most extensive plain in Italy. Its surface area is approximately 46,000 $km^2$, corresponding to 71% of all the plains in Italy. It is the filled and emerged part of a marine-continental basin that could be defined as Adriatic. Two separate stratigraphic and morphological situations are distinguished proceeding from the Alpine northern margin to the Apennine southern margin.

The plain north of the River Po is made up of a surface, the so called «Po Plain Main Level» (PETRUCCI & TAGLIAVINI, 1969), which is no longer affected by a main hydrographic pattern but which preserves traces of abandoned watercourses with flow-rates much higher than the present ones (MARCHETTI, 1990). The present rivers run within broad valleys, dug-up into the «Po Plain Main Level», of which they often occupy a very small portion. New conditions at the northern side of the Po Plain arose after deglaciation, as the South Alpine lakes retained a large amount of debris and regulated the bankfull stages, reducing water supply and sediment discharge. This new situation produced the abandonment of the Po Plain Main Level, which therefore shows on a small scale the end of each aggradation stage (MARCHETTI, 1996) and the onset of extremely erosive conditions in the left-hand tributaries of the River Po.

On the other hand, the plain south of the River Po is made up of a N-NE slightly inclined surface, on top of which the right-hand tributaries of the River Po flow. On it, also the abandoned Holocene watercourses are usually visible, which are often characterised by convex relief landforms (alluvial ridges) (BONDESAN & *alii*, 1989). The aggradation of fluvial deposits in the southern part of the Po Plain, due to high subsidence, has been almost continuous. An important change from high-energy sedimentation to finer sedimentation occurred in sedimentary facies at the beginning of the Holocene (ALESSIO & *alii*, 1981), indicating a drop in the erosion rate and transport capacity of the watercourses (CREMASCHI & MARCHETTI, 1995). In fig. 30 the Po Valley portion which will be crossed during the journey to Bologna, where the excursion ends, is shown.

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