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POST-NAPPE FOLDING SOUTHEAST OF THE MISCHABELRÜCKFALTE (PENNINE ALPS) AND SOME ASPECTS OF THE ASSOCIATED METAMORPHISM

BY

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ABSTRACT

Three important folding phases, F_1, F_2 and F_3, have been found in the Pennine Nappes and the nappe-separating Palaeozoic and Mesozoic rocks between Saas Fee and Villadossola. The F_1 folds were formed at the beginning of the temperature increase that marks the second Alpine metamorphic event, and they resulted from penetrative deformation of the nappes and the nappe-separating rocks during the final stage of nappe emplacement. This stage was preceded by a stage of non-penetrative deformation during which most of the nappe transport is believed to have taken place. This earlier stage was characterized by thrusting along the interfaces between undeformed slabs of Hercynian or older crystalline basement rocks, and in the zones of nappe-separating rocks. In this study, the structural and metamorphic history after the beginning of F_1 will be chiefly considered, and a reconstruction of the structural geology before F_2 will be made.

Nearly all the F_1 folds outside the Moncucco zone have the same asymmetry, which reflects penetrative directional movement during the last stage of nappe emplacement. The axial directions of the F_1 folds, which are tight to rootless intrafolial folds of the layering, initially were not everywhere parallel. The folds occur on a microscopic and mesoscopic scale and have a pervasive axial-plane schistosity, S_1, which is the regionally prevalent S-plane. F_1 folding may have taken place at about 60–65 m.y. before present.

The F_2 folds postdate the nappe movements; they have refolded the nappe boundaries and the structures of F_1 age. The relationships between plagioclase porphyroblastesis, F_2 deformation and the time-temperature curve for the second Alpine metamorphic event show that F_2 folding took place at about 36 m.y. before present. The similar-style F_2 folds, that occur on every scale, are characteristically disharmonic and sometimes extremely so. A common type of disharmonic fold has limbs which are convex towards the axial plane, the outer layers having been more tightly folded than the core, and some major folds also have this geometry. Many of the F_2 folds have an S_2 axial-plane schistosity that in the majority of cases was formed by reorientation of pre-existing micas. It is found that the degree of reorientation depends on the lithology, on the fold geometry, and probably also on other factors.

All the major folds except one have F_2 age, and evidence for intensive F_2 deformation can be found almost anywhere. Well-known structures like the 'Synclinale rétrograde de Saas' and the Antrona Syncline were formed at this time. It is believed that the Antrona Syncline continues through the Monte Rosa Nappe to Mattmark, where it meets another major F_2 fold (the Trifthorn Antiform) and both folds terminate. The age and characteristics of F_2 folding in the Moncucco zone differ from the remainder of the area. This and the fact that the lithology of the Moncucco zone and the style of F_1 folding there also differ, suggests that the Moncucco zone may not be a part of the Pennine Nappes.

The F_3 folds postdate all mineral growth, and one major fold, the Brevettola Antiform, was formed during this phase. It is the hinge between the 'root-zone' of the nappes and the nappes themselves, and the overturning of the 'root-zone' is therefore the result of F_3 folding. The Brevettola Antiform is believed to be the continuation in the Camughera–Moncucco Complex of the Vanzone Antiform. All other F_3 folds occur on a microscopic to mesoscopic scale; they can be separated in two contemporaneous sets of folds with opposite asymmetry and opposed axial-plane dip directions.

RIASSUNTO

Tra la media Valle di Saas (piega retroflessa del Mischabel, vedi anche Zwart, in prep.) e la Valle della Toce sono state osservate tre fasi di piegamento che hanno dato origine rispettivamente a pieghe F_1, F_2 e F_3 nelle falde pennidiche, nel Paleozoico e nel Mesozoico interposti.

F_1 – Sono contemporanee all'ultimo stadio della messa in posto delle falde del cristallino (vedi oltre) e non deformano le superfici di contatto tra le falde stesse ed il Mesozoico. Le pieghe F_1 hanno una grandezza da centimetrica a decametrica, a seconda dalle potenza del layering litologico. Il profilo è fortemente tendente ad isoclinale, o anche rootless. Non sono state osservate scistosità anteriori allo sviluppo regionale della S_1; quest'ultima è piano assiale delle pieghe F_1, le quali deformano quindi esclusivamente l'alternanza di composizione litologica regionale. La mancanza locale di un'alternanza di composizione litologica può mascherare l'esistenza di pieghe, chiaramente rilevabili in aree adiacenti, dove essa è presente.

Nel cristallino gli assi inclinano verso N o NE ed assumono direzione SW nel complesso Camughera–Moncucco, mentre nel Mesozoico inclinano spesso verso W o NW.

Non è escluso che strutture alpine precedenti a F_1 si possano trovare nel Mesozoico.

S_1 – E' definita, tra l'altro, dalle miche e dalla clorite, le quali impongono un controllo sulla orientazione preferenziale dimensionale del quarzo.

L_1 – Rappresenta l'intersezione tra il compositional layering ed i piani S_1. E' diffusa nella zona Moncucco, dove le pieghe F_1 hanno fianchi più aperti.

F_2 – Le pieghe della seconda fase ripiegano i contatti tra le falde e il Mesozoico e sono quindi posteriori alla messa in posto delle falde. Esse ripiegano le pieghe F_1 (ed S_1), e in genere l'angolo tra gli assi delle pieghe F_1 e F_2 è grande.

Dalle relazioni microstrutturali è possibile ricavare per F_2 un'età rispetto alla blastesi metamorfica per F_2; nell'area di Saas i porfiroblasti di plagioclasio, il cui orlo oligoclasico segna la culminazione termica di 38 m.a. (fase lepontina), sono precedenti a F_2, mentre dopo questa fase di deformazione si registra solo la crescita di quantità minori di individui esclusivamente albitici. Dal complesso delle osservazioni microstrutturali viene indicata per F_2 un'età di poco posteriore a 38 m.a.

Le pieghe F_2 sono di stile simile e quasi sempre disarmoniche; esse hanno fianchi variabili tra il mm e qualche km. Su tutte le scale, la geometria delle pieghe è controllata dalla litologia. Molto caratteristiche sono le pieghe con fianchi convessi verso il piano assiale, che si ritrovano frequentemente a piccola scala, ma anche su scala regionale, come la sinforme del Mittaghorn e l'antiforme del Trifthorn.

Gli assi delle pieghe F_2 immergono verso W o NW, così come i loro piani assiali. In tutta l'area le pieghe F_2 sono responsabili delle strutture a scala maggiore (ad eccezione dell'antiforme di Brevettola).

E' possibile individuare in corrispondenza ad interi fianchi normali od inversi di grandi pieghe F_2 zone altamente omogenee; esse sono comprese tra le tracce dei piani assiali delle sinforme del Mittaghorn, l'antiforme del Trifthorn e la sinclinale di Antrona.

La sinforme del Mittaghorn sembra coincidere con la 'Sinclinale rétrograde de Saas' di Staub (1936). L'antiforme del Trifthorn è la continuazione (Nieuwland, 1975) della piega retroflessa del Balmahorn (Te Kan Huang, 1935) nelle valli di Zwischbergen e Laggin, a S del Sempione. La sinclinale di Antrona (Argand, 1911) è interrotta verso NE dalla faglia Sempione–Centovalli, e il suo fianco settentrionale continua verso SW costituendo ancora una vasta zona a deformazione omogenea, tra Bognanco e la Valle di Saas, che comprende gran parte della Zona del Portjengrat e quasi tutta la Zona di Furgg.

E' probabile che la sinclinale di Antrona continui nella falda Monte Rosa fino a S della Zona di Furgg e ne rovesci la parte frontale. E' proprio qui che il piano assiale della sinclinale di Antrona e quello dell'antiforme del Trifthorn si incontrano, nella regione di Mattmark (Valle di Saas), ove entrambe le pieghe muoiono.

S_2 – E' diffusa in tutti i tipi litologici delle varie unità; l'obliterazione di S_1 nella Zona Moncucco ne ostacola talvolta l'individuazione. Nella Zona del Portjengrat nessun motivo microstrutturale suggerisce che una nuova generazione di mica bianca abbia contribuito a costituire questa scistosità. Si osserva invece una progressiva rotazione di individui di mica bianca preesistenti alla S_2 verso la nuova orientazione, anche quando i fianchi delle pieghe F_2 sono aperti (60°–90°).

L_2 – La lineazione L_2 può essere definita dalla orientazione preferenziale di anfibolo, mica bianca e aggregati quarzo–feldspatici, da una lineazione di crenulazione o può essere anche l'intersezione tra S_1 e S_2.

F_3 – Le pieghe F_3 ripiegano le pieghe F_2 ed i piani S_2. Gli assi sono paralleli agli assi F_2, ad eccezione della Zona Moncucco ove formano con questi un angolo anche grande. Pieghe F'_3 si sviluppano contemporaneamente alle F_3, ma con asimmetria opposta e piani assiali che inclinano a SE, mentre per le F_3 essi hanno inclinazione verso NW o sono verticali con direzione EW. Tutte le pieghe di terza generazione hanno stile parallelo e sono in certi casi delimitate da piani di dislocazione. Le pieghe F'_3 non hanno importanza su scala regionale; al contrario una grande piega F_3 è presente nel complesso Camughera–Moncucco, ove affiora in Val Brevettola (antiforme di Brevettola). Quest'ultima costituisce la transizione tra la 'zona di radice' e le falde ed è sicuramente prodotta da un evento successivo ad F_2 perché i piani S_2 cambiano nettamente inclinazione, da semiorizzontale nel fianco settentrionale a semiverticale nel fianco meridionale. Molti fattori indicano che l'antiforme di Brevettola sia il prolungamento verso NE dell'antiforme di Vanzone, per la quale Laduron (1974) ha pure proposto un'età di terza generazione. Non è improbabile che il piano assiale dell'antiforme di Brevettola tagli la faglia Sempione–Centovalli, e si prolunghi in direzione della grande piega ove la falda Antigorio si raccorda con la sua 'zona di radice', ad Enso (Val Divedro).

Sulla tettonica a grande scala si possono trarre le seguenti conclusioni:

a. la Zona di Furgg, ora in posizione rovesciata perchè si trova nel fianco inverso (settentrionale) della sinclinale di Antrona, costituiva prima di F_2 la copertura della Zona del Portjengrat, ed era a sua volta ricoperta dalla parte frontale della falda Monte Rosa, che ne ha causato la intensa deformazione durante il sovrascorrimento.

b. la Zona di Camughera fa parte della falda Gran San Bernardo. Svolgendo le pieghe F_3 ed F_2 nell'area studiata, la Salarioli-Mulde risulta essere la copertura della Zona di Camughera ed è a sua volta ricoperta della Zona Moncucco.

c. durante il processa di messa in posto delle falde si individuano due fasi rispetto al regime termodinamico ed alle condizioni di strain; esse si realizzano in tempi successivi. Si può concludere, in accordo con le osservazioni di Bortolami & Dal Piaz (1970) e di Compagnoni & Lombardo (1974) sulla falda Monte Rosa e nel Massiccio del Gran Paradiso, che la riattivazione tettonica del basamento prealpino inizia con l'individuazione, in regime termico basso, e/o ad alto strain rate, di piani di scorrimento isolati, mentre la deformazione penetrativa, associata allo sviluppo di pieghe F_1 e piani S_1, ha luogo solo in un tempo successivo, quando il regime termico è divenuto più elevato.

SAMENVATTING

In de volgende pagina's wordt de geologie beschreven van een deel van de Penninische Alpen tussen het Saasdal (Wallis, Zwitserland) en het dal van de Toce (Novara, Italië). Het onderzochte gebied wordt geologisch begrensd door de zuidelijke, overkiepte flank van de Mischabelrückfalte noordelijk van Saas Fee (Zwart, in voorber.) en de wortelzone van het Monta Rosa dekblad bij Villadossola. Het omvat het noorden van het Monte Rosa dekblad (d.w.z. ongeveer de Portjengrat zone), een klein zuidelijk deel van het Bernhard dekblad, en een deel van het Camughera–Moncucco Complex samen met de tussenliggende Paleozoïsche en Mesozoïsche gesteenten. Deze dekbladen behoren tot de Penninische dekbladen (Wenk, 1956), en ook het Camughera–Moncucco Complex rekent men daartoe, ofschoon sommige auteurs waaronder de schrijver menen dat dit op de Moncucco zone niet van toepassing is (zie hierna). De dekbladen bestaan uit metamorfe en stollingsgesteenten van Hercynische of Caledonische ouderdom, wier mineraalbestand door de Alpine metamorfose sterk gewijzigd is. De dekbladscheidende gesteenten, die jonger zijn dan de gesteenten van de dekbladen, zijn uitsluitend Alpien-metamorfe sedimentaire en stollingsgesteenten, zoals in de Furgg zone (veel Paleozoïsche gesteenten), in de Antrona

Syncline (voornamelijk stollingsgesteenten van Mesozoïsche ouderdom), en in de directe omgeving van Saas Fee waar Mesozoïsche sedimenten van betekenis zijn.

In dit gebied komen drie belangrijke plooiingsfasen voor, die van oud naar jong F_1, F_2 en F_3 worden genoemd. Zij zijn identiek aan de drie plooiingsfasen in het Mischabelrückfalte gebied.

De plooien van de eerste fase zijn gedurende het laatste bewegingsstadium van de dekbladen ontstaan; zij zijn meestal asymmetrisch en door de gerichte beweging domineert een asymmetrie. Er zijn geen voorbeelden bekend van F_1 plooien die het contact tussen de dekbladen en de dekbladscheidende gesteenten vervormen. Zij zijn plooien van de gelaagdheid, spits tot isoclinaal of ook wel van het 'intrafolial' of 'rootless intrafolial' type met een assenvlakschistositeit S_1. Deze laatste wordt gedefiniëerd door phyllosilicaten in de schisten en gneisen, en door amfibolen in de hoornblendegneisen en amfibolieten; in de laatste gesteenten echter hoeft hij niet altijd aanwezig te zijn. S_1 is het dominerende schistositeitsvlak in het onderzochte gebied. Er zijn geen aanwijzingen gevonden voor het bestaan van een vroegere schistositeit van Alpine ouderdom. Toch zijn oudere plooien dan F_1 wel waargenomen en het lijkt niet uitgesloten dat speciaal in de dekbladscheidende gesteenten een oudere Alpine plooiingsfase plaatsgevonden zou kunnen hebben. De geschatte ouderdom van de eerste plooiingsfase is 60–65 miljoen jaar.

De plooien van de tweede fase herplooien de structuren van de eerste fase, waarbij een grote hoek tussen de asrichtingen van de F_1 en F_2 plooien overweegt. De F_2 plooien hebben het contact tussen de dekbladen en de dekbladscheidende gesteenten vervormd en zijn daarom na de laatste dekbladbewegingen ontstaan. De relaties tussen de F_2 deformatie, de porfiroblastese van plagioklaas, en de curve die het temperatuurverloop in de tijd aangeeft, maken het mogelijk om de ouderdom van de tweede plooiingsfase op om en nabij 36 miljoen jaar vast te stellen. De F_2 plooien zijn in beginsel vormgetrouwe plooien, maar zij zijn bijna altijd disharmonisch door viscositeitsverschillen tijdens de deformatie, die het gevolg zijn van verschillen in de lithologie. Dit is toepasselijk op plooien van iedere afmeting. De disharmonie heeft bij veel F_2 plooien een karakteristieke geometrie veroorzaakt, die gekenmerkt wordt door flanken met een convexe vorm naar het assenvlak toe, zodat de buitenste geplooide lagen een spitsere plooi vormen dan de binnenste. Er zijn zowel symmetrische als asymmetrische plooien, en in veel gevallen hebben de F_2 plooien een assenvlakschistositeit S_2 die door de voorkeursoriëntatie van kleurloze mica en biotiet wordt gedefiniëerd, behalve in de ambibolieten waar de amfibolen S_2 definiëren. De voorkeursoriëntatie van kleurloze mica's in S_2 wordt met name in het westelijk deel van het gebied veroorzaakt door heroriëntatie van bestaande mica's. Biotiet is in S_2 gekristalliseerd of gerekristalliseerd en het is niet onmogelijk dat in het oosten ook kleurloze mica in S_2 is gerekristalliseerd. Afgezien van de heroriëntatie van kleurloze mica's in isoclinale plooien, kon de aard van het heroriënterend mechanisme niet volledig worden bepaald. Wel is gebleken dat de kwaliteit van de heroriëntatie afhankelijk is van de gesteentesamenstelling en er een zeker minimaal kwartsgehalte aanwezig moet zijn.

De amplitudes van de F_2 plooien variëren tussen minder dan een millimeter en enige kilometers, en de kleinere plooien komen als parasitische plooien op de grotere voor. De grootste plooien zijn: (1) de Mittaghorn Synform, vermoedelijk identiek met de 'Synclinale rétrograde de Saas' (Staub, 1936), een extreem disharmonische plooi die de zuidelijke overkiepte flank van de Mischabelrückfalte in de normale stand terug brengt; (2) de Trifthorn Antiform – volgens Nieuwland (1975) de voortzetting van de Balmahornrückfalte noordoostelijk van het hier behandelde gebied – die een antiformale structuur van de Portjengrat zone heeft veroorzaakt en de Furgg zone heeft overkiept. Zijn zuidelijke flank komt overeen met de noordelijke flank van (3), de Antrona Syncline ('Cuillère de Bognanco', Argand, 1911). Deze laatste zet zich voort in het Monte Rosa dekblad en het moet als vrijwel zeker worden beschouwd dat de assenvlakken van de Antrona Syncline en de Trifthorn Antiform elkaar in de westelijke hellingen van het Saasdal nabij Mattmark ontmoeten. Dit heeft tot gevolg dat beide ombuigingen verdwijnen en de overkiepte S_1 van hun gemeenschappelijke flank ophoudt, zodat de normaal liggende S_1 van de zuidflank van de Mittaghorn Synform doorloopt in het Monte Rosa dekblad zuidelijk van het assenvlak van de Antrona Syncline.

Van betekenis is de grote parasitische plooi in de Camughera zone, die door herhaling van de orthogneisen suggereert dat de schijnbare dikte van de Camughera zone groter is dan zijn werkelijke dikte. De F_2 plooien in de Moncucco zone verschillen naar ouderdom en eigenschappen van de gelijksoortige plooien elders in dit gebied. De lithologie van de Moncucco zone en de karakteristieken van de F_1 plooien aldaar zijn ook afwijkend, en dientengevolge wordt de Moncucco zone hier niet beschouwd als een deel van de Penninische dekbladen.

De plooien van de derde fase hebben oudere structuren herplooid, en met uitzondering van een gebied dat ongeveer de Moncucco zone omvat en een klein deel van de Antrona Syncline noordoostelijk van het Bacino Alpe dei Cavalli, zijn hun assen subparallel aan die van de F_2 plooien. De derde fase plooien kunnen gescheiden worden in twee gelijktijdig gevormde groepen van plooien, F_3 en F'_3, die een aantal eigenschappen gemeen hebben (zie hierna) maar die verschillen door hun tegengestelde asymmetrie en de tegengestelde hellingsrichtingen van hun assenvlakken. Alle derde fase plooien zijn afstandsgetrouw of van het chevron type, en werden na iedere groei van metamorfe mineralen gevormd. Hun ouderdom kon niet nauwkeurig worden vastgesteld, maar zou tussen 16 en 8 miljoen jaar kunnen liggen.

Van regionaal standpunt bekeken zijn de F'_3 plooien niet van betekenis, in tegenstelling tot de F_3 plooien. Tot deze laatste groep van plooien behoort de Brevettola Antiform die de overgang is tussen de steilstaande 'wortelzone' en de vlakker liggende dekbladeenheden noordelijk daarvan. Deze plooi demonstreert door zijn F_3 ouderdom dat het steilzetten van de 'wortelzone' een laat verschijnsel is. Het is vrijwel zeker dat de Brevettola Antiform zich naar het zuidwesten voortzet in de Vanzone Antiform, de klassieke ombuiging die het Monte Rosa dekblad met zijn 'wortelzone' verbindt. Er wordt vermoed dat de Brevettola Antiform zich ook in noordoostelijke richting voortzet, naar de antiform bij Enso (Val Diveria) waar het Antigorio dekblad in zijn 'wortelzone' overgaat. Dit brengt met zich mee dat de tussenliggende Simplon–Centovalli storing door F_3 geplooid moet zijn.

Tot de resultaten van deze studie, die deels hiervoor genoemd zijn, behoort een reconstructie van de structurele geologie voor F_2 (Fig. 7-1) met als uitgangspunt dat de dekbladoverschuivingen geen of slechts een geringe hoek met het horizontale vlak maakten. Het blijkt dat de Portjengrat zone over korte afstand op het Bernhard dekblad is overschoven en dat het zuidelijke deel van het Monte Rosa dekblad is overschoven op de Portjengrat zone met de Furgg zone als bewegingszone; deze laatste wordt hier beschouwd als de (para?)autochtone bedekking van de Portjengrat zone. Deze beweging is lang voor F_1 begonnen maar het lijkt waarschijnlijk dat hij tijdens de vorming van F_1 en S_1 nog niet was opgehouden. De Camughera zone wordt vanwege zijn ligging onder de Antrona amfibolieten als een deel van het Bernhard dekblad beschouwd. De Moncucco zone was reeds voor F_2 de bedekking van de Camughera zone en dus van een klein deel van het Bernhard dekblad, een relatie die overeenkomt met die van de Furgg zone t. o. v. de Portjengrat zone die volgens Dal Piaz (1966) ook geldt voor de Furgg zone en het Monte Rosa dekblad als geheel.

Uit de geraadpleegde literatuur en de tijdens het onderzoek gemaakte observaties kan verder worden afgeleid dat er twee stadia in de vorming van de dekbladen waren. Het eerste stadium omvat het grootste deel van het dekbladtransport, d.w.z. gerichte beweging langs overschuivingsvlakken waarbij geen interne deformatie van de dekbladeenheden optrad. Het tweede stadium bestaat uit gerichte beweging gekoppeld aan interne deformatie, waarbij F_1 en S_1 werden gevormd. In het hier onderzochte gebied, met name in het noordelijke deel van het Monte Rosa dekblad, werd het latere stadium geassisteerd door een temperatuursverhoging die het begin vormt van het tweede deel van de Alpine metamorfose, waaraan alle in deze studie beschreven mineraalgroei gekoppeld is.

CONTENTS

CHAPTER 1

INTRODUCTION

1.1. THE AREA STUDIED AND THE SUBJECTS OF RESEARCH

The structural geology of the Pennine Nappes (Wenk, 1956) between Saas Fee in the Swiss Canton of Wallis and the Val d'Ossola (Province of Novara, Italy) is the subject of this paper. Three key areas were studied in detail and they are (Fig. 1-1; Enclosure 2):

1. The Saas area, between Saas Fee and the Swiss–Italian border, which covers the northern part of the Monte Rosa Nappe (Tekt. Karte der Schweiz, A. Spicher, 1972).
2. An area enclosing the eastern end of the Furgg zone (Wetzel, 1972)* and the adjoining parts of the Antrona Syncline (Blumenthal, 1952)*.
3. A section through the eastern parts of the Antrona Syncline and the Camughera–Moncucco Complex (Bearth, 1956a) which terminates in the 'root-zone' of the Monte Rosa Nappe.

The following subjects will be dealt with:

a. An evaluation of the importance of the post-nappe folding phases south and southeast of the Mischabelrückfalte (Zwart, in prep.).
b. The construction of a profile (Enclosure 1) from Saas Fee through the frontal part of the Monte Rosa Nappe, the Antrona Syncline and the Camughera–Moncucco Complex to the 'root-zone' of the Monte Rosa Nappe.
c. The interpretation of the 'root-zone' as a late structure (Laduron & Merlyn, 1974).
d. The relations between the folding phases and the growth of the metamorphic minerals.

The use and spelling of topographic names occurring in the present work are after the existing and official topographic maps.

*) References are confined to the most recent papers until 1/1/1975.

1.2. GEOLOGICAL SETTING

1.2.1. *The present area (Fig. 1-1; Enclosures 1 and 2)*

The northern part of the Saas area consists of paragneisses belonging to the zone of Mischabel–Siviez, a constituent of the Bernhard Nappe (Bearth, 1962). Most of the Saas area lies in the Portjengrat zone, the frontal element of the Monte Rosa Nappe (Bearth, 1964, 1967; Tekt. Karte der Schweiz 1:500.000, A. Spicher, 1972). The Portjengrat zone consists of para- and orthogneisses (Bearth, 1952a, 1957a), and is separated from the Bernard Nappe by an interrupted level of Mesozoic rocks (Te Kan Huang, 1935; Bearth, 1939; Roesli, 1946).

The Furgg zone, which separates the Portjengrat zone from the main body of the Monte Rosa Nappe, also reoccurs in the Monte Rosa Nappe further south (Dal Piaz, 1964, 1966; Wetzel, 1972). Its main constituents are metasediments and metamorphic mafic rocks of Palaeozoic and Mesozoic age, subjected to particularly strong Alpine deformation (Bearth, 1953; Wetzel, 1972). For ease of reference, the Monte Rosa Nappe south of the Furgg zone is called here the Monte Rosa Nappe s.s., in contrast to the Monte Rosa Nappe s.l. that includes the Portjengrat zone and the Furgg zone. The Monte Rosa Nappe s.s. comprises Alpine metamorphic rocks and pre-Alpine metamorphic and igneous rocks (Bearth, 1952a). It also contains the Stelli zone (Bearth, 1952a, 1957a) which has a lithology similar to the Portjengrat zone, and adjoins the Furgg zone to the south.

The Antrona Syncline ('Cuillère de Bognanco', Argand, 1911; 'Antrona Mulde', Blumenthal, 1952) contains the Antrona amphibolites (metamorphosed Mesozoic mafic rocks) and subordinate calcareous schists. Its southern limb separates the Monte Rosa Nappe s.s. from the Camughera–Moncucco Complex.

The eastern part of the Bernhard Nappe is enclosed by the northwestern extension of the Antrona amphibolites,

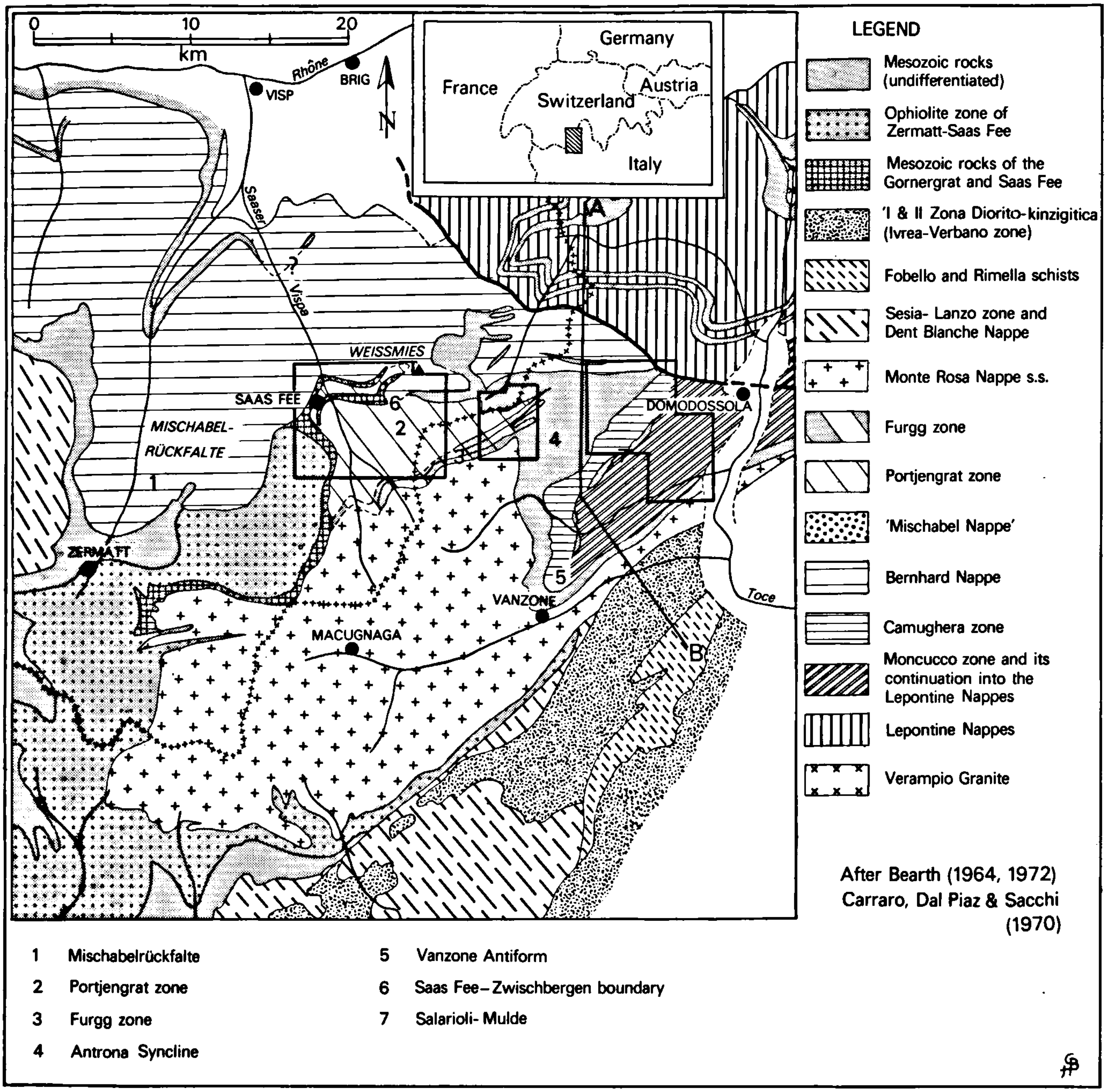


Fig. 1-1. An outline of the regional geology of the present area and its surroundings. The line A–B is the approximate profile scale for Fig. 1-4. The enclosed areas were studied in detail for this work.

the northern limb of the Antrona Syncline and the Simplon–Centovalli Fault (Hunziker, 1970), and its rocks are called the Verosso gneisses.

The Camughera–Moncucco Complex contains the Camughera zone, where orthogneisses predominate, which is followed by the 'Salarioli-Mulde' (Argand, 1911; Bearth, 1939, 1956a). The 'Salarioli-Mulde' is defined by a number of lenses of calcareous schists associated with other, presumably Mesozoic rocks. Argand (1911, 1916) and Blumenthal (1952) consider it as a nappe boundary, but Bearth (1956a) gives it a lower status and regards it as the boundary between the Camughera and Moncucco zones. To the south lies the Moncucco zone, composed of paragneisses, amphibolites and orthogneisses, which have been correlated with the 'Orselina-Serie' (Bearth, 1956a; Knup, 1958; Wieland, 1966) and the 'Isorno-Serie' (Wieland, 1966), both occurring on the northern side of the Simplon–Centovalli Fault. The Moncucco zone contains the ultramafic body of Montescheno and the pegmatite of the 'Cava di Mica I Mondei' (Traverso, 1895; Pagliani & Martinenghi, 1941; Ferrara et al., 1962).

South of the Camughera–Moncucco Complex lie the 'root-zone' occurrences of the Antrona amphibolites and the Monte Rosa Nappe s.l. The 'root-zone' of the latter is synonymous with the Monte Rosa zone.

1.2.2. *Surroundings of the present area*

The Monte Rosa Nappe s.l. is imperfectly covered by Permotriassic metasediments (Bearth, 1952a, 1957a), upon which lies the Ophiolite zone of Zermatt–Saas Fee, according to Bearth (1974), the equivalent of a former oceanic crust. The zone of Zermatt–Saas is covered by the Combin zone s.l., comprising ophiolite-free and ophiolite-bearing metasediments that were deposited on continental crust, and later emplaced upon the zone of Zermatt–Saas (Bearth, 1974). The Dent Blanche Nappe overlies the Combin zone s.l.

The zone of Mischabel–Siviez is overlaid by Carboniferous, Permotriassic and younger rocks in normal stratigraphical order. The internal structure of the zone of Mischabel–Siviez is determined by the southward-facing, anticlinal Mischabelrückfalte ('Pli en retour du Mischabel', Argand, 1911; Güller, 1947; see Bearth (1964) and Zwart (in prep.) for profiles). This fold continues into the Mesozoic rocks northwest of Zermatt where its geometry changes (Bearth, 1964; Enclosure 1, Inset 2).

South of the Monte Rosa zone lies the zone of Sesia–Lanzo and the 'Seconda Zona Diorito-kinzigitica'; the latter is a portion of the Ivrea–Verbano zone (the 'Zona Diorito-kinzigitica' proper) that has been overthrust on the Sesia–Lanzo zone (Carraro et al., 1970), where a large synform postdates the thrusting event since it folds the contact between the Sesia–Lanzo zone and the 'Seconda Zona Diorito-kinzigitica' (Dal Piaz et al., 1971). The overthrust is of large dimensions, for both units reoccur in the Dent Blanche Nappe where the rocks from the 'Seconda Zona Diorito-kinzigitica' form the upper unit, the 'Serie di Valpelline'; and rocks from the Sesia–Lanzo zone the lower unit, the 'Serie di Arolla' (Carraro et al., 1970). Between the Ivrea–Verbano zone and the 'Seconda Zona Diorito-kinzigitica' lie the so-called Fobello and Rimella schists, that according to Baggio & Friz (1968) have been produced by intense Alpine metamorphism and deformation of Sesia–Lanzo zone rocks. Johnson (1973) calls the Fobello and Rimella schists a mylonite zone which is separated from the Ivrea–Verbano zone by the Insubric Line. This was confirmed by Isler & Zingg (1974).

1.3. THE PENNINE NAPPES AND THEIR 'ROOT-ZONES'

1.3.1. *General*

The name Pennine Nappes, derived from the Pennine Alps where they occur, is used in the sense of Wenk (1956) who separated them from what he called the Lepontine gneiss region, the Lower Pennine Nappes of earlier authors and synonymous with the Lepontine Nappes.

Argand's (1911, 1916) classic synthesis of the Pennine Nappes, which incorporates also the Dent Blanche Nappe and the Lepontine Nappes as far east as the Toce culmination, contains among others the following important views on nappe geometry (the 'root-zone' concept will be discussed later):

a. The nappes are separated by continuous layers of Mesozoic rocks.

b. The nappes are cylindrical, recumbent folds with frontal hinges.

Argand states that it is necessary to demonstrate the existence of the inverted limb and the recent substrate in order to prove the existence of a nappe.

As a result of (a), the boundary between the Monte Rosa and Bernhard Nappes has been a subject of constant discussion in the literature (see 1.4.2). Point (b) is now considered to be obsolete (Rutten, 1969; Ayrton & Ramsay, 1974) and the nappes are thrust sheets. In the Pennine Alps, this view was held as early as 1925 by Hermann and it has recently been reconsidered by Bortolami & Dal Piaz (1970), who describe a gradual transition from an overthrust to a more or less autochthonous 'root-zone' in the Val d'Aosta.

Argand (1911) attributed great importance to the frontal hinges of his fold nappes, and in the older literature they contribute to the evidence for the existence of a nappe (e.g. Wenk, 1953; see also 5.2.1). These frontal hinges will have to be reinterpreted now that the nappes are interpreted as thrust sheets.

The nappes have been subjected to one or more phases of deformation after their emplacement. In the present area, this has been demonstrated by Bearth (1952a, 1956a, 1958, 1967), Wetzel (1972), and Zwart (in prep.).

1.3.2. *The 'root-zones'*

In Argand's (1911) view, the nappes are collectively 'rooted' in the so-called root-zone in the southeast. The Bernhard Nappe 'root' is the Camughera zone, and the Monte Rosa Nappe is 'rooted' in the Monte Rosa zone. Argand (1911, 1916) considered the 'root-zone' as the place where the fold nappes originated, and as a region of extreme crustal shortening, where the nappes are connected to autochthonous rocks. Argand's concept includes an initially almost recumbent position of the 'root-zone', that would have obtained its present steeply dipping position after the formation of the nappes (Argand, 1911, 1916; Staub, 1924).

In later work, the steeply dipping position of the 'root-zones' was considered, if not as a primary feature, at least as a characteristic of 'root-zones'. Cadisch (1953) defines 'root-zones' as "steeply dipping central zones of mountain chains formed by nappes", and Zawadyński (1952) uses 'root-zone position' as synonymous with 'steeply dipping'. Wenk (1953) considers the steeply dipping attitude as a characteristic property of the 'root-zones' in the Lepontine Nappes.

Some definitions of 'root-zones' have been cited by

Metz (1967): Cornelius (1940): "The areas that lie opposed to the direction of movement of the nappes and where they finally disappear into the depth". Staub (1949): "The zones where the overthrust masses have slowly grown out of". Bailey (1935) said that tracing a nappe to its root is the search for an exposure where the nappe finally passes downward out of sight.

These definitions do not contain the 'characteristic' steep dip of the 'root-zone', but neither do they attribute any special properties to a 'root-zone', by which it could be recognized as such, apart from the disappearance of the nappe under the surface. These definitions would not allow an investigator to recognize a 'root-zone' when the corresponding nappe has been eroded away, and this is also impossible with the definitions of earlier authors. Though a 'root-zone' would have to be 'strongly tectonized' (doubted by Wenk, 1953), a zone subjected to high deformation certainly is not always a 'root-zone'.

It is likely that the steep dip of many 'root-zones' has been emphasized because a 'root-zone' has insufficient characteristic properties to define it without incorporating the corresponding nappe in the definitions. This is substantiated in the United States by Freedman et al. (1964), who conclude that if roots to the nappes of the Great Valley of Pennsylvania do exist, the only difference between the 'root-zones' and the nappe bodies is the steeply dipping attitude of the former.

Kündig (1936) states that the transversal zone of Maggia is comparable to the 'root-zone' of the Lepontine Alps. In the Upper Val Verzasca (Corona di Redorta, Val Pertusio) a large antiform connects the flat-lying rocks of the Lepontine Nappes in the east to the steeply dipping rocks in the transversal zone of Maggia. This reminds Kündig of the backfolding of the Pennine Nappes, but the connexion between the 'root-zone' of the Lepontine Nappes and the transversal zone of Maggia makes him believe that the latter is another 'root-zone'. Kündig, in fact, observed that the steeply dipping attitude of the 'root-zones' could have been caused by folding of the nappes.

In more modern work, there is a return to Argand's concept that the 'root-zone' was brought in its present position after the formation of the nappes. This was suggested by Bearth (1957b, 1958; see below). Wenk (1962b), recalling his previous interpretations, wrote that the 'root-zone' might not have the meaning traditionally attributed to it, but he did not suggest an alternative interpretation. He was inspired by the find of large, tight folds with vertical axial planes in the 'root-zone' at the construction site of a dam in the Val Verzasca.

Wunderlich (1966) wrote that "the root-zone seems to represent the most compressed portion of the orogeny, possessing its present steep position thanks to its erection after the transport of the nappes". This is essentially the opinion of Argand (1911, 1916).

Bortolami & Dal Piaz (1970) use the name 'root' for the area of tectonic provenance, in a mainly palaeographic sense. The steep dip of the 'root-zone' is in their opinion the result of a final phase of compression of the nappes.

1.3.3. *The Vanzone Antiform*

Gerlach (1883) found this large fold, the classic hinge between the recumbent body of the Monte Rosa Nappe and its overturned and steeply dipping 'root-zone', near Vanzone (Val Anzasca) where it also affects the Camughera–Moncucco Complex and the Antrona amphibolites. It has been described by Argand (1911, 1916), Blumenthal (1952), Bearth (1939, 1956a, 1957b, 1958), Laduron (1974), and Laduron & Merlyn (1974).

Blumenthal (1952) surmised that the Vanzone Antiform changes its shape, becoming isoclinal at a deeper level and in the direction of the present area, since its axial plunge is towards the southwest. Bearth (1957b) traced it into the Monte Rosa Nappe and demonstrated its importance there, but did not trace it into the Camughera–Moncucco Complex. He (Bearth 1957b, 1968) argues that the formation of the Vanzone Antiform and the overturning of the 'root-zone' are related to a phase of compression of the nappes.

Laduron (1974), and Laduron & Merlyn (1974) consider the Vanzone Antiform as an essentially postcrystalline fold that has deformed the nappes. They show that it folded the Alpine albite–oligoclase isograde (mapped by Bearth (1958) in, and east of the Monte Rosa Nappe), which convincingly demonstrates that the formation of a steeply dipping 'root-zone' is a late and post-metamorphic event. Laduron and Merlyn (1974) observe that the Vanzone Antiform might have its equivalent in the Val Brevettola, in the Camughera–Moncucco Complex.

1.3.4. *Folds in the 'root-zones'*

Wenk (1962b) described folds on a scale of several tens of metres from the 'root-zone' of the Lepontine Nappes.

Numerous folds of this scale and larger have been found in the 'root-zones' of the Pennine Nappes. In a profile through part of the Sesia–Lanzo zone, Baggio & Friz (1968, 1969) found close-tight folds with subvertical to NW dipping axial planes, a nearly horizontal enveloping surface, and folding a schistosity plane. These folds also occur in the Fobello and Rimella schists, and in a small portion of the Ivrea–Verbano zone, where they are contiguous with an antiform previously observed by Schilling (1957). Baggio & Friz (1968) consider these folds as "contemporaneous with the formation of the major Alpine structures", and believe them to be related to faults that they associate with the Insubric Line. Carraro et al. (1970) describe a large synform (see 1.2.2) from the Sesia–Lanzo zone, and Dal Piaz et al. (1971) attribute a late Alpine age to it. Isler & Zingg (1974) observed two generations of small-scale Alpine folds, also in the Sesia–Lanzo zone.

Blumenthal (1952) and Reinhardt (1966) describe a large, tight antiform east of the Val d'Ossola in the Monte Rosa zone. In Blumenthal's opinion this fold is characteristic of a 'root-zone', but Wenk (1956) does not share this view. The hinge of this antiform has a width of approx. 1 km (Reinhardt, 1966), and his profiles suggest that this antiform refolds earlier isoclinal folds. Reinhardt considers it to be contemporaneous with the 'major deformation phase of Alpine orogenesis', but it is pre-crystalline (Reinhardt, 1966) and therefore evidently not of the same age as the postcrystalline folds mapped by Baggio and Friz in the Sesia–Lanzo zone.

Laduron (1974) found four generations of folds in the Monte Rosa zone, and his observations have important implications for the structure of the present area (see 7.3).

1.4. PREVIOUS STRUCTURAL WORK IN THE PRESENT AREA

1.4.1. *Some disputed aspects of Argand's synthesis*

A small part of the structural map by Argand (1911), covering the present area, is reproduced here in Fig. 1-2A (after Bearth, 1939). The following aspects of this map have raised controversies:

1. The boundary between the Bernhard Nappe and the Monte Rosa Nappe. Argand connects the Mesozoic calcareous schists in the Saastal near Saas Grund with similar rocks at Zwischbergen, north of the Portjengrat, where the northern limb of the Antrona Syncline terminates.
2. East of the Antrona Syncline exists a connexion between the Bernhard Nappe and the Camughera zone. Argand considers the Camughera zone as the 'root' of the Bernhard Nappe, traceable until Villadossola.
3. The assignment of the Moncucco zone to the Monte Leone Nappe.
4. The termination of the Furgg zone within the Monte Rosa Nappe.

1.4.2. *The boundary between the Bernhard Nappe and the Monte Rosa Nappe*

Te Kan Huang (1935) observed that the Mesozoic rocks between Saas Grund and Zwischbergen were discontinuous. He traced another nappe-boundary further to the north, along isolated occurrences of, presumably, Mesozoic quartzitic rocks.

Staub (1937) used the discontinuity of the Mesozoic rocks near Saas Grund as an argument for the existence of a Mischabel Nappe, a very large nappe comprising the Bernhard and Monte Rosa Nappes. Bearth (1939) shared this view, arguing that the quartzites forming Te Kan Huang's boundary belong in the Bernhard Nappe. He investigated the continuity of the Bernhard Nappe east of the Antrona Syncline, and concluded that the connexion between the Bernhard Nappe and the Camughera zone was fictitious, thus depriving the Bernhard Nappe of its 'root' (see also 1.4.3). This he used as another argument for the existence of a Mischabel Nappe, that would be 'rooted' in the 'root-zone' of the former Monte Rosa Nappe (Fig. 1-2B).

Roesli (1946) postulated another boundary, from the Mesozoic rocks at Saas Grund towards the Tällihorn, east of the Weissmies, but he could not find Mesozoic rocks in the Weissmies itself. Roesli stated that the Portjengrat zone (in his terminology the unit between the Furgg zone and his new boundary) might be a more or less individual element between the Bernhard and Monte Rosa Nappes.

Bearth (1952a) proposed the name 'Monte Rosa-Bernhard Nappe' for the Mischabel Nappe, subdividing it in the Monte Rosa and Bernhard component nappes because of the different internal structure and metamorphic history of the two components. Bearth (1953, 1957a) understood the structural importance of the Furgg zone, extending from the Antrona Syncline to the Mesozoic rocks covering the Monte Rosa Nappe, and with some reserve, he gave it the status of the boundary between both component nappes. The unit between the Furgg zone and the Saas Fee–Zwischbergen boundary, Bearth called the 'Portjengratlappen' and later the Portjengrat zone (Index map II, sheet Saas, No. 31 of the 'Geologischer Atlas der Schweiz 1:25.000' and explanation, Bearth, 1957a).

In a previous paper Bearth (1956b) had already reintroduced the concept of individual Monte Rosa and Bernhard Nappes, with the Furgg zone separating them. He did not use Argand's boundary because he believed it to pass through coherent gneisses. One of his reasons for this reintroduction was that elsewhere in the Alps the Bernhard Nappe is separated from the internal massifs, Dora Maira and Gran Paradiso, the equivalents of the Monte Rosa Nappe.

In later years, Bearth (1964, 1967; Hunziker & Bearth, 1969) returned to the point of view of Argand.

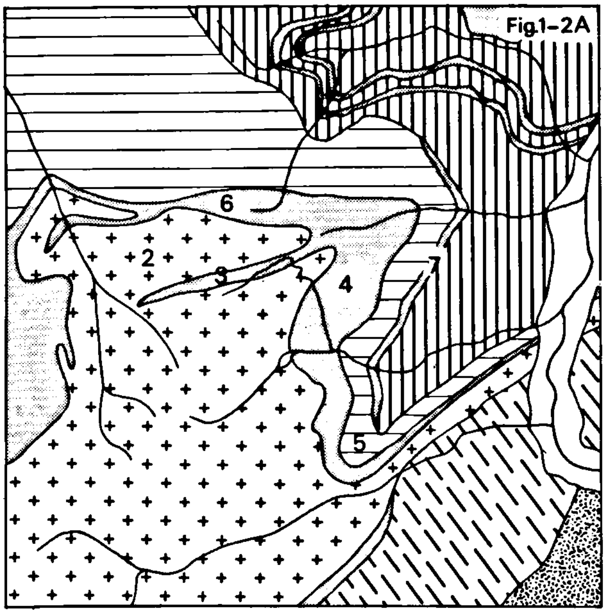
1.4.3. *The Antrona Syncline, the Camughera–Moncucco Complex and the Simplon–Centovalli Fault*

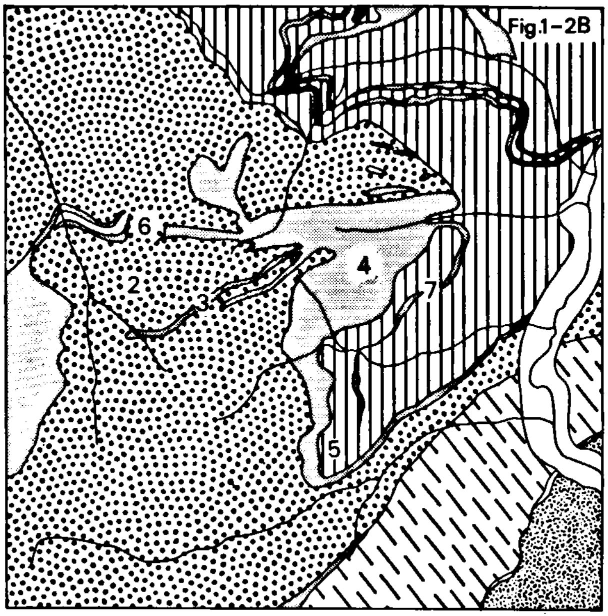
The Antrona Syncline and the Camughera–Moncucco Complex are featured on the 'Carta Geologica d'Italia Foglio 15, Domodossola' (1913). The geology was investigated by A. Stella between 1890 and 1910.

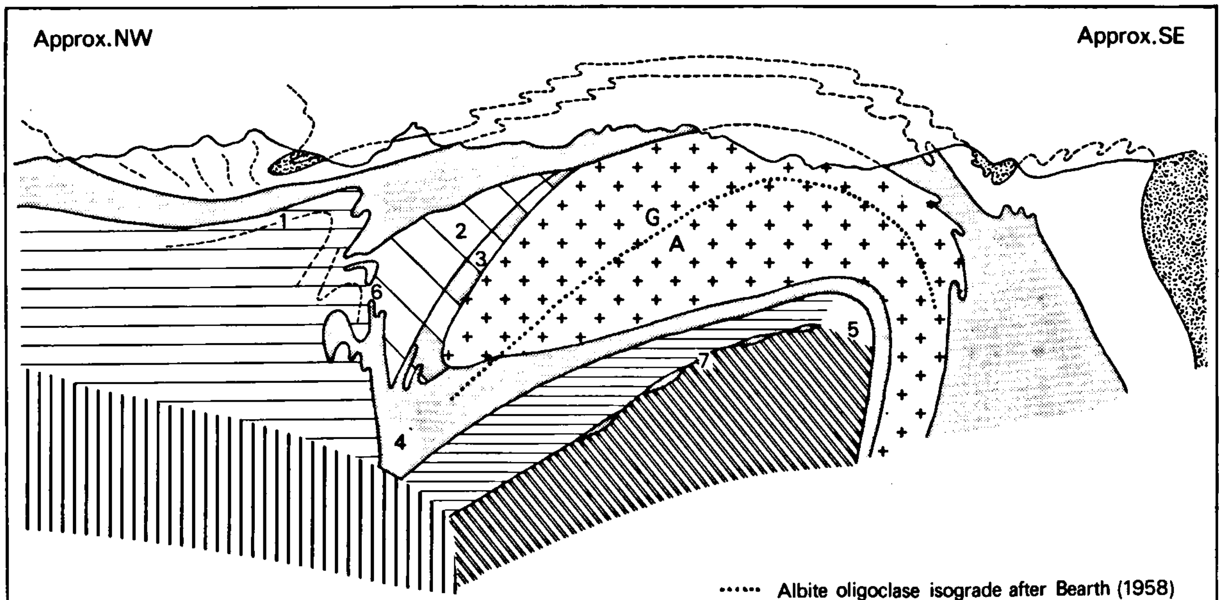
In the Camughera–Moncucco Complex, Bearth (1939) observed that near the Passo Salarioli, on the watershed between the Val Anzasca and the Val Antrona, the Camughera granite gneisses disappear and the Salarioli-'Mulde' terminates. No evidence was found for the 'root' of the Bernhard Nappe, another factor in Bearth's accepting of the Mischabel Nappe of Staub (1936, 1937).

Blumenthal (1952) studied the Antrona Syncline and the Camughera–Moncucco Complex, and in his opinion, the latter was part of the Monte Leone Nappe. Blumenthal's publication includes numerous structural sections featuring large folds, and the existence of some of them is confirmed in this paper.

While these authors generally adhere to the classic views of Argand, trying to improve them and to add detail, Amstutz (1950, 1954, 1971) holds a strongly different opinion on the structure of the Antrona Syncline and adjacent areas (Fig. 1-4). He believes that the amphibolites of the Antrona Syncline (which he interpretes as an anticline) should represent the autochthonous cover of the Monte Rosa Nappe, and have been overturned in a later stage of mountain building as the result of underthrusting of the 'Antrona







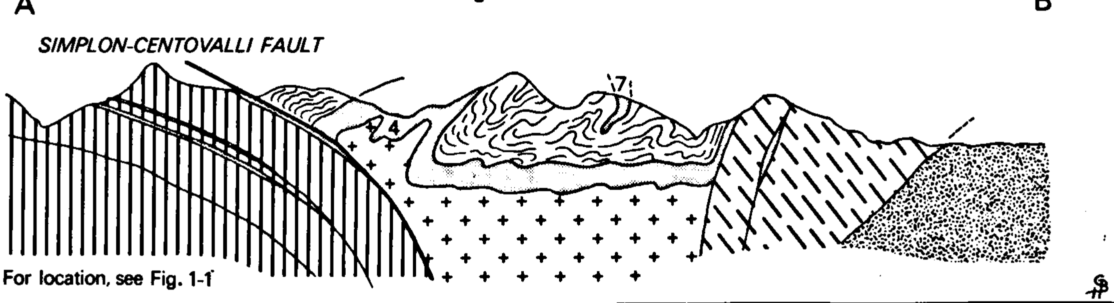
······ Albite oligoclase isograde after Bearth (1958)

G Greenschist metamorphic facies

A Amphibolite metamorphic facies

Fig.1-3
This section is schematical and not in one plane.

Fig. 1-4



For location, see Fig. 1-1

Fig. 1-2. A. The interpretation of the regional structural geology between the Saaser Vispa and the Toce, by Argand (1911).

B. Another interpretation of the regional structural geology of the same area, after Bearth (1939). Dots indicate the 'Mischabel Nappe'.
Compare with Fig. 1-1; see there for the legend. The scale is identical to the scale of Fig. 1-1. The numbers indicate the following structures: 2, Portjengrat zone; 3, Furgg zone; 4, Antrona Syncline; 5, Vanzone Antiform; 6, Saas Fee-Zwischbergen boundary; 7, Salarioli-'Mulde'.
Explanations in the text; see 1.4.1 to 1.4.4.

Fig. 1-3. A profile through the present area and its surroundings. The scale approximates the scale of Fig. 1-1; see there for the legend and the references. The number 1 indicates the Mischabelrückfalte; for the explanation of the other numbers, see either Fig. 1-1 or the legend to Fig. 1-2.

Fig. 1-4. Amstutz' (1954) profile from the Lepontine Nappes to the Ivrea–Verbano zone (see Fig. 1-1 for the profile trace and the legend, the scale of this section is approximately two times larger). The Antrona Syncline is interpreted as an anticline, and the Camughera–Moncucco zone (wavy lines) supposedly covers the Monte Rosa Nappe. The number 4 indicates the closure of the 'Antrona Anticline'; 7 indicates the Salarioli-'Mulde'. Further explanations in the text (see 1.4.3).

Anticline' by the Monte Leone gneisses. He connects the Bernhard Nappe and the Camughera zone as Argand (1911) did before, but by an anticline instead of a syncline, and he assigns the Camughera–Moncucco Complex to the Bernhard Nappe, separating it from the Monte Leone Nappe by a fault passing through the Val Bognanco from Alpe Arza (north of Bognanco San Lorenzo) to Cisore (near Domodossola). An important factor in Amstutz' (1954) interpretation is the so-called transversal fold of Antronapiana that could coincide with a large second generation fold (as defined in this paper) in the Camughera zone (see 5.4.1).

Though objecting against the majority of Amstutz' (1954) ideas, Bearth (1956a) agreed on the necessity to separate the Camughera–Moncucco Complex from the Monte Leone Nappe. A significant argument was the completely different internal structural geology of both units. Wenk (1955b) shared this view and pointed out that Schmidt & Preiswerk (1908) had made this distinction before.

Wenk (1955) describes a northern branch of the Insubric Line passing through the Centovalli and the Val Vigezzo (between Domodossola and Locarno), and he connects it to the fault in the Val Bognanco that separates the Camughera–Moncucco Complex from the Monte Leone Nappe. This is the Centovalli Fault (Knup, 1958), and together with its continuation in the Simplon area (Bearth 1956a, b) it is referred to as the Simplon–Centovalli Fault. The history of the movement along this fault, a branch of the Insubric Line, was reconstructed by Hunziker (1970) on the basis of large differences in radiometric Rb/Sr cooling ages of Alpine micas on either side of the fault (Hunziker, 1969, 1970; Hunziker & Bearth, 1969).

1.4.4. *The Mischabelrückfalte and its bearing on the present area*

Zwart (in prep.) demonstrates that in the Mischabelrückfalte area three generations of folds occur. The first generation folds (F_1), with the regional schistosity S_1 as their axial-plane schistosity, have been refolded by second generation folds (F_2) that possess subhorizontal to slightly northwest dipping axial planes, and whose axes (plunging west to northwest) are paralleled by the major lineation, L_2. The F_2 folds in the Mischabelrückfalte have the usual asymmetry relationships of parasitic folds to the hinge of an antiform, and Zwart (in prep.) considers the Mischabelrückfalte as a large F_2 fold. It postdates the emplacement of the nappes since it folds the nappe boundaries. The last generation of folds (F_3), with vertical or subvertical axial planes, and axial directions generally parallel to F_2, deforms the older structures but it is not of regional importance.

In the upper limb of the Mischabelrückfalte, the zone of Mischabel–Siviez is in normal stratigraphic succession covered by younger rocks (Bearth, 1962, 1964; see 1.2.2), but in its lower limb the zone of Mischabel–Siviez is overlying the Ophiolite zone of Zermatt–Saas Fee (Bearth, 1964). These limbs are called, respectively, the non-inverted and the inverted limb, and their position corresponds to the inverted and non-inverted limbs of the F_2 folds to be discussed later in this paper.

The inverted limb of the Mischabelrückfalte is exposed at the altitude of Saas Fee in the Saastal (Güller, 1947; Bearth, 1964; Zwart in prep.; this paper). The Mittaghorn Synform (see 5.2.1) brings this inverted limb into a non-inverted position, and the present work begins with that fold. It is likely to be the same as the 'Synclinal rétrograde de Saas' (Staub, 1936) which according to that author is closing in the Weissmies area.

The three folding phases recognized by Zwart (in prep.) occur also in the present area, and they correspond to the F_1, F_2 and F_3 of this paper.

1.5. STRATIGRAPHY AND ROCK DESCRIPTION

1.5.1. *Major rock units and the structural geology of the Pennine Nappes*

The oldest rocks are the Hercynian or possibly, older high-grade metamorphites and intrusive granites of the pre-Westphalian (Bocquet et al., 1972) basement that are also intensively affected by Alpine metamorphism. These now form the volumetrically largest part of the Pennine Nappes. This basement is covered by sedimentary rocks and mafic igneous rocks of Permocarboniferous age, some of which were transported as the nappe cover, and retain a (para-) autochthounous relation to the basement (Permocarboniferous cover of the zone of Mischabel–Siviez, Bearth, 1964), while others are contained within the nappe, and mark internal thrust planes (Furgg zone, enclosed by the Monte Rosa Nappe s.l.; Dal Piaz, 1964, 1966; Wetzel, 1972). The Permocarboniferous rocks postdate the high-grade metamorphism of the basement and have only been subjected to Alpine metamorphism. There is, however, some evidence for a Permian phase of metamorphism in the Monte Rosa Nappe s.l. (Hunziker, 1970) of strictly localized extent (Bocquet et al., 1972). In the Palaeozoic rocks, neither stratigraphical thicknesses nor a stratigraphical sequence could be established in the present area.

The youngest major rock unit is a sequence that begins with Permotriassic quartzitic rocks (Bearth, 1957a) and concludes with rocks of Upper Cretaceous age (Bearth, 1964); part of this sequence contains ophiolites. Its metamorphism is of Alpine age. Rocks belonging to this unit are sometimes contained within the nappes, forming the so-called 'Deckenscheider' (Preiswerk). Their stratigraphical sequence is summarized under 1.5.4.

1.5.2. *Basement rocks*

Zone of Mischabel–Siviez. – It is the polymetamorphic core of the Bernhard Nappe (Bearth, 1962), composed mainly of paragneisses whose sedimentary age is older than the Late Carboniferous (Bearth, 1962). It has been subjected to pre-Alpine metamorphism under amphibolite facies conditions (Bearth, 1962; Bocquet et al., 1972). Rock descriptions are to be found in Güller (1947), Jäckli (1950) and Bearth (1956a, 1957a, 1962, 1964). The Verosso gneisses have been described by Blumenthal (1952), Amstutz (1954) and Bearth (1956a). Along the contact with the Antrona amphibolites they contain granite gneisses frequently passing into granite with ill-defined schistosity planes.

The paragneisses belonging to the zone of Mischabel–Siviez are fine- to coarse-grained, grey to greyish-green rocks that weather to a brown or brownish-red colour. Where the albite content is high, the rocks have a paler shade of these colours. There is a gradual change in lithology on a scale of several metres to a few tens of metres, and a small-scale (cm–dm) layering that are both defined by different relative proportions of albite, phyllosilicates and quartz. The small-scale albite-rich layers are well-defined and have sharp contacts, also on a miscroscopic scale, but transitional contacts are not uncommon. Some albite-rich layers are formed by a concentration of albite porphyroblasts. Local epidote-amphibolite lenses of possible Mesozoic age occur in the Wysstal (Bearth, 1954), and in a presumed syn-nappe movement zone north of the Mittelrück, they contain eclogites that were metamorphosed to garnet amphibolites.

The mineral content is characteristic of greenschist facies rocks, with albite (sometimes inversely zoned oligoclase), quartz, colourless mica, chlorite and garnet. In the Almagellhorn and below the tongue of the Rotblattgletscher biotite-rich gneisses occur. Biotite is also common in the epidote-amphibolites; elsewhere, it occurs only occasionally.

Camughera–Moncucco Complex. – Bearth (1956a), although admitting that the Camughera and Moncucco zones are structurally and lithologically different, regards them as one unit, for they grade into each other in the area of the Passo di Salarioli. Laduron (1974) objects against this gradual transition and questions whether the Camughera and Moncucco zones should not be set apart in the sense of Argand, as different units separated by the Salarioli-'Mulde'. In the Val Brevettola section, the Camughera zone is lithologically and structurally reminiscent of the Portjengrat zone and the granite gneisses in the Verosso gneisses (Bernhard Nappe). The Moncucco zone, however, shows little affinity to other parts of the present area, and its continuation is found in the Lepontine Nappes (see 1.4.3). The Camughera gneisses (an informal name for the rocks of the Camughera zone) are believed to be part of the Bernhard Nappe, for they have the same structural position below the Antrona amphibolites (see 7.4).

Camughera zone. – It has been described by Blumenthal (1952), Amstutz (1954) and Bearth (1956a). Bocquet et al. (1972) consider it as a portion of the Bernhard Nappe, and consequently as a part of the pre-Westphalian basement.

In the present area, the Camughera gneisses consist mainly of light-coloured and greyish to reddish weathering coarse augengneisses. Microcline augen up to 8 cm long lie in a groundmass of oligoclase–andesine (Bearth, 1958), biotite, colourless mica and quartz. Laminated biotite-rich orthogneisses with a plagioclase-rich vs. quartz-rich layering and occasional augen are interlayered with the augengneisses. The orthogneisses wedge-out in the northeast where they are replaced by micaschists (Amstutz, 1954; Bearth, 1956a). Very light-coloured, fine-grained and unlayered orthogneisses adjoin the contacts with the Antrona amphibolites and the Salarioli-'Mulde' in the Val Brevettola; they consist chiefly of plagioclase, quartz with subordinate microcline and colourless mica. Close to the Passo del Pianino and in the R. di Molezzano, thin plagioclase- and quartz-rich layers alternate with paper-thin biotite-rich micaceous layers in phyllonitic augengneisses with many strongly elongate augen. Dark-brown weathering biotite schists and subordinate amphibolites are intercalated with the orthogneisses in the Camughera zone. They are similar to their counterparts from the Moncucco zone.

For the Salarioli-'Mulde', see 1.5.4.

Moncucco zone. – Bocquet et al. (1972) assign it to the Bernhard Nappe, which implies that the rocks of the Moncucco zone might be pre-Westphalian, but their age could be different when the Moncucco zone is related to the Lepontine Nappes instead (Bearth, 1956a).

A characterization of the rock-types in the Moncucco zone is as follows:

The micaschists and micaceous gneisses contain quartz, oligoclase–andesine, biotite, muscovite, ± epidote, ± garnet, and rarely staurolite and kyanite. The micaschists have alternating quartz-rich and mica-rich layers on a mm-scale, and some plagioclase-rich layers also occur. The micaceous gneisses have a layering of quartz-rich and plagioclase-rich lamellae, sometimes also thin mica-rich layers, but more commonly the mica (mostly biotite) is evenly distributed. Most rocks are fine- to medium-grained.

The fine- to medium-grained orthogneisses and augengneisses in the southeastern slopes of the Moncucco and along the Salarioli-'Mulde' are similar to the orthogneisses of the Camughera zone. The gneiss contained between both levels of ultramafic rock consists of andesine, microcline, biotite and muscovite. Biotite is the predominant mica. Microcline augen frequently consist of an aggregate of small grains rather than a single porphyroclast. Thin mica-rich layers are numerous.

The amphibolites and hornblende gneisses contain andesine-labradorite, hornblende, epidote, ± zoisite, ± biotite, ± muscovite, ± chlorite, ± garnet and ± quartz. They are fine-grained rocks with a conspicuous layering on mm-dm scale of alternating hornblende-rich and epidote-rich layers, or hornblende-rich and plagioclase-rich layers.

Monte Rosa Nappe s.l. – Bearth (1952a) distinguishes between an Alpine and a pre-Alpine phase of metamorphism, separated in time by granite intrusions.

The pre-Alpine phase of metamorphism, which is of Hercynian or Caledonian age (Bocquet et al., 1972), took place under high-grade conditions. Sillimanite (replaced by Alpine kyanite, Bearth, 1952a) is frequent in the metapelites, and local migmatites also occur. The metasediments were intruded by granites of Carboniferous age (Bearth, 1952a), and this has been confirmed by Hunziker (1970) who dated the time of crystallization of the granite with a 310 ±50 m.y. Rb/Sr whole rock isochron in weakly deformed Monte Rosa granite. Hunziker (1970) detected a whole rock Rb/Sr age of 260 ± 10 m.y. in gneissic Monte Rosa granite ('Augengneiss von Macugnaga'), and Rb/Sr ages of approx. 250 m.y. in muscovites oriented in the foliation. He uses these ages in favour of a weak phase of metamorphism of Permian age.

Portjengrat zone. – It is the most strongly Alpine-deformed portion of the Monte Rosa Nappe, and few traces have remained of its pre-Alpine history. The paragneisses of the Portjengrat zone are in all respects similar to the gneisses of the zone of Mischabel–Siviez (see also Bearth, 1957).

The orthogneisses of the Portjengrat zone have been described by Bearth (1948, 1957a). They occur either as large bodies or are interlayered with the paragneisses, and inclusions of orthogneiss in paragneiss or vice versa are found on every scale. The contacts are sharp, or gradational over a distance between a few cm and a few m. Some layers of orthogneiss enclosed in the paragneiss are F_1 folds (see 2.3); others are very flat lenses. When fresh the orthogneisses are grey to white. A phengite content gives them a greenish hue, but weathering of the phengite changes its colour to a rusty red, and frequently causes the orthogneisses to show a red tinge in the field. They are albite–microcline–quartz–phengite/muscovite gneisses with a compositional layering formed by variable relative proportions of these minerals. This layering occurs on any scale between a few mm and several tens of metres. Layers of aplitic composition occur with thicknesses between a few cm and a few dm. The rocks are usually fine- to medium-grained and contain isolated microcline porphyroclasts up to 5 cm long. Medium–coarse grained augengneisses are frequent and occur in layers up to a few tens of metres thick. A larger body of augengneisses is exposed south and west of the Mittelrück and it is rich in biotite.

Layers of micaschists of up to several metres thickness, intercalated with the orthogneisses, and consisting almost exclusively of colourless mica, result from early Alpine transformation of alkali feldspar-rich rocks (Bearth, 1952a, 1953; Reinhardt, 1966).

1.5.3. *Palaeozoic cover rocks*

Furgg zone. – It is the Palaeozoic cover of the Monte Rosa Nappe (Bearth, 1953; Wetzel, 1972). Wetzel (1972) believes it to be of Lower Carboniferous age, Bocquet et al. (1972) consider it as pre-Westphalian. Wetzel (1972) gives a petrographical description of the Furgg zone, and published a monograph on chlorites from this zone (Wetzel, 1973). Dal Piaz (1964, 1966) describes Furgg zone-type rocks in the southern part of the Monte Rosa Nappe s.s.

On the basis of small compositional differences, the shape of the rock bodies, and their association with other rocks, Wetzel (1972) distinguishes between carbonaceous rocks and amphibolites of Palaeozoic and Mesozoic age. Their exact distribution is not described and they are not discriminated in this paper.

A very heterogeneous lithology is characteristic for the Furgg zone (Bearth, 1953; Wetzel, 1972) but in its eastern part, paragneisses with intercalated amphibolites predominate. The former resemble the rocks from the zone of Mischabel–Siviez, but biotite is common, and the plagioclase is albite, surrounded by an oligoclase rim (Wetzel, 1972; this paper). Both rock-types are associated with Mesozoic and older (Wetzel, 1972) metasediments together with serpentinites.

1.5.4. *Mesozoic cover rocks*

Only quartz-rich and carbonate-rich Mesozoic metasediments are distinguished here (see also Enclosures 1 and 2) and they are not differentiated according to age. Stratigraphical thicknesses are unknown in the present area.

The stratigraphical sequence, as described by Bearth (1957a), is as follows:

1. (Permo) triassic quartzitic and associated rocks. They are muscovite quartzites and muscovite–albite gneisses, that according to Bearth (op. cit.) could represent Permian sediments, upon which in Lower Triassic time quartz-rich rocks were deposited that are now phengite-bearing, light-green platy quartzites. In the present area, these quartzites occur at the Zwischbergenpass, in the Furgg zone, and in the Mittaghorn where their large thickness is structurally controlled. They were not differentiated in the Antrona Syncline.

2. Triassic marbles and dolomites. They have Middle Triassic age, and occur (Geol. Atlas der Schweiz, Blatt Saas, 1954) in the Saas area and in the Furgg zone. Blumenthal (1952) observed them in the Antrona Syncline. They were mapped as Mesozoic carbonaceous rocks.

3. Jurassic calcareous schists ('Bündnerschiefer'; 'Schistes lustrés'). These rocks too were mapped as Mesozoic carbonaceous rocks, and they are grey to bluish-grey calcareous schists with brown weathering colours, that may change into bluish marbles with a low content of non-carbonaceous minerals. See Bearth (1957a, 1964) and Wetzel (1972) for detailed descriptions of these metasediments.

4. Ophiolites. Dal Piaz et al. (1972) attribute a Jurassic age to the ophiolites of the zone of Zermatt–Saas. Only a very small portion of this zone is exposed in the present area, where it consists mainly of prasinites (Bearth, 1957, 1967).

It is probable that the ophiolites which are now the Antrona amphibolites have approximately the same age, though Blumenthal (1952) thought that the oldest could be of Triassic age. The Antrona amphibolites have not been thoroughly petrographically investigated, but some particulars were published by Blumenthal (1952), Bearth (1967) and Wetzel (1972). The most common rock-types in the Antrona amphibolites are epidote and zoisite amphibolites with or without plagioclase (oligoclase–labradorite), and amphibolites consisting almost exclusively of plagioclase and amphibole. The amphibole is usually a blue-green Barroisitic hornblende (Bearth, 1967) but other amphiboles occur too (see 6.10). Minor constituents are garnet, chlorite, and rarely biotite and/or colourless mica. They are medium- to fine-grained rocks.

A well-developed layering, on a mm–cm scale, is formed by hornblende-rich and epidote-rich or zoisite-rich layers. Plagioclase-rich layers alternating with hornblende-rich layers are infrequent. The Antrona amphibolites contain serpentinite bodies, commonly with faulted margins, or as cm to dm thick layers parallel to the layering of the amphibolites. Inclusions of eclogite showing varying degrees of metamorphic change were occasionally found, and a large body containing glaucophane, is exposed at the Alpe Pasquale southeast of the huts along the track to Bocchetta Pasquale.

5. The Mesozoic(?) metasediments of the Salarioli-'Mulde'. Argand (1911, 1916), Blumenthal (1952) and Bearth (1939, 1956a) attribute Triassic age to the calcareous rocks characterizing the Salarioli-'Mulde', and Bearth (1939) assigns this age also to the quartz–mica schists that at the Passo di Ogaggia are associated with them.

The Salarioli-'Mulde' is, however, nowhere visibly connected to rocks of established Mesozoic age, and its calcareous rocks are not associated with amphibolites (see also Blumenthal, 1952). This is unexpected, since the Antrona amphibolites are the dominating Mesozoic rock-type around the Camughera–Moncucco Complex. Stella (Carta Geologica d'Italia Foglio 15, Domodossola) mapped the calcareous rocks as pre-Triassic marbles, and Bearth (1956a) reports occurrences of calcareous rocks (to which he attributes Mesozoic age) in the Moncucco zone itself.

For these reasons, the Mesozoic age of the calcareous rocks in the Salarioli-'Mulde' is open to doubt and they may belong to the pre-Westphalian basement.

CHAPTER 2

FIRST GENERATION STRUCTURES

2.1. GROUPING OF THE FIRST GENERATION STRUCTURES

The style of the F_1 folds is identical in all rock-types, and they possess an almost omnipresent axial-plane schistosity S_1. They are always folds of the compositional layering, and in the Permotriassic and Mesozoic metasediments no evidence could be found suggesting that this compositional layering might be a differentiated layering of structural origin. This layering is therefore believed to be sedimentary bedding, which confirms that the F_1 folds were the first to form in these rocks. Apart from the bedding itself, no relict sedimentary structures were found.

The presence of the S_1 schistosity in the Mesozoic rocks is proof for its Alpine age. S_1 in the Mesozoic rocks is everywhere parallel to S_1 in the basement rocks of the nappes, when some local effects of post-F_1 deformation are disregarded. This and the similarity in the style of the associated folds demonstrate that S_1 is in the same plane in the basement and in the cover rocks. The Alpine age of S_1 in the Portjengrat zone, the zone of Mischabel–Siviez and the Stelli zone is confirmed by Bearth (1952a, 1957a, 1958, 1962, 1964); and an Alpine age of S_1 in the Camughera–Moncucco Complex is implied by Bearth (1956a), Blumenthal (1952), and confirmed by Laduron (1974). The fact that S_1 in the nappes has been formed during Alpine time does not imply that all minerals defining the S_1 planes are necessarily of Alpine age (Hunziker, 1969, 1970; see also Ferrara et al., 1962).

Dal Piaz et al. (1972) present a structural and metamorphic evolution scheme of this part of the Alps, largely based on the Sesia–Lanzo zone. A significant part of their structural history precedes the formation of S_1 in the present area (during the last stage of nappe emplacement, see 7.6), and older structures might therefore have existed. Pre-F_1 structures are, however, extremely rare (see 2.2), and the F_1 folds therefore represent the first Alpine-fold generation that was observed in the present area.

2.2. STRUCTURES PREDATING F_1

Folds. – In the Val Brevettola, adjoining the southern contact of the main body of the ultramafic rocks, a close F_1 fold has refolded an isoclinal fold in the amphibolites (Fig. 2-1). Alpine age of the refolded folds could not be demonstrated positively, and the same applies to fold closures contained by layers of albite gneiss in the paragneisses of the Steintälli (Almagellertal) that predate F_1.

A Type 3 (Ramsay, 1967) interference pattern in Antrona amphibolites at the Alpe Pasquale has been formed by an F_1 fold and an earlier fold of Alpine age. They are both folds of the layering, and can only be distinguished when refolding each other, or when F_1 has an axial-plane schistosity. S_1 is frequently absent in the Antrona amphibolites, and as a consequence some uncertainty remains about the age of early folds there, although nearly all of them are believed to have F_1 age.

Schistosity planes. – A few phengites from the core of an F_1 fold in the Almagellertal are not oriented in S_1, and though this could be interpreted as a folded older schistosity plane, it seems more likely that F_1 deformation continued when S_1 was already in existence.

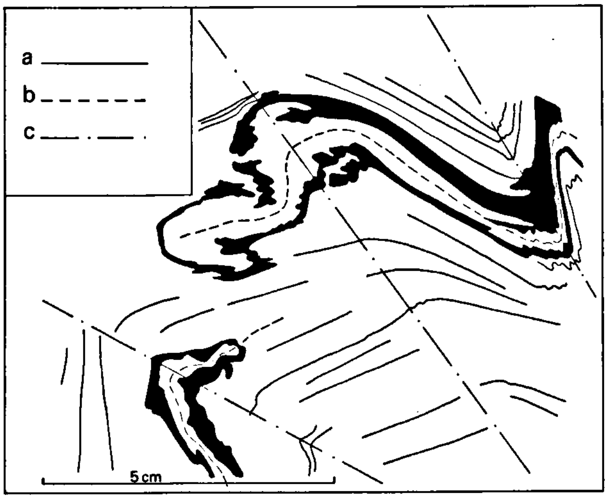


Fig. 2-1. F_1 folds refolding isoclinal folds. (a) Trace of the compositional layering. (b) Axial plane trace of the refolded folds. (c) Axial plane trace of the F_1 folds.
Amphibolite, Val Brevettola; pegmatitic layers are black.

Compositional layering. – Not all of the layering has been formed at the same time. In the axial planes of F_1 folds in suitable lithologies, thin mica-rich layers alternate with quartz-rich layers ('Zeilenbau'), and this layering is of syn-F_1 age, but the remainder of the compositional layering predates F_1 and could be of early Alpine or Palaeozoic age. No relic sedimentary structures were observed and the layering is likely to have a metamorphic origin. The compositional layering in the orthogneisses could be of early Alpine age (Callegari et al., 1969; see also 7.5), and part of this layering has been derived from compositional differences in the igneous rocks (Bearth, 1952a).

2.3. PROPERTIES OF F_1 FOLDS

Fold style. – Isoclinal and rootless folds, many of them intrafolial, are prevalent, but close-tight F_1 folds of similar style also occur (Fig. 2-2). In the Furgg zone, where paragneisses and amphibolites are interlayered on a scale between a few cm and several m, overall similar fold style is maintained by differences in the radius of curvature of closures in adjoining layers. This is strongly reminiscent of a sequence of Class 1C–Class 3 folds (Ramsay, 1967), and good examples of it occur in the Passo della Preja area. F_1 folds in the orthogneisses of the Portjengrat zone, and in the very albite-rich gneisses in particular, are frequently less tight than their counterparts from the paragneisses.

The F_1 folds occur as isolated closures, isolated asymmetric folds, or sequences of closures in a single or a few folded layers (Fig. 2-3). Belts of continuous F_1 folding are prominent in the paragneisses and amphibolites of the Moncucco zone, and they seem to be important in the Monte Rosa zone also (see below). F_1 folds are difficult to find when the layering is faintly developed, and this can strongly influence the field impression of the intensity fo F_1 folding.

The Mesozoic rocks do not seem to contain more F_1 folds than the basement rocks.

Fold sizes. – Rootless folds usually have sizes between a few cm and a few dm (measured in S_1); they can, however, be much larger. The paragneisses in the eastern face of the Mittaghorn enclose rootless folds of orthogneisses with sizes up to several tens of metres, and a rootless F_1 fold in calcareous schists in the Monte della Preja has an exposed limb-length of approx. 50 m (Enclosure 2).

The isolated tight and isoclinal folds have long limbs of unknown but presumably great length when compared with the short limbs, which in most folds do not exceed a few dm but may be quite significantly larger. The short limbs occurring in the more thickly layered rocks tend to be longer than elsewhere. F_1 folds in the Moncucco zone

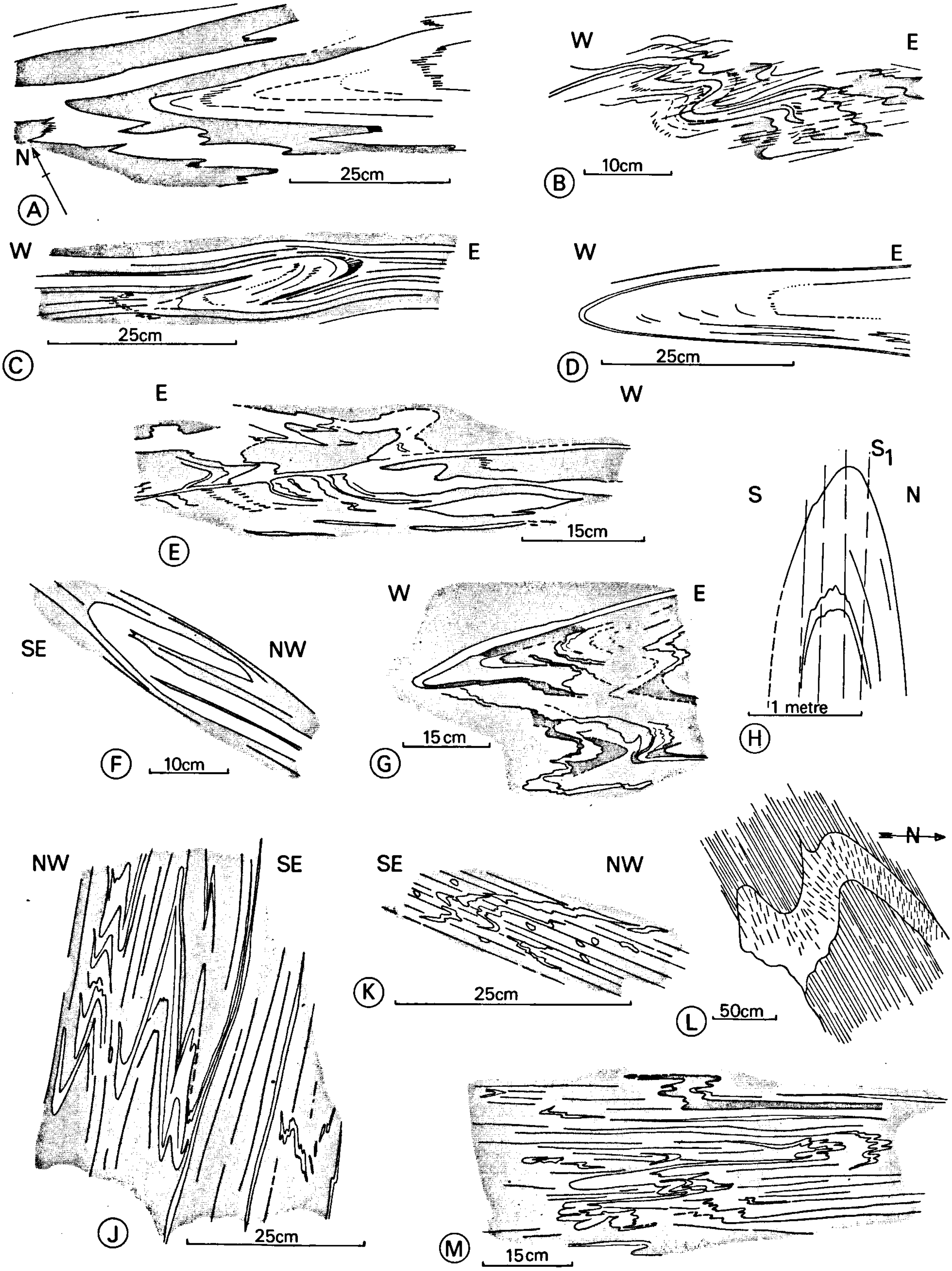

N
25cm
A
W
E
10cm
B
W
E
25cm
C
W
E
25cm
D
E
W
15cm
E
S₁
S
N
1 metre
H
SE
NW
10cm
F
W
E
15 cm
G
NW
SE
25cm
J
SE
NW
25cm
K
N
50cm
L
15cm
M

Fig. 2-2. A collection of F_1 fold profiles, selected to show their variability. All drawings were copied from photographs with the exception of L.
A: Orthogneiss, Portjengrat zone, Rotblattgletscher.
B: Paragneiss, Bernhard Nappe, Triftgletscher.
C: Orthogneiss, Portjengrat zone, Almagellertal.
D: Orthogneiss, Portjengrat zone, Rotblattgletscher.
E: Paragneiss, south of the Stelli zone, Monte Rosa Nappe.
F: Paragneiss, Furgg zone, Alpe del Gabbio.
G: Epidote amphibolite, Antrona amphibolites, Pizzo Montalto.
H: Orthogneiss, Portjengrat zone, Rottal.
J: Amphibole gneiss, Moncucco zone, Valemiola.
K: Orthogneiss, Moncucco zone, N of Villadossola.
L: Pegmatite with S_1 cleavage fan in biotite gneiss, Moncucco zone, Domodossola–Calvario.
M: Biotite gneiss, Moncucco zone, N of Villadossola.

are present as several orders of parasitic folds with short limbs ranging between a few cm and several m, but hinge 34 (see 5.4.3) is of much larger size. An F_1 fold in the Monte Rosa zone has a flat hinge of approx. 5 m width between almost parallel limbs, but here only one folded layer was observed, and its shape may not be representative. In this hinge a very strong L_1 lineation occurs, and on a larger scale this lineation is found in zones of approx. 200 m width, separated by wider zones of slightly linear rock with a well-developed S_1. These zones of linear rocks might be closures of F_1 folds.

In the calcareous schists between Saas Grund and Saas Fee, and in interlayered dolomite and prasinite of the eastern face of the Mittaghorn (layers between 2 and 5 dm thick), the limbs of F_1 folds are exposed for several metres. Folds of this size occur at the Zwischbergenpass, in the Furgg zone in the Furggtal, and at the Alpe Pasquale.

2.4. CHARACTERISTICS OF S_1

S_1 is formed by the planar preferred orientation of (001) of phyllosilicates, and/or by the planar preferred orientation of the c-axes of amphibole. Among the phyllosilicates are several kinds of colourless mica (see 6.4.) but the relations between the individual species and S_1 were not investigated.

The penetrativity of S_1 depends on the rock composition, and it is increased with an augmenting content of mica, or quartz and mica, whereas an increase in feldspar content makes S_1 less penetrative. The preferred orientation of amphibole does not always give the rock a schistose appearance in the field.

Basement rocks, general. – Where colourless mica and/or biotite occur in quartz-rich rocks, they are usually arranged in thin layers, normally of one mica thickness and forming a minute compositional layering ('Zeilenbau') that defines S_1. The micas impose (Wilson, pers. comm.) their grain boundaries on the quartz grains, and this frequently results in a dimensionally preferred

Fig. 2-3. F_1 folds in orthogneisses, Portjengrat zone, Almagellertal.

orientation of quarz in S_1. In the Camughera–Moncucco Complex, plagioclase-rich layers too occur between these mica-rich layers, and the boundaries between plagioclase and mica are the (001) planes of mica. Most of the boundaries between amphibole and plagioclase grains are formed by faces of the forms {110} and {010} of amphibole.

Paragneisses. Portjengrat zone and the zone of Mischabel–Siviez. – S_1 is defined by colourless mica and colourless and/or green chlorite; S_1 in lenses of (Mesozoic?) epidote-amphibolite (Wysstal, Mälliga-gletscher) results from the well-defined preferred orientation of amphibole, alone or in combination with biotite or chlorite having an imperfect preferred orientation.

The habit of the colourless micas and the S_1 planes is related to the lithology in which they occur (compare with the orthogneisses). In the quartz-rich and mica-rich paragneisses of the Portjengrat zone, the S_1 planes are laterally extensive, and the preferred orientation of the micas is good. The sizes of the micas range from 0.1 to 2.0 mm but are usually between 0.4 and 0.8 mm. The micas show evidence of boudinage, they have tapered ends, and their length/thickness ratios are up to 8.6 (measured perpendicular to L_2).

The paragneisses enclose numerous ellipsoidal porphyroblasts of albite that are orientated with their longest axes in S_1. This orientation probably results from the preferred orientation of colourless mica and chlorite at whose expense the albites have grown (Bearth, 1952a, 1953, 1957a; and 6.2.2), for in rocks where a phyllosilicate-preferred orientation is absent, equi-dimensional albite porphyroblasts occur exclusively. Another example of the preferred orientation of albite in S_1 is the replacement of a 'Zeilenbau' microstructure by skeletal porphyroblasts of albite enclosing the quartz grains between the former 'Zeilen'. These albites also occur in orthogneisses with a 'Zeilenbau', where they measure up to 3.5 mm in S_1. The porphyroblastic albites may be more than 10 mm long. Individual albites in layers consisting almost exclusively of albite may have a slight dimensional orientation that is, however, insufficient to define S_1. Their sizes range between 0.1 and 1.5 mm.

Orthogneisses. Portjengrat zone. – Colourless mica is the only phyllosilicate, apart from the augengneisses near the Mittelrück (see 1.5.2) where biotite and retrogressively formed chlorite are important.

The S_1 planes, of limited lateral extent, in the albite-rich (up to 70%) orthogneisses have been formed by a small number of micas with a poor preferred orientation, in some cases deviating up to 30° from the overall orientation of S_1. The thickness of such an S_1 plane is one or two micas, and it cannot be traced for more than a few cm; where it terminates it is replaced by others in the vicinity. The planes may be closely or widely spaced. In a gneiss with a feldspar/quartz/mica ratio of approx. 70/15/15, phengites (see 6.4) with an average size of 0.4 mm and a length/thickness ratio of 2.2. (measured perpendicular to L_2) have a ragged appearance. The albites are roundish or irregularly shaped grains with sizes between 0.5 and 2.2 mm, but the larger sizes are infrequent. Some albites show a polygonal fabric with triple junctions between the grain boundaries (Wilson, pers. comm.). Quartz occurs as irregularly shaped grains between the albites, and as lenses and irregularly shaped bodies, comprising a number of grains, parallel to S_1 or L_2. When feldspar, quartz and mica occur in approximately equal amounts, the dimensional preferred orientation of all minerals improves and the length/thickness ratios of the micas increases. When the albite content is less than approx. 30%, 'Zeilenbau' is frequent, and the micas have length/thickness ratios of up to 8.6. They measure max. 0.8 mm and are frayed or have tapered ends.

The augengneisses near the Mittelrück are characterized by S_1 planes, defined by colourless micas, curving around microcline porphyroclasts that are oriented with their longest axis approximately in S_1. Biotite and plagioclase do not show a well-defined preferred orientation. Apart from the porphyroclasts, microcline has only a sporadic dimensional orientation in S_1.

Paragneisses and orthogneisses. Camughera–Moncucco Complex. – S_1 is formed by colourless mica, biotite and occasional chlorite. In the amphibole-bearing gneisses and amphibolites, amphibole defines S_1 together with biotite and (retrogressive) chlorite in subordinate amounts. The planar preferred orientation of the phyllosilicates is better than in the Portjengrat zone, and in the Moncucco zone especially it may be almost perfect. Biotite dominates over, and has a better preferred orientation than colourless mica: this was previously observed by Ferrara et al. (1962). The micas are rectangular with or without frayed ends, and of smaller size when compared with the Portjengrat zone.

It frequently occurs that layers of quartz of a single grain thickness are confined between plagioclase–mica layers. They consist of quartz grains with a strong dimensional orientation parallel to the layering, and rectangular outlines, or more or less isolated flat ellipsoid grains when the layering is less well-developed. The grain boundaries of quartz are irregular because of grain boundary migration, but they are generally oriented at high angles to the layering. When the layering makes an angle with S_1, the quartz grains retain their preferred orientation parallel to the layering, and their formation is therefore not related to S_1 but depends on the presence of the thin layering. It is believed that this microstructure is formed by growth of quartz grains up to a size larger than the thickness of the layering by which they are contained (see also Spry, 1969) and they thus obtain a dimensionally preferred orientation in the plane of the layering.

Many plagioclase grains in the Moncucco zone are small, equant grains of uniform size with a strong tendency to form a polygonal microstructure, as numerous triple junctions between the grain boundaries demonstrate (Spry, 1969). They do not contribute to the delineation of S_1.

The microstructure of the Camughera augengneisses resembles the microstructure of the augengneisses near the Mittelrück, but there is a much better dimensional orientation of quartz and plagioclase, especially where the Camughera augengneisses are thinly layered. The microstructure of massive plagioclase-rich orthogneisses, occurring in the Val Brevettola adjoining the contacts with the Antrona amphibolites and the Salarioli-'Mulde', resembles that of the albite-rich orthogneisses of the Portjengrat zone.

Furgg zone. – Wetzel (1972, 1973) gives a detailed account of the microstructures in the various rock-types of the Furgg zone, and he describes the phyllosilicates that occur there. The present author's observations in the eastern part of the Furgg zone do not show much difference between the basement rocks in the Portjengrat zone, and the Mesozoic metasediments and amphibolites in the Antrona Syncline (see below). Compared with the Portjengrat zone, biotite is more frequently well-oriented in S_1. Plagioclases with a dimensional orientation in S_1, and skeletal porphyroblasts that replaced a quartz–mica 'Zeilenbau', are also more frequent.

Mesozoic rocks. – Colourless mica, colourless and green chlorite forms S_1 in both the Permotriassic quartzites and the Mesozoic calcareous rocks.

The colourless micas (phengites, Bearth, 1967) have different grain sizes in the mica-rich and in the quartz-rich layers of the Permotriassic quartzites. In the mica-rich layers (approx. 60% mica) of a specimen from the Mittaghorn, the average length of the micas (measured parallel to (001) and perpendicular to L_2) is 0.6 mm, whereas in the mica-poor layers (approx. 25% mica) it is 0.3 mm. The length/thickness ratio of the former is 2.7 and the latter's ratio is 3.8. A second specimen from the Mittaghorn gives an average mica length of 1.3 mm in the mica-rich layer, and 0.2 mm in the mica-poor layer (80% vs. 20% mica). When they are completely surrounded by quartz, all micas have tapered ends, and within any one layer they tend to be of the same size and to be uniformly distributed. In another specimen from the Mittaghorn, that contains tight F_2 folds, the largest micas are also situated in the mica-rich layers; the average length of the micas in the mica-rich layers (approx. 80–90% mica) is 0.8 mm, and in the mica-poor layers (approx. 40% mica) it is 0.4 mm. In all these specimens, quartz subgrains of F_2 age occur in older, larger grains. The older grains sometimes have a dimensional preferred orientation in S_1 and they are likely to be of F_1 age.

S_1 in the calcareous metasediments may be inconspicuous because phyllosilicate minerals are absent,

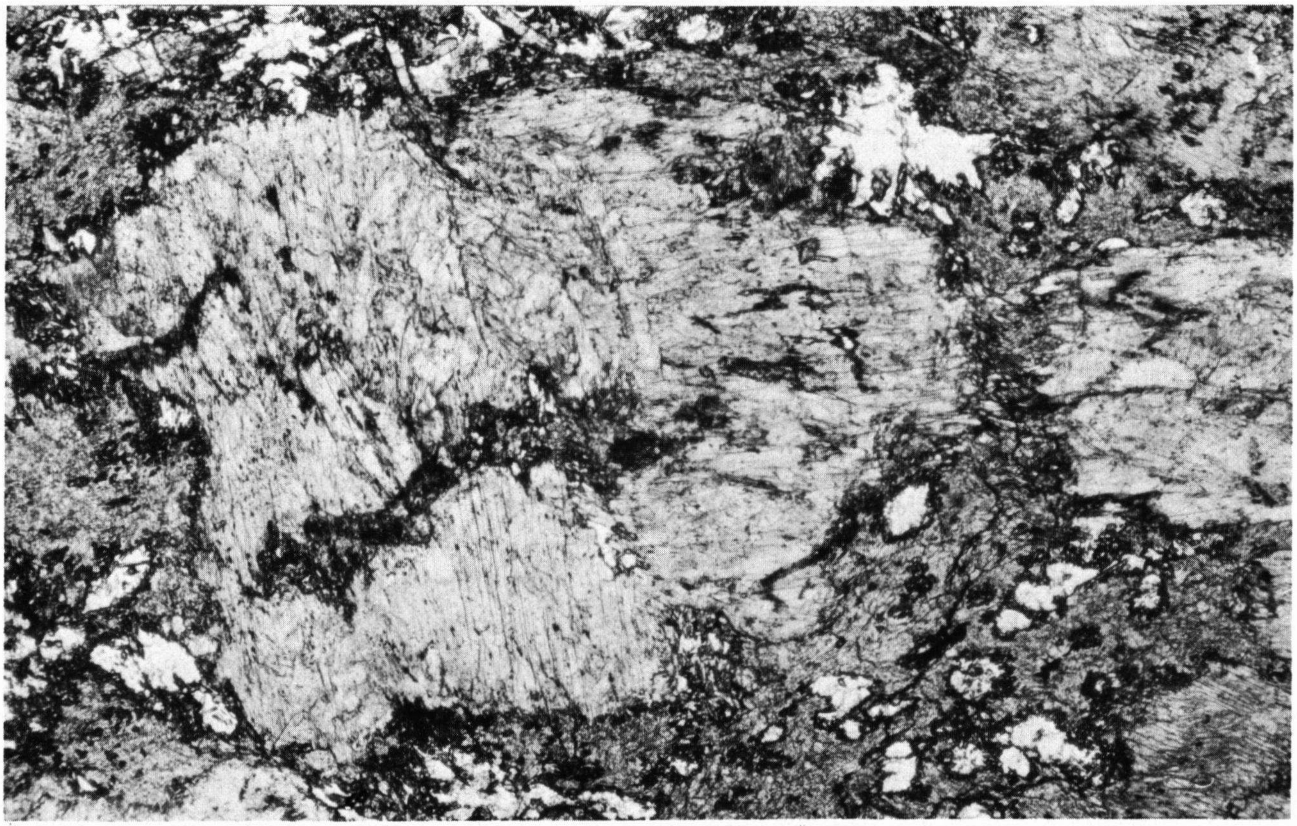


Fig. 2-4. Pseudomorphs of actinolitic amphibole after pyroxene, enclosed in a groundmass of zoisite and subordinate plagioclase. Zoisite amphibolite, Antrona Syncline. 10×, parallel polarizers.

and in a few cases S_1 is defined by a planar or linear preferred orientation of the c-axes of amphibole (see 2.5). The carbonate minerals sometimes form a polygonal microstructure; when this is absent, individual crystals show a slight dimensional orientation in S_1.

Antrona amphibolites. – S_1 is usually defined by the planar or linear preferred orientation of the c-axes of amphibole. Figs. 2-4 to 2-6 show a mode of development of S_1, out of non-foliated zoisite amphibolites. The zoisite amphibolite contains numerous porphyroblasts of actinolitic amphibole up to 7 mm long which could be pseudomorphs after igneous pyroxene (Bearth, 1967; Dal Piaz, pers. comm.). Their distribution suggests gradational igneous layering, and they are embedded in a fine-grained zoisite groundmass. In such a rock, S_1 appears as a 'Schlieren' microstructure of irregularly shaped, elongate domains of zoisite with subordinate (and probably later) plagioclase alternating with amphibole-rich domains. The latter are characterized by a mortar structure (Spry, 1969) of a second generation amphibole growing at the expense of the older porphyroblasts. The new amphiboles are oriented with their c-axes in S_1, whereas the old amphiboles have been rotated and deformed, the newly formed S_1 planes curving around them. The second generation of amphibole remains fine-grained.

Frequent massive amphibolites consist of a very fine-grained intergrowth of blue-green hornblende and plagioclase, without any dimensional orientation of either mineral, and consequently no S_1. It is likely that this microstructure postdates F_1, and was formed at the expense of earlier minerals. Sometimes S_1 is defined by elongate domains of plagioclase and chlorite, but the chlorite does not always possess a preferred orientation.

2.5. THE L_1 LINEATION

In all rock-types, L_1 is formed by the intersection of S_1 and the compositional layering. In the para- and orthogneisses of the Camughera–Moncucco Complex it is well-defined as alternating feldspar-rich and mica-rich bands in S_1. The width of these bands, measuring up to 15 cm, is dependent on the angle between S_1 and the layering. The frequency of L_1 in the Camughera–Moncucco Complex results from the large number of F_1 folds which are not always very tight. A microscopical L_1 in the orthogneisses of the Moncucco zone is formed by a dimensional orientation of biotite, quartz and plagioclase. This lineation may occur also in the paragneisses of the Moncucco zone, but a linear preferred orientation of amphibole c-axes is more frequent there.

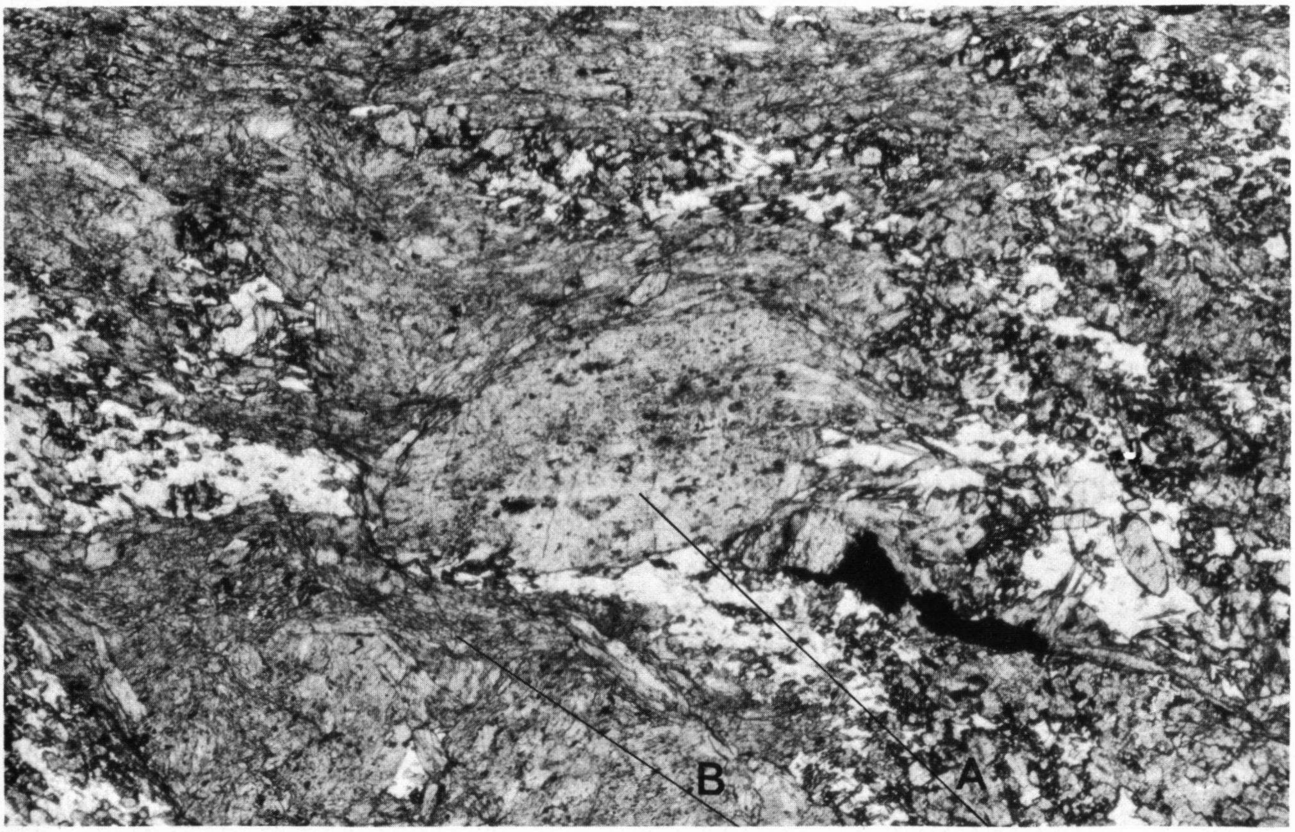


Fig. 2-5. The same rock type as in Fig. 2-4 after development of S_1. Relics of the rotated and deformed pyroxene pseudomorphs (A) are enclosed and replaced by a new generation of amphibole (B) that grew syn-F_1. The plagioclase content has increased. 10×, parallel polarizers.

Apart from the intersection between the compositional layering and S_1, these lineations have not been found in the Portjengrat zone. Because most F_1 folds in the Portjengrat zone are isoclinal or rootless, the intersection lineation is scarce, and it is found only in the hinge zones of a few close-tight F_1 folds. The width of the L_1 bands in S_1 may be up to 30 cm (Fig. 3-1A).

In a dolomite at the Alpe Pasquale (Bocchetta Pasquale), the L_1 lineation is a linear preferred orientation of grammatites (see also Wetzel, 1972, fig. 24).

The very strong L_1 lineation in the Monte Rosa zone (see also Reinhardt, 1966) is caused by pencil-shaped domains of plagioclase and/or quartz, very elongate inclusions of biotite schist, and locally the intersection between the layering and S_1.

2.6. BOUDINAGE

The effects of boudinage are wide-spread in the basement rocks, and particularly evident in the interlayered para- and orthogneisses of the Portjengrat zone, where the orthogneisses occur as lenses, boudins and boudinaged folds (see 2.3). Internal boudinage has been observed in the orthogneisses in the Almagellertal, and the axes of the boudins are approximately parallel to the F_1 fold axes, pointing to a syn-F_1 age. Boudinage is not so evident in the Camughera zone but it is more wide-spread in the Moncucco zone, where it affects the amphibolites.

The Furgg zone is known to contain strongly boudinaged rocks (Bearth, 1953; Wetzel, 1972), and in the portion investigated, boudinage is thought to be contemporaneous with F_1 deformation. Some of the lenses of Mesozoic or older (see 1.5.3) calcareous metasediments in the eastern part of the Furgg zone appear to be boudinaged F_1 folds, for they contain more or less distinct F_1 folds that frequently describe a complete closure. The best example is the large lens in the Monte della Preja (see 2.3 and Enclosure 2).

Elsewhere, the Mesozoic metasediments contain abundant evidence of having been boudinaged contemporaneously with F_1. Rotated boudins sometimes occur (i.c. NW of the Dri Hörlini).

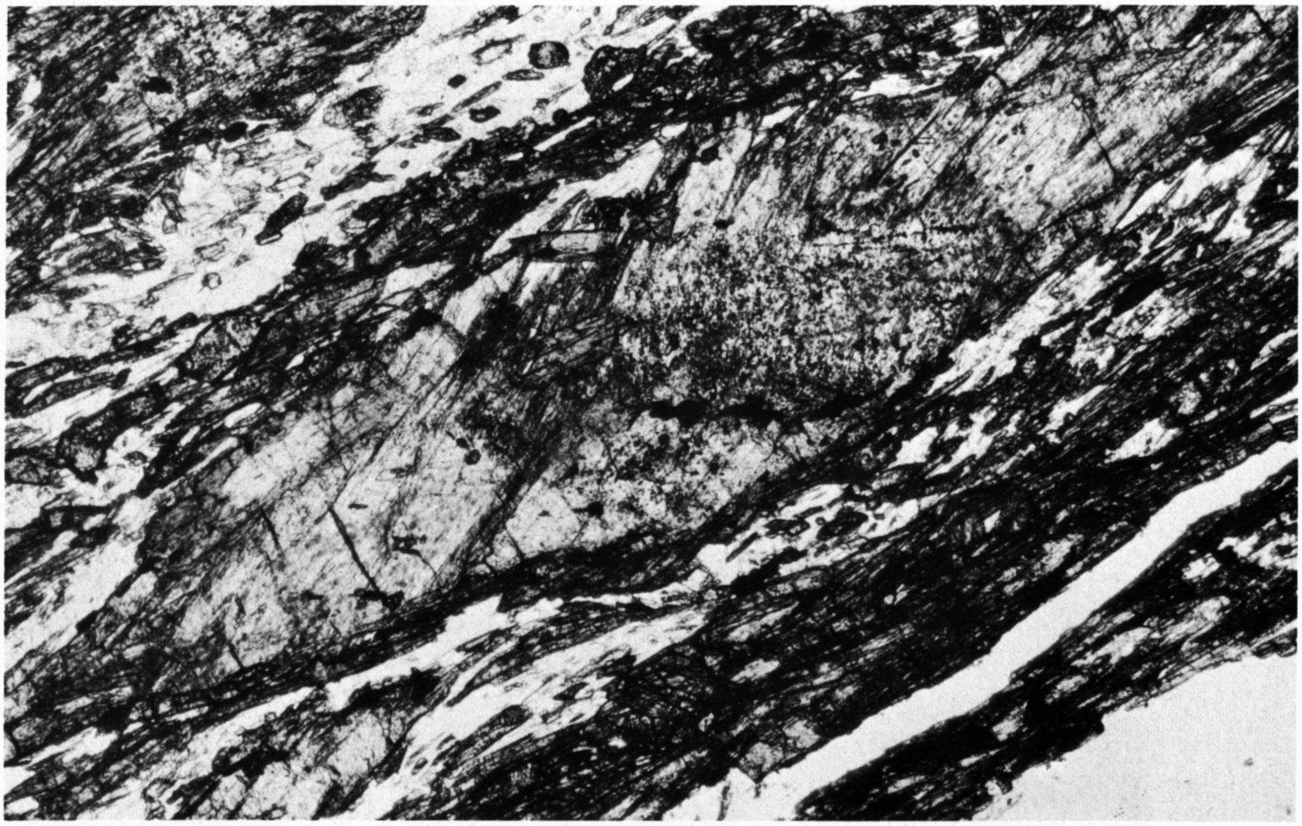


Fig. 2-6. Further development of S_1 (compare Figs. 2-4 and 2-5). A relic of the pyroxene pseudomorphs is enclosed by new amphiboles and elongate plagioclase crystals with a good preferred orientation in S_1. Amphibolite, Antrona Syncline. 25×, parallel polarizers.

CHAPTER 3

SECOND GENERATION STRUCTURES

3.1. GROUPING OF THE FOLDS AND THEIR RELATIONS TO EARLIER STRUCTURES

The F_2 folds were grouped on the basis of (1) refolding of first generation structures and (2) fold style. Numerous F_2 folds in the Antrona amphibolites, where S_1 and L_1 may be absent, have exclusively folded a compositional layering.

Interference between F_1 and F_2 folds is not particularly common, but refolded F_1 folds are distributed over the whole area. The angle between the axial directions of the F_1 and F_2 folds is variable. Small angles are prominent in the marginal parts of the Portjengrat zone, in some of the surrounding Mesozoic rocks, in the Furgg zone in the Val Loranco (Fig. 3-1B, C), and in the Furggtal. A small angle can also be observed in the Antrona amphibolites at the Alpe Pasquale (Fig. 3-1D), in the Fornalino area, and above Colorio. Large angles might, however, exist in the Antrona Syncline.

A high angle between F_1 and F_2 was found in the central part of the Portjengrat zone, where the F_1 folds are steeply plunging (Fig. 3-1A). In the western slopes of the Almagellhorn, northerly plunging F_1 folds have been refolded by westerly plunging F_2 folds, parasitic on fold 5 (Enclosures 1 and 2), producing a mushroom-like interference pattern. The high angle between the F_1 and F_2 axial directions, and the steep plunge of the F_1 folds are prominent in the Portjengrat zone east of the Pizzo d'Andolla, in the Furgg zone east of the Monte Rosa Nappe s.s., and in the frontal part of the Monte Rosa Nappe s.s.

The variability of the angle in the Camughera zone is less than in the Portjengrat zone. In the eastern spur of the Camughera, refolded F_1 folds show a Type 3 (Ramsay, 1967) interference pattern, but in the Val

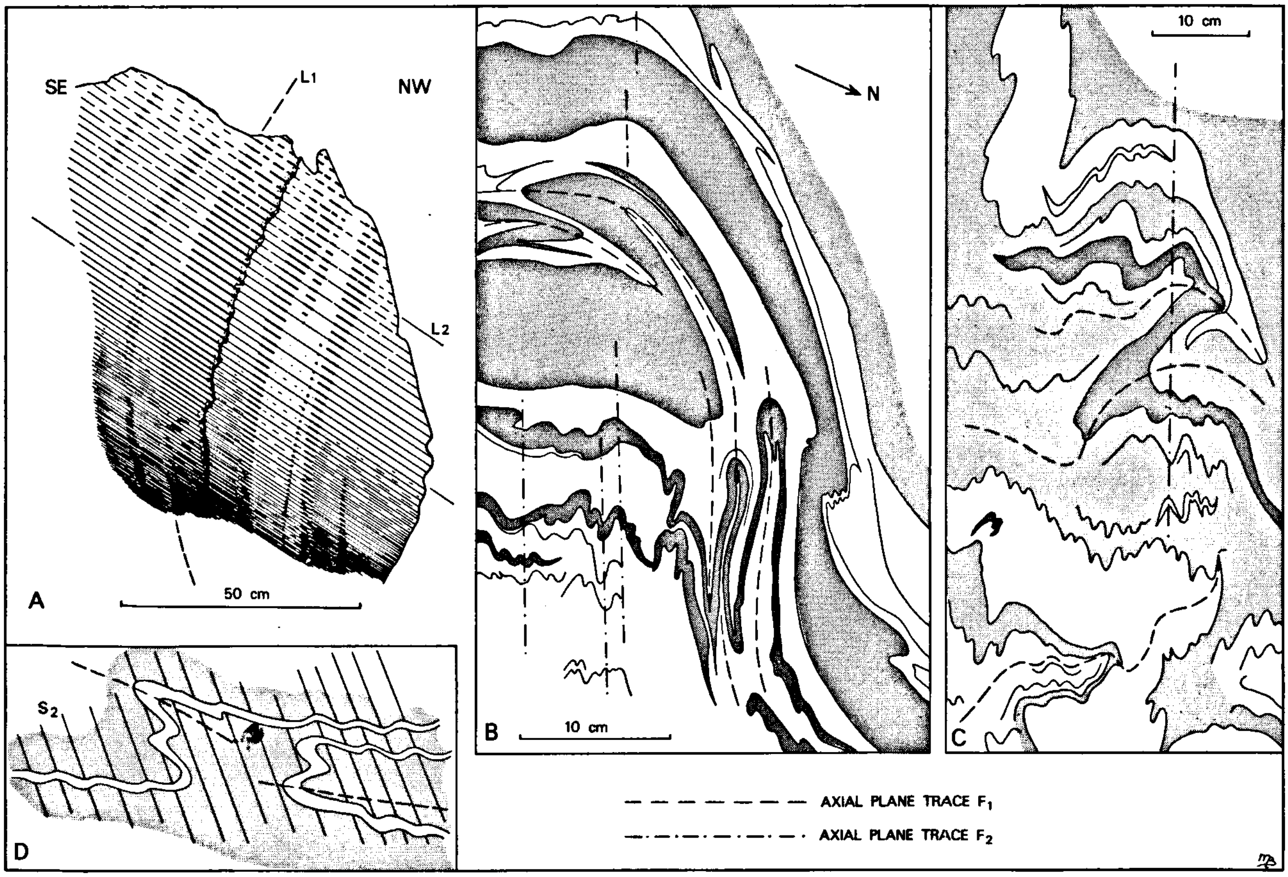


Fig. 3-1. Refolding of F_1 folds by F_2 folds. (A) Orthogneisses of the Portjengrat zone, Rotblattgletscher area. L_1 and L_2 enclose a very large angle. (B) Paragneisses of the Furgg zone, Alpe Camasco (Val Loranco). F_1 and F_2 fold axes are subparallel. (C) Paragneisses of the Portjengrat zone, northern slopes of the Rottal. F_1 and F_2 fold axes are subparallel. (D) Amphibolites of the Antrona Syncline, Alpe Pasquale. An S_2 axial plane schistosity intersects an F_1 fold (nearly actual size). All drawings were copied after photographs.

Brevettola river section, angles of 41° to 43° were measured between the F_1 and F_2 fold axes, these being the largest angles found. The large spread of the angular relations in the Portjengrat zone is found again in the Moncucco zone, but in the Monte Rosa zone, the F_1 and F_2 folds have subparallel axes.

These observations show that relations between the F_1 and F_2 folds are far from constant. The causes for this will be discussed later; see 5.4.3 and 7.5.

3.2. SIZES AND ORDERS OF F_2 FOLDS

Major folds, with amplitudes and limb lengths measured in kilometres, are the Mittaghorn Synform, the Trifthorn Antiform and the Antrona Syncline. The remaining, smaller F_2 folds are parasitic on the major folds, and have been called minor folds (Elliott, 1965). The minor folds have been subdivided into first order parasitic folds, parasitic on a major fold, and lower order parasitic folds; they do not have a characteristic size or periodicity. Folds with a more or less constant size and periodicity do occur in the field, but these qualities always depend on the rock association (local layering and nearby presence of other units which deform differently) in which the folds were formed, making generalization of little value. The term mesoscopic fold, used on Enclosure 1, is for the present purpose synonymous with minor fold.

There is considerable variability in the size of first order parasitic folds. The amplitude and length of limbs of the Zwischbergen and Camughera parasitic folds, and the fold formed by hinges 29 and 30 in the Antrona Syncline, are measured in hundreds of metres, but folds 25 and that formed by hinges 14 and 15 are much smaller. The folds contained by the Permotriassic quartzites in the Mittaghorn have short limbs with lengths between one and several metres, and long limbs measuring up to some tens of metres. First order parasitic folds of this size also occur in the paragneisses in the eastern part of the Furgg zone. The lower orders of parasitic folds vary correspondingly in size.

3.3. GEOMETRY OF THE F_2 FOLDS

3.3.1. *General*

The F_2 folds have similar (Turner & Weiss, 1963) fold style but true similar folds (Class 2, Ramsay, 1967) are almost absent. The geometry of Class 2 folds is closely approached when the folded rock body is nearly homogeneous and possessing a minimal layering and/or schistosity to show the folds. In most cases, however, the folds have similar style only when the profiles of a few adjacent layers are compared. In orthogneisses containing thin mica-rich layers (Fig. 3-2), the shape of the mica-rich layers in similar style fold hinges compensates for the difference in curvature of adjoining orthogneiss layers, demonstrating that these folds do not belong to Class 2. Analogous structures were found in other rocks.

Some degree of disharmony is characteristic of the F_2 folds, and many of them are strongly disharmonic (Enclosure 1; Fig. 3-3). Whitten (1966) defined a disharmonic fold by an abrupt change in profile in passing from one S-surface, or lithic unit, to another within a folded sequence. This definition encloses two kinds of disharmonic folds, distinguished by the fashion of fold profile change:

(a) The first kind occurs mainly in sequences of uniformly layered rocks where virtually each layer has a slightly different fold profile. The sequence has a dominant wave-length affecting all layers, but many individual layers contain apparently parasitic folds of shorter wave-length imposed upon folds of the dominant wave-length. The shape of the enveloping surfaces is almost the same, and the overall fold style resembles similar folding, but most of the folds in the sequence show some disharmony. (b) When individual layers have greatly different thickness or composition, the other kind of disharmonic fold is formed. This is characterized by a progressively changing fold profile from one lithic unit to another, and on an observational scale larger than the

Fig. 3-2. Similar-style F_2 folds in orthogneisses (Portjengrat zone). Thin micaceous layers compensate for geometrical differences of the folds in adjacent orthogneiss layers.

scale of the layering, the folds do not look like similar folds. The difference between fold profiles becomes greater with the distance between the layers.

The shape of all disharmonic folds could be related to changes in the folded layering (kind of lithology, thickness or change in thickness, termination of layers), and the degree of disharmony appears to be related to the importance of the change. The Mittaghorn Synform is an extreme example; it is isoclinal in the micaceous paragneisses east of Saas Grund but becomes a close fold in the prasinites of the zone of Zermatt–Saas Fee. Generally, plagioclase-rich layers behave competently unless plagioclase is present as isolated porphyroblasts; micaschists and quartz-mica schists behave as the less competent layers. The tighter portions of disharmonic folds are always situated in the more mica-rich layers, or in sequences with a high content of micaschists and quartz–mica schists. Plagioclase-rich layers determine the dominant wave-lengths of folds in thin adjoining mica-rich layers.

Numerous disharmonic folds have a somewhat unusual geometry (Figs. 3-3, 3-5G, 3-6G), that is independent of scale. The fold in Fig. 3-3 shows that the fold shape in the more feldspar-rich layers resembles a Class 1C fold (Ramsay, 1967) but that in adjoining more mica-rich layers the fold limbs are convex towards the axial plane, and the outer folded layers consequently form a tighter fold than the inner layers. No dip isogons (Elliott, 1965; Ramsay, 1967) can be drawn along part of the fold profile. The Mittaghorn Synform (see Enclosure 1 and 5.2.1) has this geometry, and it is also shown by the Trifthorn Antiform, although to a lesser degree. S_2 axial-plane schistosity is frequently well-developed in and near such hinges (Mittaghorn Synform, Figs. 3-8, 3-9).

Fig. 3-4. Some profiles of F_2 folds in paragneisses. All drawings have been copied from photographs. The folds in profiles A, B, E, H and J have folded compositional layering and S_1, the others have folded quartz veins and and S_1.
A: Steintälli, Portjengrat zone.
B: Almagellertal, idem.
C: Grundberg, idem.
D: Steintälli, Bernhard Nappe.
E: Vallone Pasquale, Furgg zone.
F: Steintälli, Bernhard Nappe.
G: Hannig, idem.
H: North of Villadossola, Moncucco zone.
J: Alpe del Gabbio, Furgg zone.
K: Val Brevettola, Moncucco zone.
L: Idem.

Fig. 3-5G shows that the axial planes of disharmonic folds can be a curved surface, and this is likely to be applicable to the disharmonic major folds. It is apparent that the disharmony leads to the formation of polyclinal folds (Greenly, 1919; Turner & Weiss, 1963; Ramsay, 1967). This is demonstrated by a synform in Antrona amphibolites above Pizzanco (Fig. 3-6B) where divergence of the axial plane takes place within one layer.

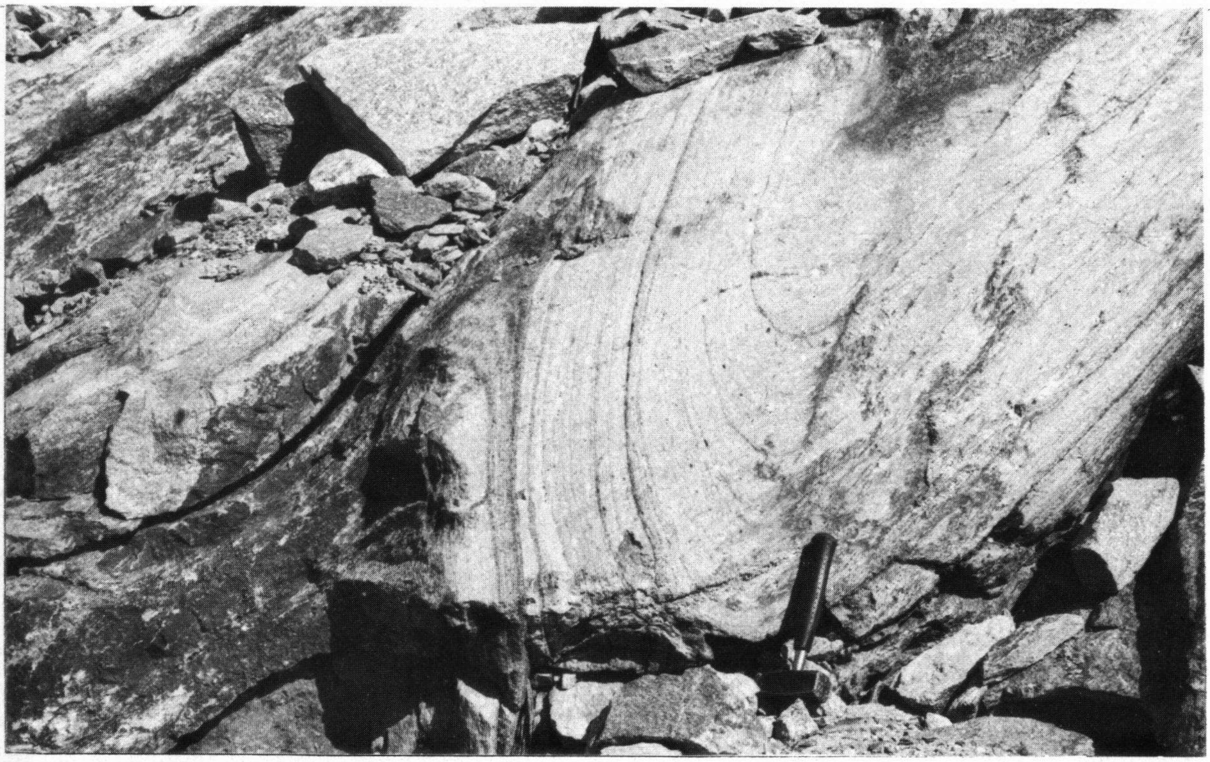
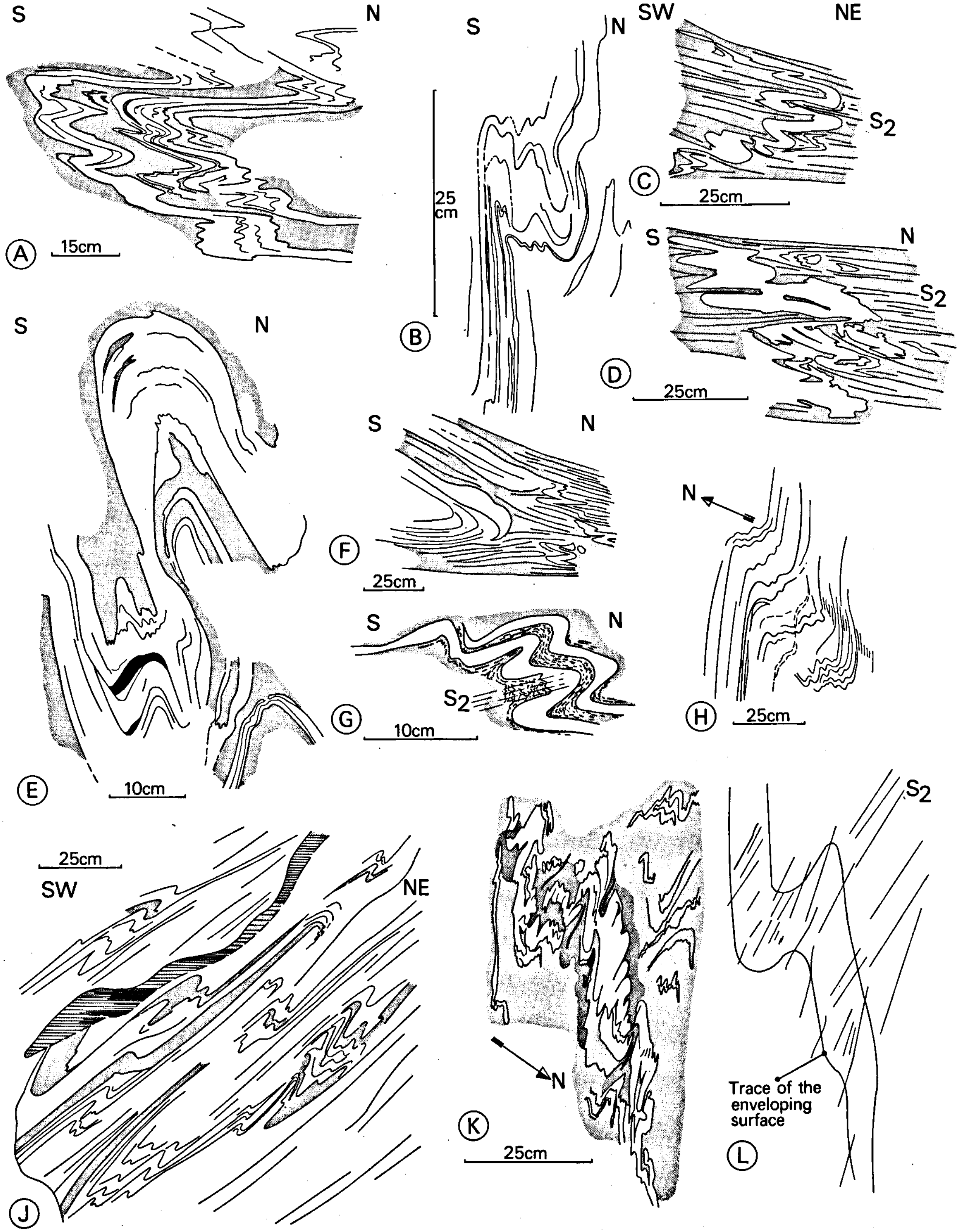


Fig. 3-3. Minor F_2 folds in orthogneiss northeast of the Rotblattgletscher. The limbs of the antiform are convex towards the axial plane.


S
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S2
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25cm
S
N
S2
G
10cm
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H
25cm
25cm
SW
NE
J
N
K
25cm
S2
Trace of the
enveloping
surface
L

Adjoining fold hinges in a coupled fold (Hansen, 1971) may have a different geometry, and in every case this can be attributed to some change in lithology.

Open, close and tight F_2 folds occur in all rocks, but isoclinal F_2 folds are limited to schistose and relatively mica-rich rocks, such as the majority of the paragneisses. In certain portions of the orthogneisses, notably between the Sonnighorn and the Almagellhorn, the Moosgufer area, the northern part of the Camughera gneisses and the Moncucco zone, isoclinal F_2 folds occur. Rootless F_2 folds are usually confined to the paragneisses near the axial planes of important folds. In orthogneisses they were found only twice, southwest of the Rotblattgletscher, and in the Camughera gneisses.

In some minor folds, the cylindricity is affected by lithological changes, and this is likely to apply to the major folds also, whose cylindricity is believed to be poor. Unfortunately it is very difficult to verify the cylindricity in the field, because of the prominent axial plunge of the F_2 folds.

The preceding text has shown that lithology difference exercised a strong control over the geometry of the F_2 folds. Different wave-lengths have been observed in layers of contrasting thickness and composition, and clearly the layering was not passive during folding (Donath & Parker, 1964; Ramsay, 1967). Biot (1965) states that truly similar folds develop only when the viscosity contrasts are very small, and it must be concluded that in this area viscosity contrasts were present during F_2 folding. Flinn (1962), Ramsay (1962, 1967) and Ramberg (1963a, 1963b, 1964) suggest that many apparently similar folds owe their initial development to the buckling mechanism, and Hudleston (1973) could demonstrate this for medium-grade metamorphic rocks near Loch Monar. It is believed here that the geometry of the F_2 folds in the present area has been determined by the buckling mechanism and a superimposed strain (see also Ramsay, 1967).

3.3.2. *Some properties of F_2 folds in various rock-types. The distribution of S_2*

Paragneisses. – A number of F_2 fold profiles are shown in Fig. 3-4. In the Portjengrat zone, first order parasitic folds in the Steintälli have close lower order parasitic folds in their hinge zones, and tight lower order parasitic folds on their limbs. This distribution of parasitic folds with unlike tightness has also been found in hinge 32, in the Camughera orthogneisses.

An S_2 axial-plane schistosity is frequent in the marginal parts of the Portjengrat zone, and particularly so in the hinge zones of folds 1–4 (Enclosures 1 and 2). S_2 is less common in the paragneiss between the two large orthogneiss bodies, where it was found in the Rotblattgletscher area, the Almagellhorn and in the Furggtal.

F_2 folds in the Camughera and Moncucco zones show less disharmony than in the Portjengrat zone, and many are almost pure Class 2 folds. S_2 axial-plane schistosity is frequent, particularly in quartz–mica schists in the Moncucco zone where F_2 folds may be difficult to distinguish from F_1. The amphibolites and amphibole gneisses of the Moncucco zone contain close-tight similar style F_2 folds that only rarely have an axial-plane schistosity. The hinge zones of F_2 folds present a prominent L_2 lineation of oriented amphibole, while in the limbs the L_1 amphibole preferred orientation is still present.

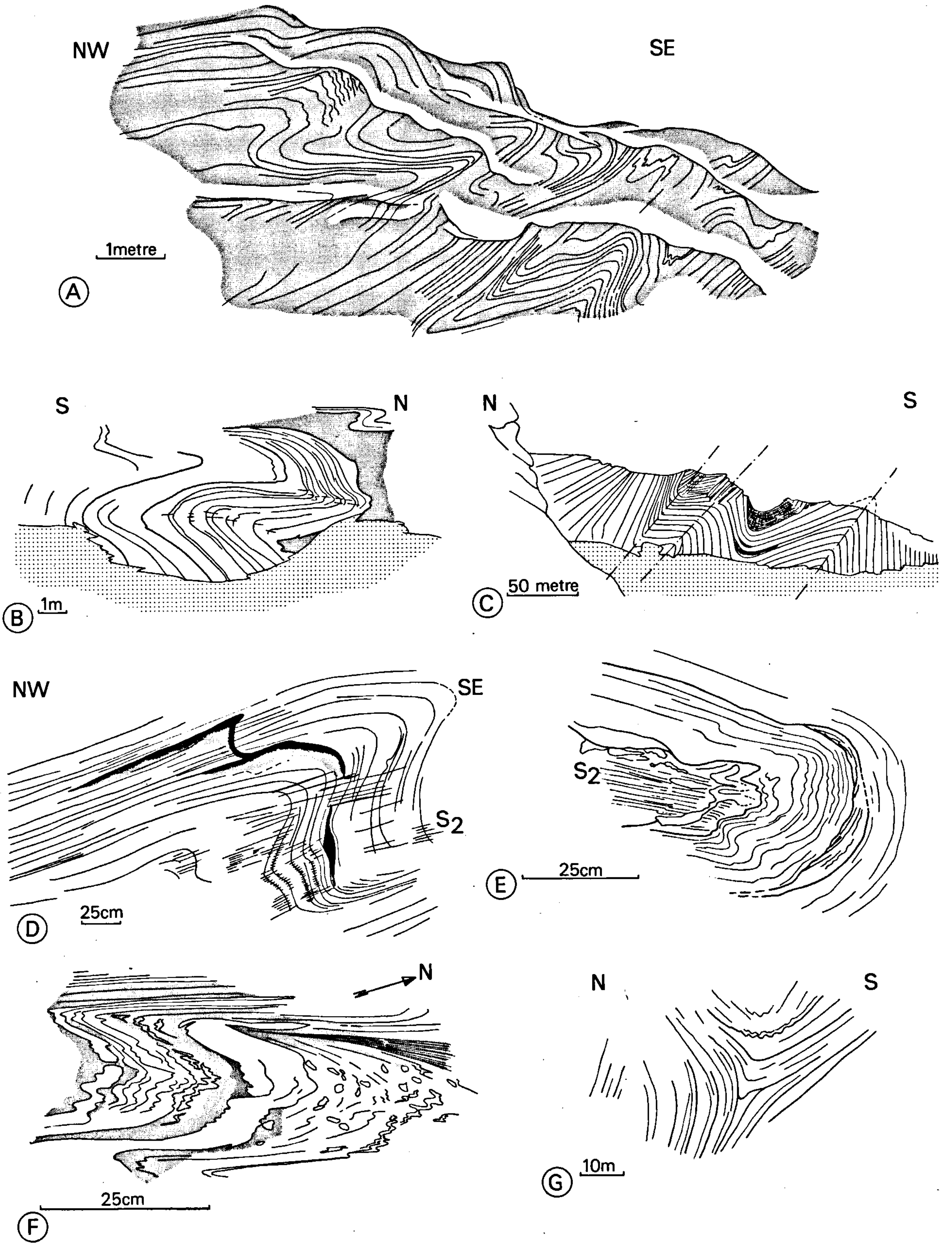
Orthogneisses. – Fig. 3-5 shows fold profiles from orthogneisses, and because these rocks usually have a coarser layering than the paragneisses, minor F_2 folds have larger dimensions. S_2 axial-plane schistosity is wide-spread but frequently non-penetrative. Its occurrence is not limited to the tightest folds but it is more penetrative in the more tightly folded or more quartz-rich rocks. S_2 is frequent in the Camughera–Moncucco Complex.

Furgg zone. – Strongly disharmonic F_2 folds occur in interlayered paragneisses and amphibolites in the eastern part of the Furgg zone, notably at the Alpe della Preja. An S_2 schistosity is locally important (eastern and western slopes of the Monte della Preja, and in the rocks adjoining the Portjengrat zone).

Mesozoic metasediments. – Fig. 3-6 shows fold profiles of minor F_2 folds in the Mesozoic metasediments. In the eastern slope of the Mittaghorn, F_2 folds occur in Permotriassic quartzites where they resemble close-tight Class 1C (Ramsay, 1967) folds. Lower order parasitic folds in the hinge zones of first order parasitic folds may form fold mullions (Wilson, 1953, 1961). The folds that most closely approach similar style contain an S_2 axial-plane schistosity (see 3.5.2), but in the calcareous schists where mica is not always present, S_2 may be absent even in the tightest folds. Disharmonic F_2 folds occur in interlayered calcareous schists and ophiolites. In the Antrona Syncline close to i Gerbi, F_2 folds of similar style were formed in calcareous schists with a subordinate mica content, and in their hinge zones the layering has been transposed into lenses parallel to the axial plane.

Antrona amphibolites. – Some fold profiles are shown in Figs. 3-6. S_2 axial-plane schistosity is infrequent and may be difficult to detect in the field, with the exception of some F_2 folds at the Alpe Pasquale (see 3.5.4).

Fig. 3-5. Some profiles of F_2 folds in orthogneisses. All drawings have been copied from photographs. All folds have folded a compositional layering parallel or subparallel to S_1; in B and E the orthogneisses are interlayered with schist.
A: Rotblattgletscher, Portjengrat zone.
B: Italian side of the Portjengrat (Pizzo d'Andolla).
C: Southern part of the Portjengrat as seen from the Rotblattgletscher.
D: Kehrenrück (Furggtal), Monte Rosa Nappe.
E: Loose block northwest of the Rottblattgletscher.
F: North of Villadossola, Moncucco zone.
G: Alpe Saudero, Camughera zone.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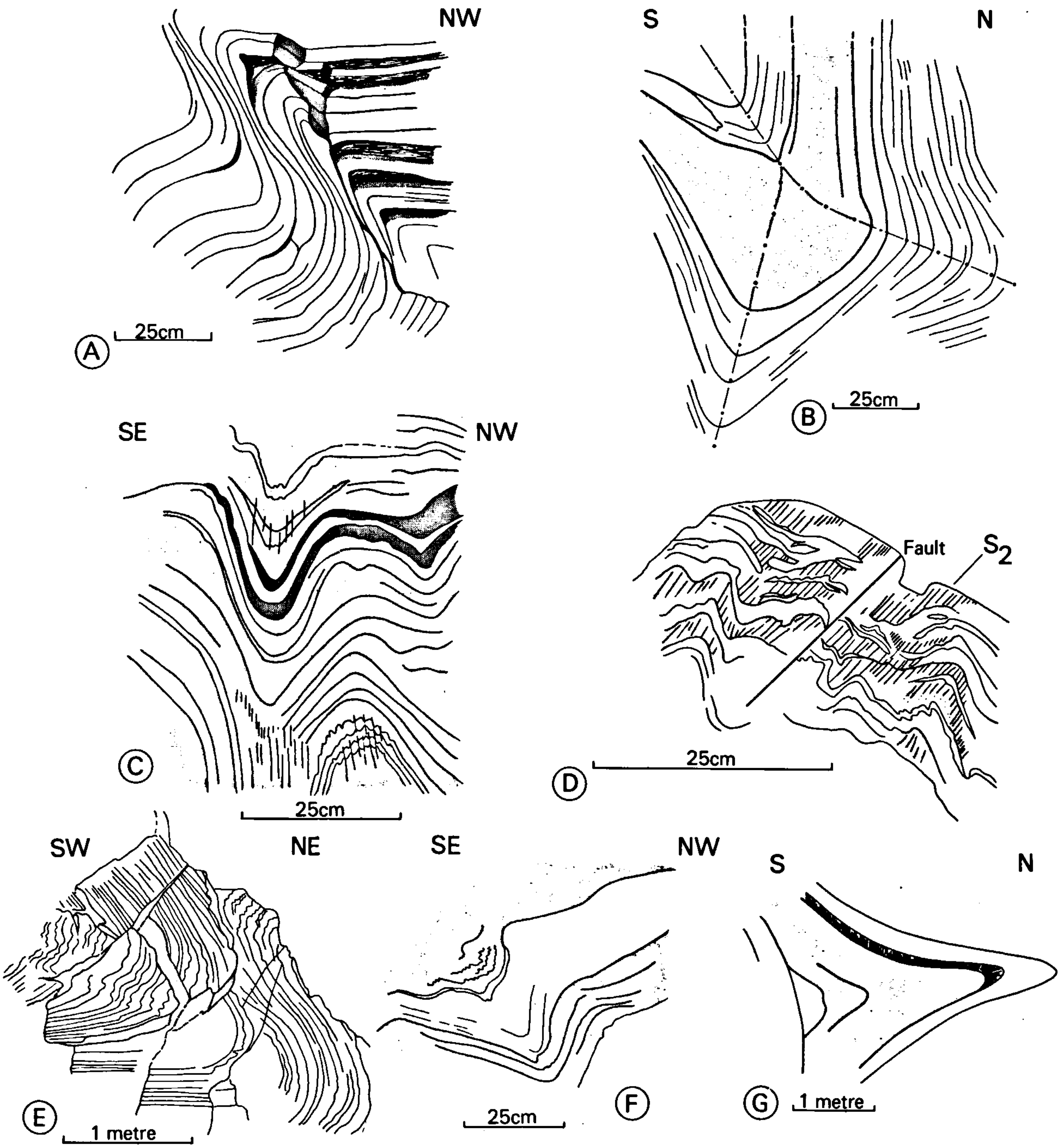


Fig. 3-6. Some profiles of F_2 folds in Permotriassic and younger metasediments and the Antrona amphibolites. All folds have folded a compositional layering parallel or subparallel to S_1. All drawings, with the exception of G, have been copied from photographs.
A: Antrona amphibolites, Passo d'Arnigo.
B: Antrona amphibolites, Alpe Pranzano.
C: Bündnerschiefer, I Gerbi.
D: Antrona amphibolites, Alpe Pasquale.
E: Permotriassic quartzites, Mittaghorn area.
F: Antrona amphibolites, Cresta.
G: Antrona amphibolites, Alpe Pranzano.
The 'competent or incompetent' behaviour of the Antrona amphibolites is chiefly governed by the intensity of S_1 development.

3.4. MICROSCOPIC PROPERTIES OF F_2 FOLDS AND THE DEVELOPMENT OF S_2, BASEMENT ROCKS

3.4.1. *Paragneisses – Saas area*

F_2 folds. – Colourless mica and chlorite have been deformed together with S_1, and have frequently retained their preferred orientation in that plane. They have undulose extinction and curved (001) planes, or they may have formed polygonal arcs (see below) that are frequent in quartz–mica schists but infrequent in plagioclase-rich rocks.

The width of the hinge zone in a single close-tight folded layer is frequently seen to increase from the core towards the outer part of the fold, and this is probably a remnant of the early buckle folding. Post-F_2 grown albite and/or chlorite can occur in the hinges, and colourless mica frequently has a smaller grain size in fold hinges than in the limbs. Isoclinal folds have the same microstructure, apart from the rotation of S_1 in the limbs into parallelism with S_2 which results in a preferred orientation with (001) parallel to S_2 of the micas contained within these limbs.

Polygonal arcs of colourless mica (Figs. 3-7, 3-10) are defined as fold hinges pictured by polygonized micas (Spry, 1969), and because the micas are rarely completely strain-free, the term is used here when the cumulative curvature of (001) in the micas is less than the curvature of the fold hinge. A decrease in mica grain size accompanies their formation. In tight-isoclinal folds, it frequently occurs that the quartz content is lower in the limbs than in the hinges of the polygonal arcs, where the quartz is present as irregularly shaped grains between frayed micas or as triangular grains between straight micas. Biotite and chlorite also occur as triangular grains (Figs. 3-8, 6-5; see 6.5-6), but colourless micas with a triangular shape and enclosed by other colourless micas have not been found.

Quartz subgrains have been formed in older grains, some of these subgrains are elongate and oriented subparallel to the a-direction of the F_2 fold. The preferred orientation of quartz c-axes describes a great-circle girdle in the plane perpendicular to L_2.

In the massive albite-rich gneisses the albites were deformed, whereas in the porphyroblastic gneisses the albites were rotated but not always deformed. In gneisses consisting mainly of albite porphyroblasts and colourless mica, the microstructure of the F_2 folds may be very complicated due to irregular folding in the micaceous part of the rock between the rotated albite porphyroblasts.

S_2 schistosity. – In the hinge zones of tight-isoclinal minor folds in the Triftgrätji, S-shaped microfolds are separated by an S_2 schistosity of colourless mica and

Fig. 3-7. Polygonal arcs of colourless micas. F_2 fold in paragneisses of the Portjengrat zone, Almagellertal. 10×, parallel polarizers.

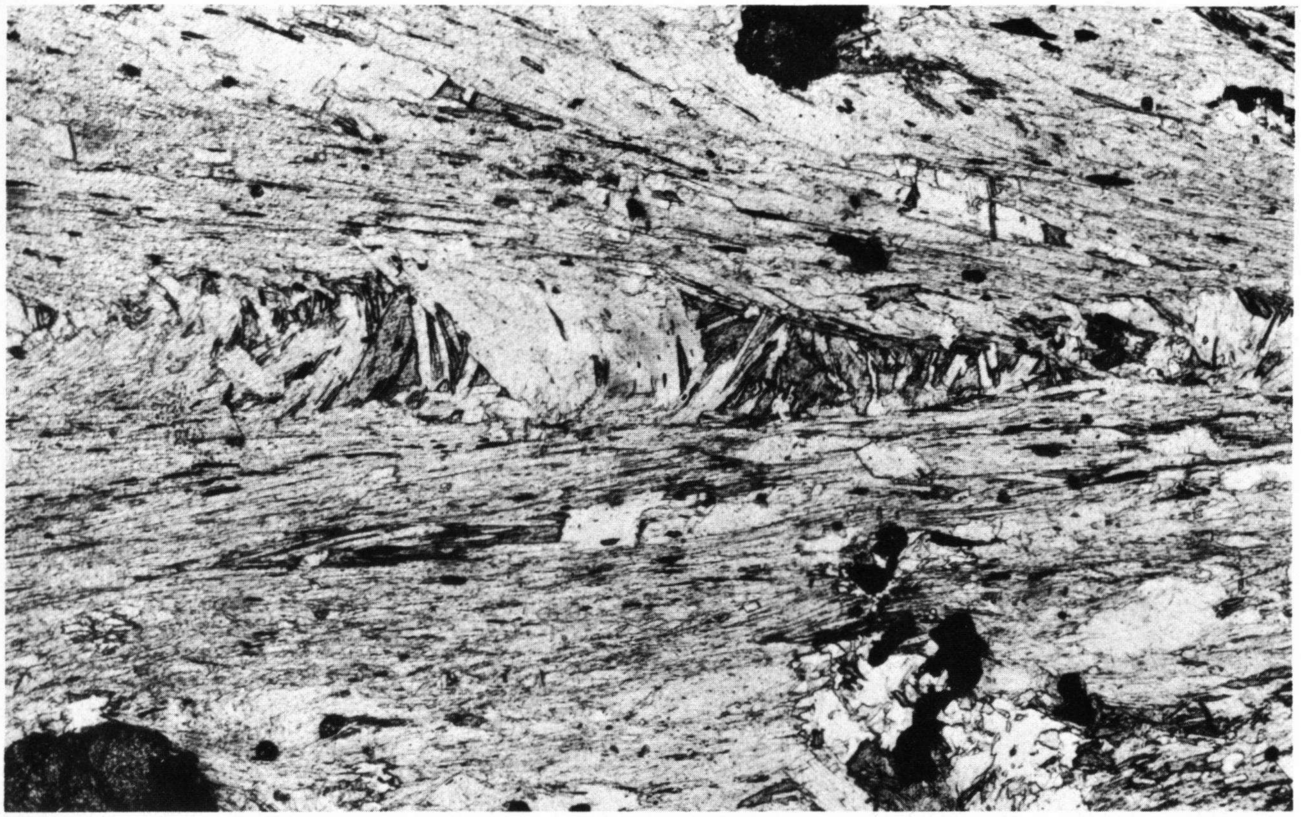


Fig. 3-8. F_2 fold in paragneisses of the Bernhard Nappe, Hannig. S_1 has been rotated almost into parallelism with S_2, but the latter is not a material plane. Chlorites tend to form triangular grains in the hinge zone, and they also occur in the stress shadows of some albites. F_2 folding changed the shape of the muscovite metablast in the center, normally these are elongate parallel to S_1. 10×, parallel polarizers.

subordinate chlorite. The width of the S_2 planes is small when compared with the width of the folds (Fig. 3-9), and the form of the S-shaped folds varies between very open, where S_1 makes a small angle with S_2, and tight where the enveloping surface of folded S_1 makes a high angle with S_2. In the open F_2 folds, the colourless micas in S_1 show less evidence for deformation than in the tight F_2 folds, and many of them lie partly in S_1 and partly in S_2, so that the S_1 planes curve into an S_2; the latter is defined by the short limbs of the F_2 folds. Gradual transitions towards more tightly folded folds exist, and they show that mica grain boundaries form at the location where the S_1 and the S_2 planes meet, and micas curving from S_1 into S_2 become rare. Finally, all micas in S_1 terminate abruptly against S_2 (Fig. 3-8) which now contains perfectly oriented micas, and forms the long limb of the F_2 folds. Adjoining folded micas in S_1 curve in opposite directions towards S_2, which could be evidence for deformation by flattening. Triangular grains of chlorite occur between these colourless micas (Fig. 3-9).

When the F_2 folds are isoclinal, and S_1 has been rotated into parallelism with S_2, except for the fold hinges, isolated domains of transverse micas representing relics of these hinge zones are the only remnants of S_1 between the S_2 planes.

In quartz-mica schists adjoining the southern contact of the Portjengrat zone orthogneisses in the Val Loranco, S_2 is well-developed, and the microfolds associated with it change their habit according to the mica-content of the rock (Fig. 3-10). The mica-rich layers contain isoclinal folds with S_1 in the limbs rotated into parallelism with S_2. Numerous micas in the mica-poor layers are well-oriented in S_2, others form polygonal arcs between the S_2 planes, whilst very few micas lie partly in S_1 and partly in S_2. A comparison of the folds in the mica-rich and in the mica-poor layers shows that the S_2 schistosity is formed by reoriented micas. The S_2 planes do not persist for long, and replace each other 'en échelon' on thin section scale. Chlorite and subordinate biotite have a preferred orientation in S_2, and they have probably grown in that position. This applies in particular to the biotite in S_2 that shows no evidence for reorientation.

At many places in the Grundberg and close to the valley-floor of the Saastal a 'Zeilenbau' dominates in the quartz-rich layers. It can be dated as S_2 because of the presence of albite porphyroblasts, since folded S_1 is only visible where the porphyroblasts deviate the S_2 planes, and the latter's spacing becomes larger.

Discussion. – It is believed that S_2 in isoclinal folds in

Fig. 3-9. F_2 folds and S_2 in the closure of the Mittaghorn Synform, Triftgrätji. S_2 is a material plane that was formed by reorientation of S_1, and it is defined by colourless micas of pre-F_2 age that were rotated with S_1. In the lower right hand corner of the photograph, colourless micas are curving in opposite directions towards S_2, suggesting a flattening element in the deformation. New grains of chlorite occur selectively in the fold hinges. 10×, parallel polarizers.

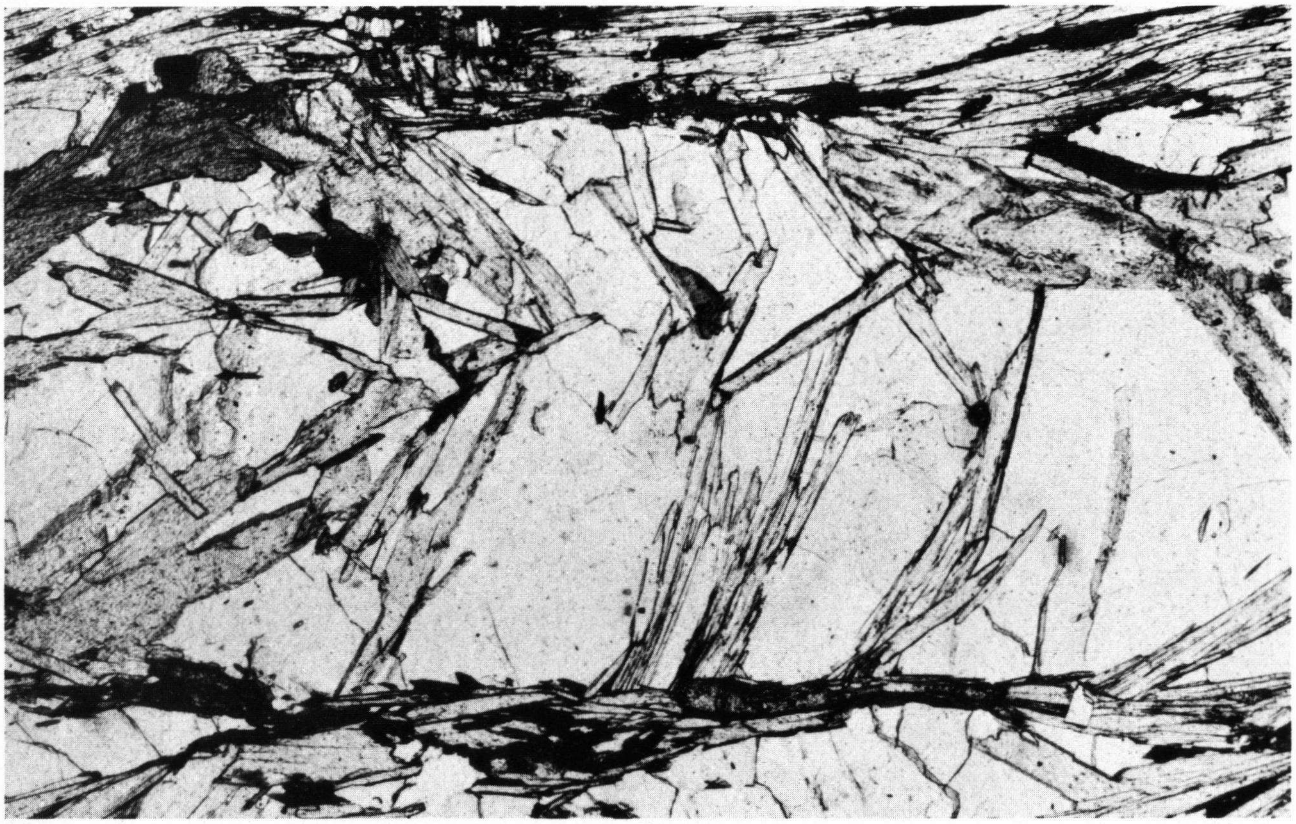
the paragneisses of the Portjengrat zone has been formed by reorientation of pre-existing colourless micas. Important arguments are the presence of micas that curve from S_1 into S_2, and the, not previously mentioned, absence of two optically recognizable generations of colourless mica. Another argument is the absence of axial-plane schistosity in open-close F_2 folds. Where the micas oriented in S_2 have not recrystallized, they appear strongly deformed also when no evidence is found for F_3 deformation, and this can only be explained by deformation during the F_2 reorientation into their present position.

Shear-S_2. – A coarse S_2 schistosity occurs in the Triftgrätji, below the summit of the Portjengrat, and in the southwestern slopes of the Almagellhorn. The S_2 planes are between 3 mm and 2 cm apart, and are sharply bounded layers of very fine-grained quartz and mica, much finer grained than outside the S_2 planes, and offsetting individual quartz-rich layers. They are therefore considered as shear planes. The S_1 surfaces always form acute angles with S_2, and show a slight curvature towards S_2 near to the S_2 planes. Their orientation differs significantly from the usual orientation of the F_2 planar and linear elements, but in most cases the angular relationship between S_1 and shear-S_2 can be related to the F_2 fold closures. This shear-S_2 has possibly been formed during the last stage of F_2 deformation.

3.4.2. *Paragneisses – remaining area*

In the Camughera zone, F_2 microfolds are not much different from the Saas area but polygonal arcs are more frequent. F_2 folds with an S_2 axial-plane schistosity are common, and S_2 is defined by biotite with a preferred orientation of (001) parallel to the axial planes of the folds. These include close-tight folds, where the rotation of the limbs has been insufficient to reorientate the pre-existing micas parallel to S_2, and where colourless micas are conspicuously absent in S_2. It is therefore believed here that in the Camughera zone biotite crystallized in S_2, and that in contrast with the Saas area, significant biotite growth took place synchronously with F_2. The preferred orientation of biotite in S_1 (see 2.4), and the presence of undeformed biotite as triangular grains in polygonized F_2 folds demonstrate that in the Camughera zone biotite growth began syn-F_1 and outlasted F_2, and as the temperature culmination in the Camughera zone took place just before F_2 (see 6.11), the conditions for biotite growth must have been favourable during F_2.

The same applies to the Moncucco zone, where S_1



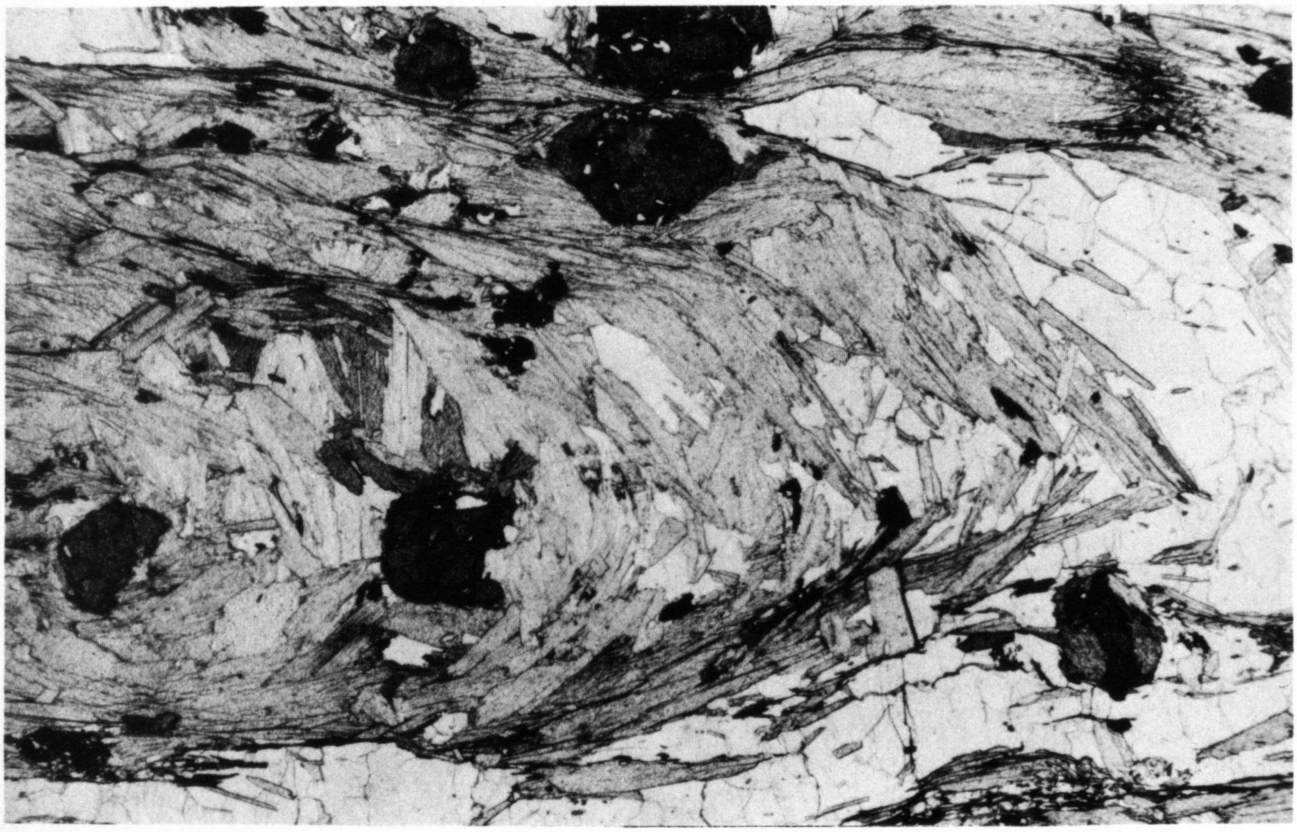


Fig. 3-10. A (upper photograph) and B (lower photograph). F_2 fold in paragneiss of the Portjengrat zone, Val Loranco. S_2 is parallel to the long sides of the photographs. The same fold is shown in a quartz-rich layer (A) and in a mica-rich layer (B). Reorientation of colourless micas is due to isoclinal folding and forms S_2 in the mica-rich layer, for apart from mica polygonization no distinct structural break occurs between the micas oriented in S_1 and those oriented in S_2. Therefore, also the fold in the quartz-rich layer must have formed in this manner, and the colourless micas in S_2 are reoriented older micas. Reorientation of the micas in the fold limb into S_2 may have been facilitated by redistribution of quartz between S_2 and the fold hinges. Biotite (dark), however, crystallized in S_2 and parallel to it. 10×, parallel polarizers.

and S_2 are not easy to distinguish. Few traces of S_1 remain when an S_2 schistosity has been formed, and the areal distribution of S_2 could therefore not be accurately determined. S_2 is a tectonic layering of thin (mm-size) quartz-rich layers separated by layers of colourless micas and biotites, with a thickness corresponding to one or two micas. Biotite occurs also between the S_2 planes, oriented in S_2 or in folded S_1. In many instances S_2 can only be safely dated when plagioclase porphyroblasts are present, either because they possess curved inclusion trails (plagioclase growth syn-F_2) or because rotated porphyroblasts with a straight S_i (= S_1) occur between the S_2 planes. Locally, S_1 and S_2 are intersecting at a small angle.

3.4.3. *Orthogneisses – Saas area*

F_2 folds. – The F_2 folds in the thinly layered, quartz-rich portions of the orthogneisses have much the same microscopic characteristics as in the paragneisses. In the albite-rich orthogneisses, infrequent polygonal arcs are defined by slightly deformed colourless micas lying in the fold limbs and meeting at an angle in the hinge. In some cases the micas do not actually touch each other. Quartz and albite show undulose extinction. F_2 microstructures are geometrically irregular where ill-defined S_1 schistosity planes are common. Numerous microcline porphyroclasts have been rotated.

S_2 axial-plane schistosity. – Two general cases can be distinguished:

a. An S_2 of discontinuous planes, defined by strongly deformed colourless micas which are arranged in those planes but have a weak preferred orientation of (001) in S_2. Some of the micas have been boudinaged. In a good example of this microstructure (Fig. 3-11, from the hinge of fold 14, Enclosure 1) the thickness of the S_2 planes is no more than one or two micas, 0.3 mm at the most, and the length of the planes in sections perpendicular to L_2 averages 4.5 mm. They are spaced at distances between 1.3 and 5.1 mm, and when they terminate they are replaced by others in the vicinity. The planes are, however, not always clearly defined on microscopic scale, and they may be faint zones of higher mica content parallel to S_2.

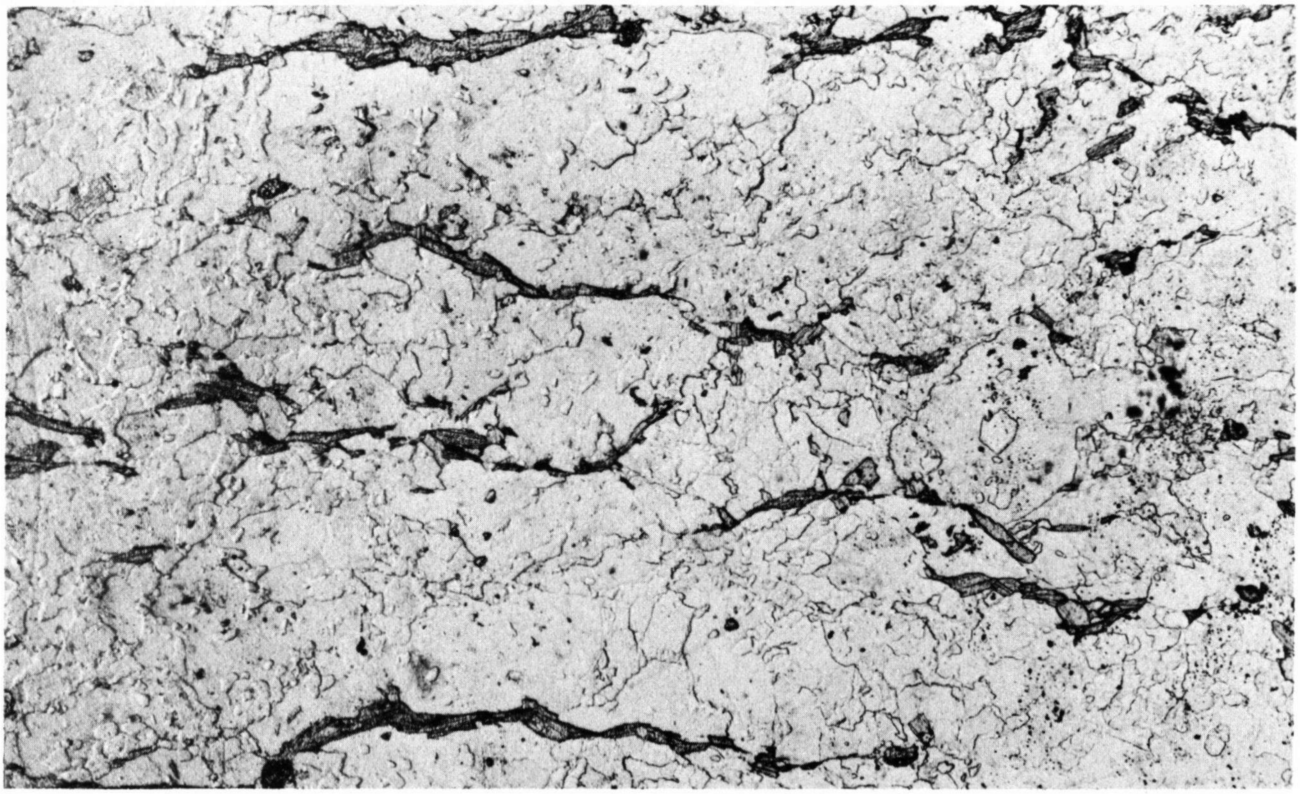


Fig. 3-11. S_2 axial-plane schistosity in orthogneiss (Portjengrat zone, Almagellhorn) consisting of discontinuous planes defined by strongly deformed colourless micas. 25×, parallel polarizers.

A specimen from the closure of fold 7 (Zwischbergenpass) contains a Class 1C (Ramsay, 1967) fold with an amplitude of 5 cm. The S_2 planes occur in the core of this fold, where the rock composition approximates albite 70%, quartz 15% and colourless mica 15%. S_2 is absent in a layer of pure albite gneiss and in the outer part of the fold, where folded micas occur exclusively. In the core of the fold, colourless mica is concentrated in the S_2 planes that also contain quartz grains of much smaller-than-average size, presumably new grains of syn-F_2 age. Some of the micas lie partly in S_1 and partly in S_2, but usually a grain boundary exists at the place where a mica has been folded. A few micas situated between the S_2 planes are oriented in S_1 or do not show an obvious preferred orientation, and they are of larger grain size than the micas oriented in S_2. No evidence was found for the growth of new, undeformed and perfectly oriented micas in S_2. This microstructure predominates in the closures of F_2 folds, and good examples of it occur in the Portjengrat, between the Portjengrat and the Mittelrück, in the Moosgufer area, and in the Almagellhorn. The new quartz grains in S_2 were found only at the Zwischbergenpass.

b. An S_2 of colourless micas that are not always arranged in well-defined schistosity planes (Fig. 3-12), but form a penetrative S_2. Their preferred orientation is good, and (001) is generally parallel to the axial plane of the F_2 fold. Between them occur micas that are oriented in S_1, and define occasional F_2 fold hinges. All micas show undulose extinction, and there is no apparent difference in grain size between the micas in S_1 and in S_2. Some micas lie partly in S_1 and partly in S_2, but only when polygonal arcs are absent, for in that case, grain boundaries separate the micas in S_1 from those in S_2. The S_1 and S_2 planes may be difficult to distinguish. This S_2 occurs in the closures and in the limbs of F_2 folds, but its occurrence is restricted to relatively quartz-rich rocks (>40% quartz).

In both examples the quartz microstructure consists of equant or elongate subgrains in older grains. The boundaries of some of the elongate subgrains were imposed upon them by colourless mica (Wilson, pers. comm.), but other subgrains are elongate in S_2, parallel or subparallel to the a-direction of the fold, without being bounded by mica.

Discussion. – From (1) the absence of two optically recognizable generations of colourless mica, (2) the presence of deformed colourless micas oriented in planar S_2, (3) the colourless micas that curve from S_1 into S_2, and (4) the difference in grain size frequently shown by the colourless micas in S_1 and in S_2, it has been concluded that in the orthogneisses of the Portjengrat

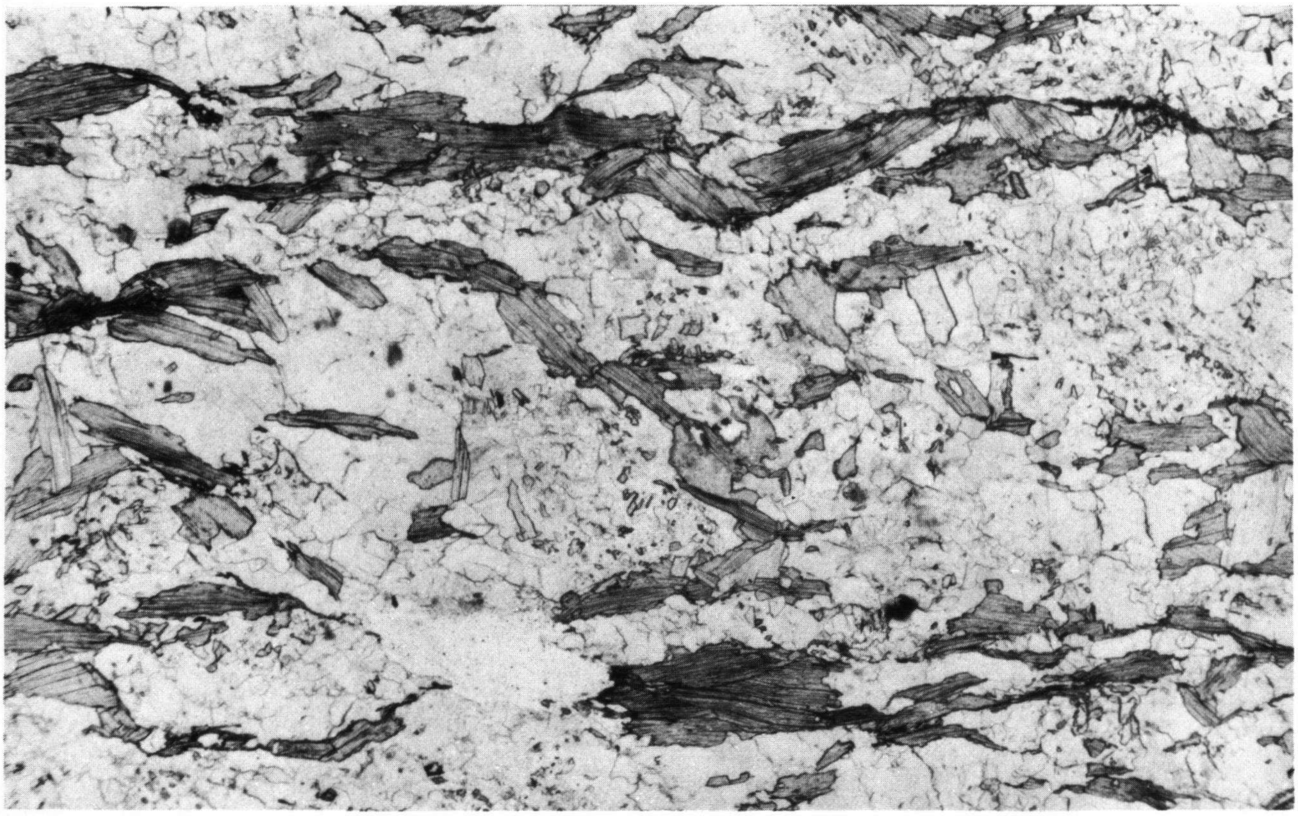


Fig. 3-12. S_2 axial-plane schistosity in orthogneiss from the Portjengrat, with remnants of S_1 between S_2. Relatively large colourless micas have a good individual preferred orientation parallel to S_2, but they are not always arranged in planes. 25×, parallel polarizers.

zone the S_2 axial-plane schistosity was formed by reorientation of pre-existing colourless micas. It differs from the process of reorientation in the paragneisses, because in the orthogneisses S_2 is frequent in close-tight folds, and common in fold closures where the angle between S_1 and S_2 is large. Though simple rotation of pre-existing micas into S_2, as in the paragneisses, could have taken place in isoclinally folded quartz-rich orthogneisses, the process of reorientation has to be more complicated to account for the reorientation of colourless micas into parallelism with the axial planes of close-tight folds. This is demonstrated by the small micas in the quartz-rich layer of Fig. 3-13, that tend to lie in S_2 in contrast with the large folded micas in the mica-rich layer.

This and the previous observations suggest that the grain size of the micas, and the rock composition affect the process of reorientation. In the orthogneisses, the plagioclase vs. quartz content seems to be of prime importance, and observations in the paragneisses (see 3.4.1) and in the Permotriassic quartzites (see 3.5.2) show that reorientation of micas in mica-rich layers is less common than in micaceous layers with a moderate to high quartz content, with or without subordinate plagioclase.

It is not yet understood why the colourless micas in the orthogneisses occur in planes, ill-defined as they may be. Williams (pers. comm.) suggested that some kind of volume change due to the removal of material (particularly quartz) during folding might have taken place, and this could have been a factor in the formation of the S_2 planes. If this should prove to be the case, it must be considered as a possibly contributing mechanism for the reorientation of colourless micas in the paragneisses and in other rocks. Migration of quartz may be indicated by numerous instances of cm- to dm-sized quartz bodies adjoining the more competent layers in the cores of strongly disharmonic folds.

3.4.4. *Orthogneisses – remaining area*

The F_2 folds in the orthogneisses of the Camughera zone resemble those from the Saas area, when allowance is made for the somewhat different S_1 (see 2.3) and compositional layering (see 1.5.2). Post-F_2 recrystallization is more important than in the Saas area, and well-developed polygonal arcs are frequent. An S_2 axial-plane schistosity, defined by the preferred orientation of (001) of biotite, occurs in open and close folds as well as in isoclinal folds. By analogy with the paragneisses of the Camughera zone (see 3.4.2), biotite

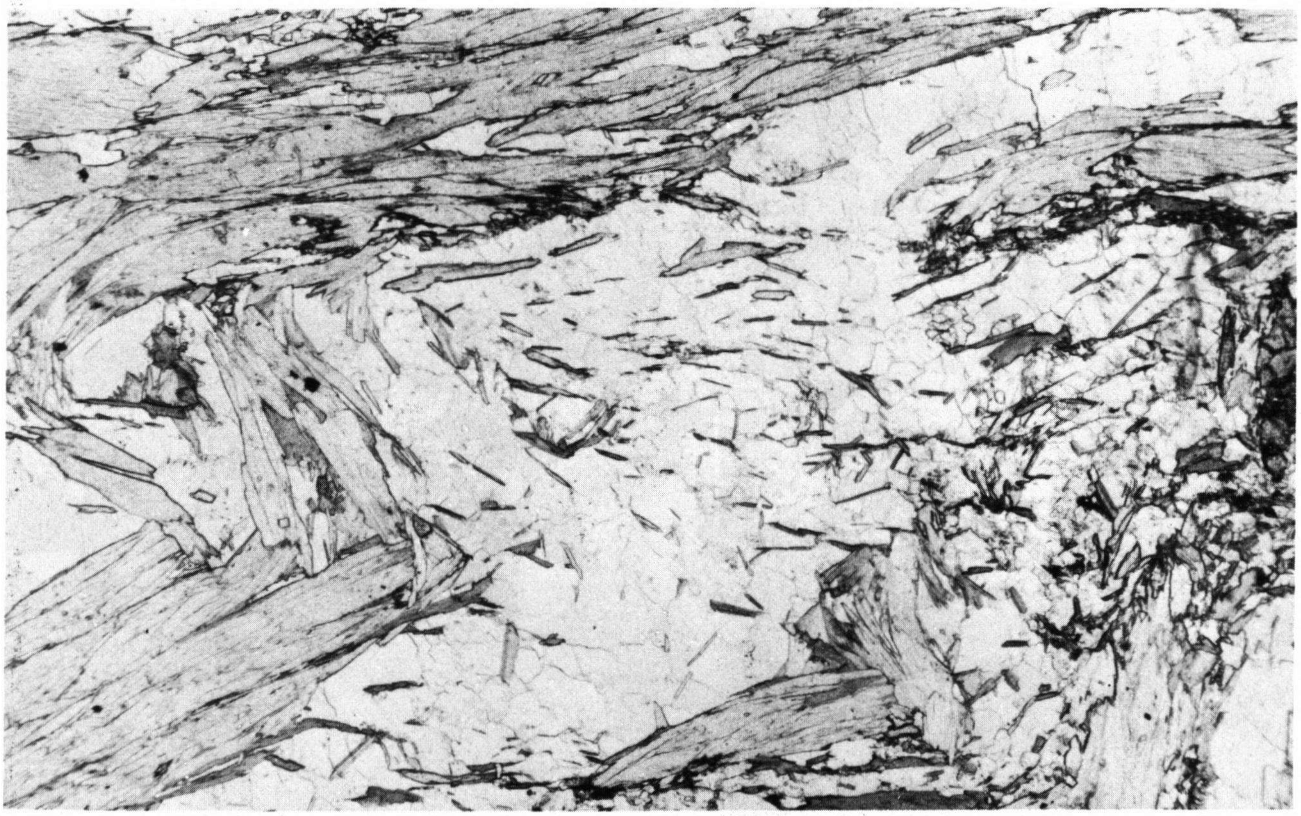


Fig. 3-13. Small colourless micas and biotites define an S_2 axial-plane schistosity in an F_2 microfold, while the large colourless micas defining S_1 have been folded. Biotite occurs as triangular grains between colourless micas in the polygonal arcs. Crystallization of biotite is syn-post F_2; rotation of small colourless micas only. Orthogneiss, Rotblattgletscher (Portjengrat zone). 10×, parallel polarizers.

growth must have taken place simultaneously with F_2 deformation. A tenuous biotite-defined S_2 in similar-style minor folds in the hinge of fold 31 (Enclosure 1) makes a small angle with the axial planes of these folds, whose wave-lengths are indicative for a buckling element in the deformation (see 3.3.1). It seems possible that this S_2 results from superimposed strain on already formed folds, with the XY plane of the superimposed strain ellipsoid not coinciding with the axial plane, and assuming that the S_2 schistosity is mainly or completely the product of the superimposed strain.

F_2 folds in the orthogneisses of the Moncucco zone are much the same; they are fairly prominent northwest of Villadossola, where S_2 is also important.

3.5. MICROSCOPIC PROPERTIES OF F_2 FOLDS AND THE DEVELOPMENT OF S_2, FURGG ZONE AND MESOZOIC ROCKS

3.5.1. *Furgg zone*

The descriptions of the F_2 folds and the S_2 axial-plane schistosity in the paragneisses of the Saas area apply also to the eastern part of the Furgg zone, although polygonal arcs tend to be better developed. A strong dimensional orientation of quartz subgrains in the axial plane of F_2 folds is common. Small newly grown quartz grains were found in S_2 shear planes in the eastern slopes of the Monte della Preja (compare with Pizzo Montalto, 3.5.2).

3.5.2. *Permotriassic quartzites and other quartzitic Mesozoic rocks*

F_2 folds. – The best examples occur in the eastern face of the Mittaghorn where the following observations were made. When an S_2 axial-plane schistosity is absent, the colourless micas retain their preferred orientation in S_1 during F_2 folding, and they are rotated and deformed along with S_1. Polygonal arcs are absent in layers consisting exclusively of colourless micas, though some microfolds are formed by straight micas meeting at an angle in the hinge. The micas are smaller in the quartz-rich layers, and polygonal arcs are more frequent there.

The quartz microstructure consists of subgrains in older, larger grains, and the preferred orientation of quartz c-axes is a great-circle girdle perpendicular to the L_2 lineation.

S_2 schistosity. – In the eastern part of the Mittaghorn, S_2 is present in thin mica-rich layers with a thickness of up to three or four micas, and in quartz-rich layers with uniformly distributed micas. In thicker mica-rich layers, crenulation of S_1 took place, resulting in a local and

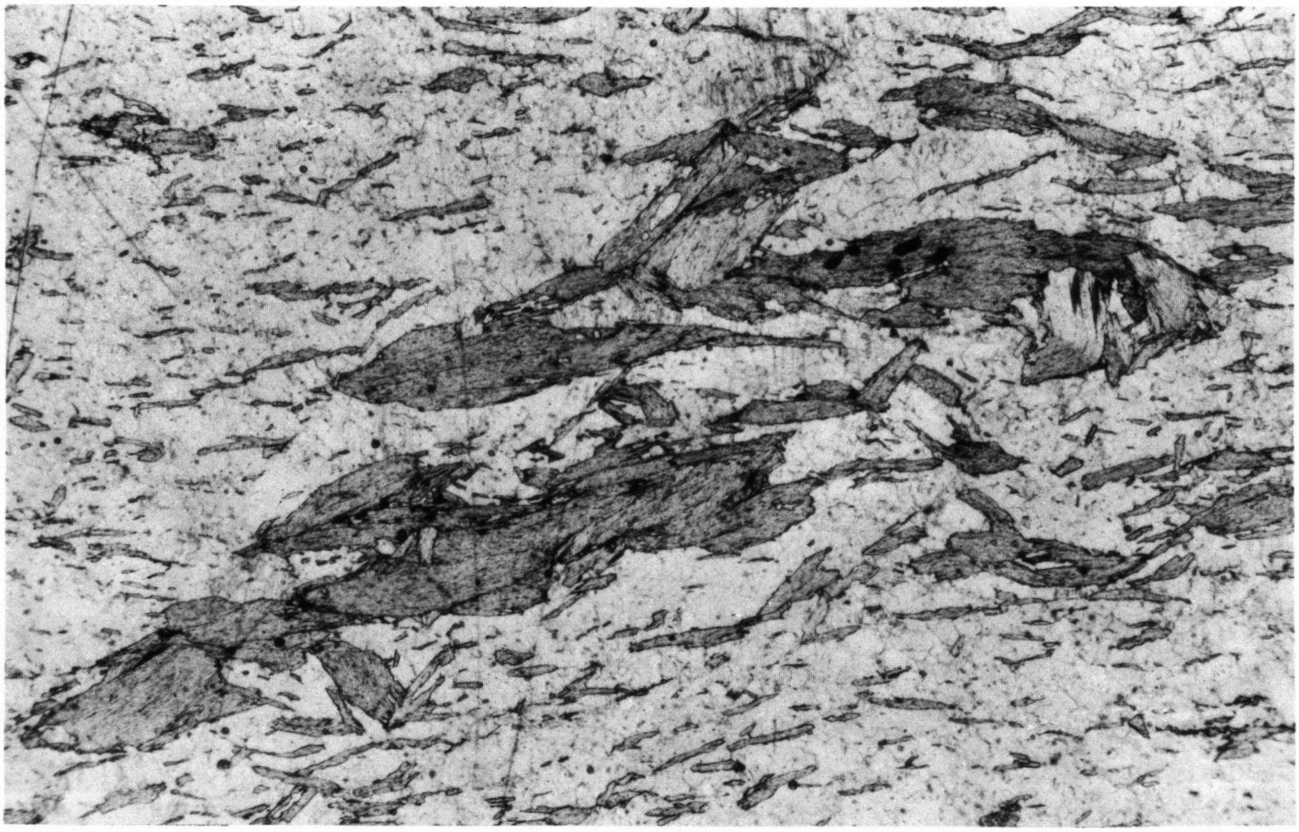


Fig. 3-14. The hinge area of an F_2 similar-style fold in the Permotriassic quartzites (Mittaghorn). The smaller micas have a better preferred orientation in S_2 than the larger ones. S_2 is almost parallel to the long side of the photomicrograph. 25×, parallel polarizers. The thin section was made available by H. J. Zwart.

poorly developed preferred orientation of colourless micas parallel to S_2, which improves with increasing tightness of the folds.

Thin mica-rich layers (up to 80% mica) in tight F_2 folds with an S_2 axial-plane schistosity, contain aggregates of micas with low-angle grain boundaries that are parallel to (001) of one of the adjoining micas (Fig. 3-14). Differing undulose extinction on either side of these boundaries demonstrates that they are movement planes. The aggregates and isolated micas (also undulose) have slightly sigmoidal outlines, pointed ends, and they are separated or almost separated from each other by quartz. Evidence for boudinage comes from micas connected by a thin string of strongly deformed mica (Wilson, pers. comm.). The (001) planes of the micas are oriented parallel to the axial plane of the fold, or intermediate between the axial plane and folded S_1.

In mica-poor layers (20–30% mica) the grain size of the colourless micas is smaller (see also 2.4), and they have a preferred orientation with (001) parallel or subparallel to the axial plane of the fold. These micas too, show undulose extinction. In both cases, the microstructure of the quartz is formed by subgrains, presumably formed during F_2, in older grains.

Discussion. – As in the para- and orthogneisses of the Saas area, undeformed and newly grown colourless micas with a strong preferred orientation in S_2 are conspicuously absent, and all micas show the effects of deformation, although F_3 folding was insignificant in the Mittaghorn area. S_2 occurs in close-tight folds which implies that the micas have been individually reoriented, and evidence for this is the sigmoidal outline of the micas, and their frequent preferred orientation intermediate between S_1 and S_2. The observations indicate that the grain size of the colourless micas has been important, for the smaller micas possess a better preferred orientation than the larger ones (see also Fig. 3-13). Reorientation of individual micas seems to be possible only when the rock has a certain minimal quartz content, as no such reorientation took place in layers consisting exclusively of colourless mica.

F_2 and S_2 east of the Saas area. – Other characteristics remaining much the same, the development of polygonal arcs improves towards the east (i.e. in the direction of increasing temperature), where they also form in layers consisting exclusively of colourless mica, but of no more than a few micas thickness. Polygonal arcs were not formed in very tight to isoclinally folded micas.

A Mesozoic quartz-rich rock from the Pizzo Montalto, found as a loose specimen above Alpe Meri Superiore, contains wavy S_2 planes and presents a poor preferred orientation of colourless mica in S_2. The micas show undulose extinction and have been boudinaged. Tiny, presumably newly grown quartz grains have appeared in S_2.

3.5.3. *Calcareous schists*

Isoclinal F_2 folds from calcareous schists show polygonal arcs in thin mica-rich layers, where the micas are separated or almost separated by carbonate. An S_2 axial-plane schistosity is not particularly frequent, but of some significance in the calcareous schists associated with the Mittaghorn Synform. In the calcareous schists exposed by the lake at Alpe Cheggio, a microscopic S_2 is formed by very tightly folded S_1 that appears to be defined by (110) of minute amphiboles.

3.5.4. *Antrona amphibolites. The F_2 folds at the Alpe Pasquale*

F_2 folds. – Their most prominent aspect is the preferred orientation of amphibole c-axes parallel or subparallel to the F_2 fold axes. The amphiboles that deviate strongly from this orientation show polygonal arcs, or, in rare cases, undulose extinction and curved outlines. The plagioclases are undulose and contain pericline twins. Numerous zoisites, whose outlines are suggestive for deformed hypidiomorphic crystals, consist of several separate grains with slightly different optical orientations, and these are suspected to be subgrains, formed in older zoisite single crystals.

S_2 schistosity. – In some F_2 folds (Alpe Pasquale, Passo di Fornalino, northern limb of the Antrona Syncline) an axial-plane schistosity is defined by aggregates of amphibole that have a preferred orientation parallel to the axial plane. The c-axes of the amphiboles are parallel to the F_2 fold axes. The aggregates are separated by strongly undulose xenoblastic plagioclase grains with pronounced pericline twinning and a dimensional preferred orientation parallel to S_2.

In the lower reaches of the T. Asinera an S_2 schistosity occurs in fine-grained amphibolites. It consists of a tectonic layering with a thickness of 0.2 mm, and defined by amphibole-rich and plagioclase-rich layers with locally large porphyroblasts of plagioclase. The amphibole-rich layers are formed by hornblende needles whose c-axes parallel the F_2 fold axes. The plagioclase between these layers contains curved inclusion trails representing folded S_1, defined by amphiboles with a smaller grain size than those outside the plagioclase. The S_2 planes curve round the plagioclase porphyroblasts; these and the finer-grained plagioclase between the amphibole-rich layers have undulose extinction, indicating that this structure is of early syn-F_2 age.

F_2 folds at the Alpe Pasquale. – From east to west, slightly to moderately plunging close-tight F_2 folds with a poorly defined S_2, change into open-close, steeply plunging to vertical folds with an S_2 axial-plane schistosity that is readily observed in the field, and microscopically defined by amphiboles with subparallel c-axes making moderate angles with the F_2 fold axes. The slightly to moderately plunging, and the steeply plunging to vertical folds do not significantly differ in fold style. F_3 folding cannot have caused the change in orientation (see 5.3.5).

The orientation in space of the amphiboles in the steeply plunging to vertical folds does not differ much from the amphibole-defined L_2 in the eastern part of the Alpe Pasquale, which is parallel to the F_2 fold axes. From a regional point of view, the latter have their normal orientation and parallelism between the F_2 fold axes, and the L_2 mineral lineation occurs everywhere else in the present area. Therefore, the steeply plunging to vertical F_2 folds in the western part of the Alpe Pasquale are abnormal and do not fit in the regional picture of F_2 folding. As a consequence, it is unlikely that during F_2 folding the orientation of the F_2 finite strain elements was much different from the remainder of the present area. Rather it is believed that before F_2 the orientation of the compositional layering and S_1 was already different, and that the changing orientation of the F_2 folds at the Alpe Pasquale must be explained by their deforming a curved surface which towards the western part of the Alpe Pasquale obtained a steeper westerly dip. This also explains why the steeply plunging to vertical folds are less tight than the slightly to moderately plunging folds; the greater the angle between the would-be axial direction of the folds (as imposed by the stress conditions) and the surface to be folded, the more open the resulting folds become (Ramsay, 1967). The direction of the L_2 amphibole lineation was apparently unaffected by the initially differing orientation of the compositional layering and S_1, which resulted in a changing angle between the amphibole-defined L_2 and the F_2 fold axes (Fig. 3-15).

The preceding discussion shows that in the steeply plunging to vertical folds two contemporaneous lineations with different orientations may coexist. They are (1) the amphibole-defined L_2 making an angle with the F_2 fold axes, and (2) an L_2 crenulation that parallels the F_2 fold axes. Unfortunately, a good example of both lineations occurring in one exposure could not be found at the Alpe Pasquale, largely because a crenulation

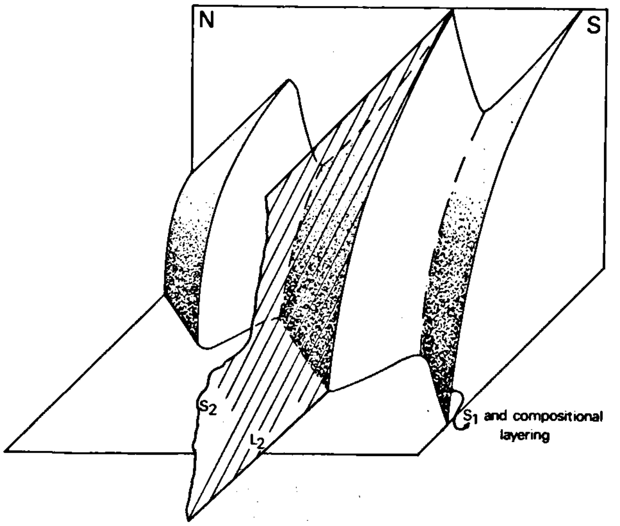


Fig. 3-15. The relation between the L_2 amphibole lineation and the F_2 fold axes at the Alpe Pasquale. The closure of the Antrona Syncline lies to the north. For explanation see text (3.5.4).

lineation is poorly developed in the amphibolites, and micaschists containing a vertical crenulation lineation (close to the huts of Alpe Pasquale) are devoid of amphibole.

3.6. THE L_2 LINEATION

L_2 occurs as a frequent crenulation lineation in all schistose rocks in the present area. It is also found as the intersection between S_1 and S_2 where the latter is well-developed. L_2 is always parallel to the F_2 fold axes, also when it is a mineral lineation, but with the exception of fold 8 (Rottalbach, see 5.2.2) and part of the Alpe Pasquale area (see 3.4.5). The L_2 mineral lineations in the various rock units can be specified as follows:

Basement rocks and Palaeozoic cover rocks. – In the paragneisses of the Saas area, L_2 occurs infrequently as an amphibole preferred orientation, and as such it is present only in the epidote-amphibolites. It is, however, common in the eastern part of the Furgg zone, and very frequent in the Camughera–Moncucco Complex where the amphibole preferred orientation is the dominant L_2 lineation in all amphibole-bearing rocks.

In the strongly linear orthogneisses of the Saas area, L_2 is a mineral lineation defined by elongate domains (0.5–5 mm long) of feldspar and/or quartz. These domains are separated by the S_1 and S_2 planes where the colourless micas are concentrated. Microscopically the lineation is a direction of mechanical stretching of existing minerals and mineral aggregates, not a direction of preferred growth. This is demonstrated by the deformed aspect and the undulose extinction of the feldspar/quartz domains, and by boudinaged micas in S_1 and S_2. In thin sections parallel to L_2, the S_1 and S_2 planes are indistinguishable, and it could be that in these linear rocks, the S_1 planes were not passive during F_2 deformation.

This microstructure is almost absent in the orthogneisses of the Camughera–Moncucco Complex, probably because syn- to post-F_2 recrystallization has been more important there (see 6.11).

Mesozoic cover rocks. – In the Permotriassic quartzites and other Mesozoic quartzitic rocks, an L_2 mineral lineation is formed by quartz rods or elongate quartz-rich domains that may not be clearly separated from their surroundings. In the Permotriassic quartzites in the eastern slopes of the Mittaghorn, length/width ratios of colourless micas have been measured in thin sections perpendicular to S_1, some parallel and others perpendicular to L_2. The average of fifty micas was taken in each section. In one specimen, the ratios were 5.0 perpendicular to L_2 and 7.9 parallel to L_2, and in another specimen the ratios were 3.3 perpendicular to L_2 and 9.3 parallel to L_2. This indicates that the colourless micas have a dimensional preferred orientation parallel to L_2 and hence form a mineral lineation.

In the amphibole-bearing calcareous schists, the amphiboles form an L_2 mineral lineation. The L_2 lineation in the Antrona amphibolites has been described previously (see 3.5.4).

CHAPTER 4

THIRD GENERATION AND YOUNGER STRUCTURES

4.1. THIRD GENERATION STRUCTURES IN THE SAAS AREA

Two contemporaneous (see below) sets of folds, that postdate F_2, have been called F_3 and F'_3.

F_3 folds. – They have refolded the first and the second generation structures. Apart from the area west of the Mälligagletscher (see below), their axes are parallel to, or make a small angle with, the F_2 fold axes and L_2. Interference patterns (Figs. 4-1, 4-2) are of Type 3 (Ramsay, 1967). Most F_3 folds are open-close folds of parallel style (Ramsay, 1967) from a few metres amplitude down to microscopic size, without axial-plane schistosity, and forming a crenulation lineation in micaceous rocks. Larger F_3 folds, with amplitudes of approx. 45 m, occur southwest of the Rotblattgletscher, and one of these is isoclinal (Fig. 4-1). The geometry of this fold implies that it terminates against a dislocation surface, which is not an uncommon feature. The axial direction of an F_3 fold west of the Mälligagletscher (Enclosures 1 and 2) makes a greater angle than usual with the F_2 fold axes whose orientations have been changed; they have an abnormal plunge of up to 65° southwesterly.

The F_3 folds are generally symmetrical, with vertical axial planes, or alternatively have an asymmetry corresponding to an antiform in the south; in that case their axial planes dip steeply north to northwestwards.

F'_3 folds. – They postdate F_2 because they have refolded an S_2 schistosity in the Triftgrätji. With the following exceptions, they have the same properties as the F_3 folds:

1. They do not form folds with more than a few metres amplitude.
2. Their axial planes dip around 50° to the south or southwest.
3. Their asymmetry always corresponds to an antiform in the north.

Relations between F_3 and F'_3 folds. – Both sets of folds have similar properties and were formed after F_2, which indicates that they have approximately the same age.

Fig. 4-1. An F_3 fold (Fold 18, Enclosure 2) refolding isoclinal F_2 folds in orthogneisses of the Portjengrat zone above the Rotblattgletscher. The triangular mark indicates the location of Fig. 4-2. The exposure is about 50 m high.

Fig. 4-2. Refolding of isoclinal F_2 folds by open F_3 folds. The exposure is about 2.5 m high. See Fig. 4-1 for its location.

Moreover, they do not refold each other, and their mutual relations resemble those of the opposite limbs of a box-fold. The arguments for this are:

1. The asymmetry of the F_3 folds is consistently opposed to the asymmetry of the F'_3 folds.
2. Each set of folds has its own characteristic direction of dip of the axial plane (Fig. 4-3). The distributions of the F_3 and F'_3 fold axes and the associated lineation share a common axis.
3. Small box-folds, on a cm-dm scale and having F_3

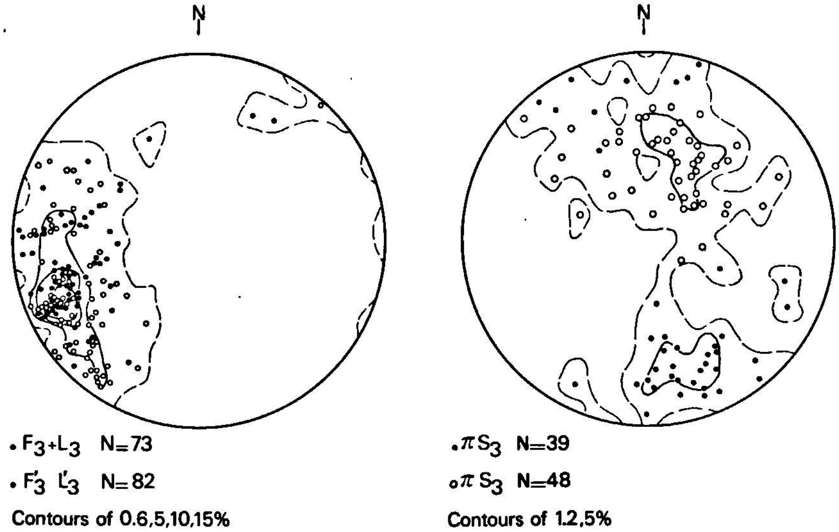


Fig. 4-3. Distribution of F_3 and F'_3 fold axes, lineations and axial planes in the Saas area. While the distribution of linear elements is not much different, the F_3 and F'_3 axial planes characteristically have an opposite direction of dip. Counting circle 1%.

and F'_3 properties, adjoin quartz-filled dislocation surfaces in the Rottalbach. Fig. 4-1 shows a polyclinal (Greenly, 1919; Turner & Weiss, 1963; Ramsay, 1967) F_3 fold that has a wide hinge zone with small curvature and a rapid change in shape of the folded surfaces along the curved axial planes. It complies to Ramsay's (1967) definition of a box-fold and shows that this type of structure could be formed in the Saas area.

The conclusion is that the F_3 and F'_3 folds were produced by a single deformation phase. They have a tendency to be spatially separated; F_3 folds occur principally in the northern and eastern part of the Saas area, and F'_3 folds occur chiefly in the northern and western part. The F_3 and F'_3 are exclusively found in areas where a shallow dip of S_1 prevails, and have little effect on the regional structural geology.

Microstructure of F_3 and F'_3 folds. – They are postcrystalline (see 6.11), and all minerals have been deformed. Some albite porphyroblasts have been rotated, and there may be elongate quartz subgrains oriented parallel to a.

4.2. THIRD GENERATION STRUCTURES IN THE ANTRONA SYNCLINE AND IN THE CAMUGHERA–MONCUCCO COMPLEX

Antrona Syncline. – F_3 and F'_3 folds occur sporadically. F_2–F_3 and F_2–F'_3 interference patterns belong to Type 3 (Ramsay, 1967), and were found north of Cresta, north of the Passo di Fornello, and at the Alpe Pasquale. The F_3 folds are open-close folds of parallel style with a maximum size of some tens of metres (Alpe Pasquale), but folds with sizes of up to a few dm are the most common. Other characteristics are the same as in the Saas area, and with the same exceptions the properties of the F'_3 folds correspond to those of the F_3 folds. North of Cresta, the shape of an F_3 fold with an amplitude of 20 m changes in profile from close to tight, and on the tight side it is bounded by a dislocation surface.

Camughera–Moncucco Complex. – F'_3 folds are absent but F_3 folds are frequent in the Camughera zone and in the northwestern part of the Moncucco zone, where a major F_3 fold (the Brevettola Antiform, see 5.4.1) was found, and large-scale refolding of S_2 took place. Most of the F_2–F_3 interference patterns are of Type 3, but the angle between the F_2 and F_3 fold axes can be larger, up to 35° as at Alpi Sogno. The L_3 crenulation lineation occurs throughout the Camughera–Moncucco Complex.

The F_3 folds are open-close parallel folds or chevron (Turner & Weiss, 1963) folds without axial-plane schistosity. The wave-lengths of first order parasitic folds on the Brevettola Antiform range between tens of metres (Val Brevettola) and decimetres (Alpi Sogno). The lower order parasitic folds may occur exclusively in the short limbs of first order parasitic folds.

Microstructure of the F_3 folds. – Although F_3 folds are generally postcrystalline, polygonal arcs of colourless mica have been found in the vicinity of Vagna. A preferred orientation of elongate quartz subgrains, approximately parallel to the a-direction of the F_3 folds is not unusual.

4.3. YOUNGER STRUCTURES

Saas area. – West-facing, open kink-folds, with their axial planes spaced at distances between a few cm and one dm, occur in profusion in the orthogneisses of the Almagellertal. Their axes plunge northwest to northeast and their axial planes dip northeast (Fig. 4-4). They are associated with tension fissures, often of sigmoidal shape, and containing chlorite, quartz and albite. The kink folds represent the last folding phase (F_k) in the Saas area. Occasional vertical faults presumably postdate the kink folds.

The Antrona Syncline and the Camughera–Moncucco Complex. – A late generation of folds, postdating F_3, and characterized by subhorizontal axial planes, occurs occasionally in the Antrona Syncline and rarely also in the Monte Rosa zone. The size of the folds ranges from a few cm or a few dm (Alpe Pasquale) to several metres at the Passo del Fornalino, where they refold F_3 folds. They represent a postcrystalline, and regionally insignificant fourth folding phase (F_4).

The last deformation phase formed kink folds, not necessarily synchronously with F_k in the Saas area, and most of them occur in the flat-lying schists close to the Simplon–Centovalli Fault, in the R. di Molezzano and in the Val Brevettola section. They usually have subhorizontal axial planes and fold axes with variable directions. Sporadic steeply plunging kink folds were found in the R. di Molezzano and east of Montescheno. At the latter location they seem to be associated with the small regional change in strike over the T. Brevettola.

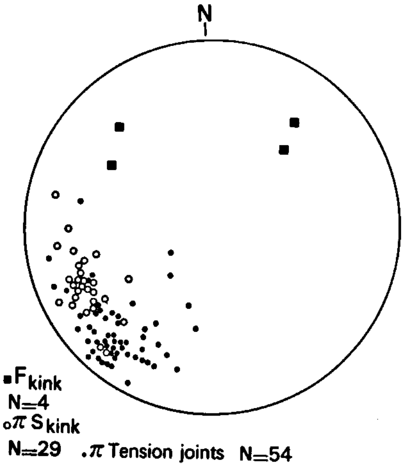


Fig. 4-4. Distribution of the structural elements of kink folds and poles to tension joints from the Saas area.

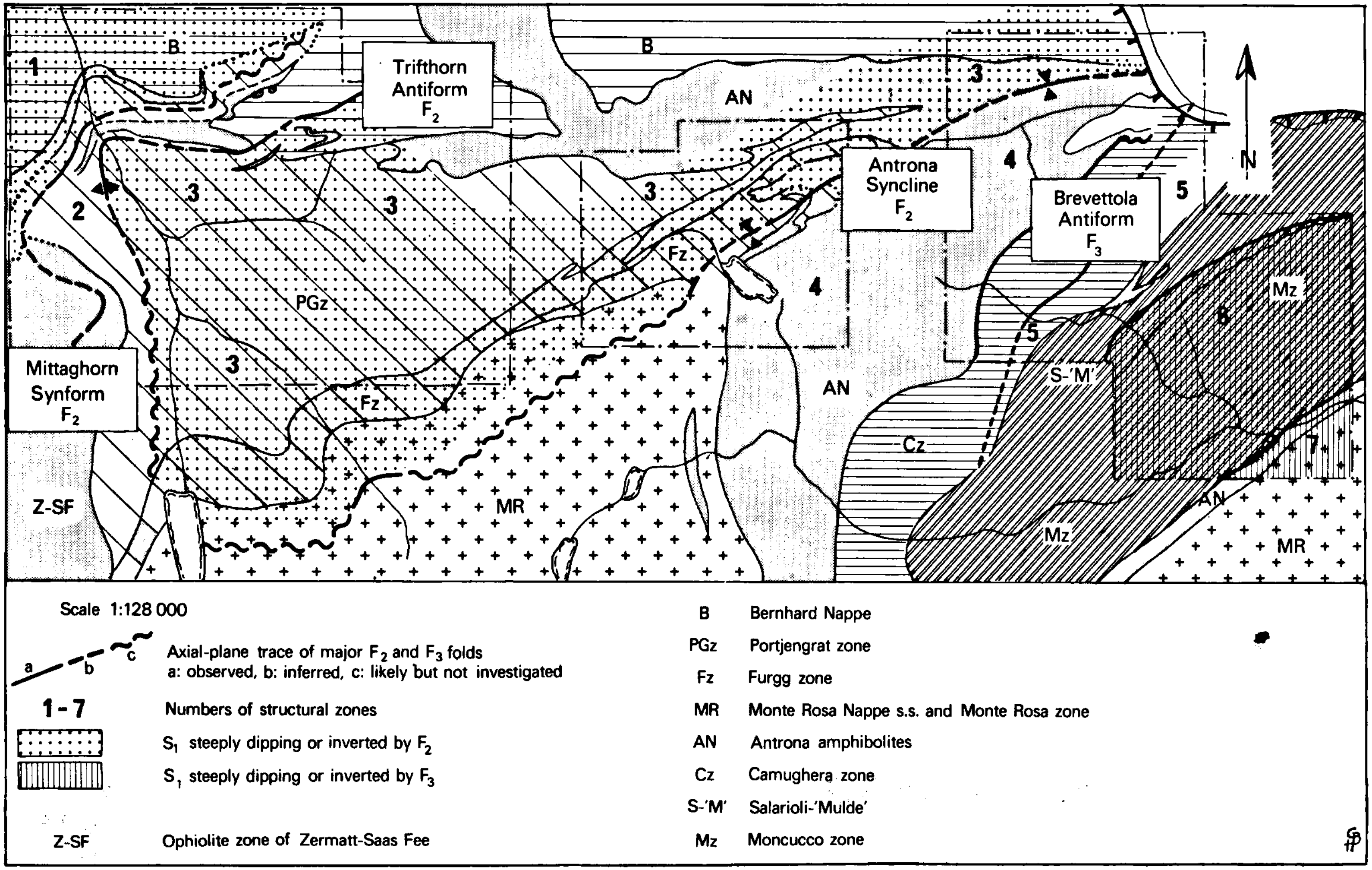


Fig. 5-1. Subdivision of the present area into structural zones. The boundaries between structural zones are defined by the axial-plane traces of major folds, the contact between the Camughera and Moncucco zones, and the contact between the Moncucco zone and the Antrona amphibolites.

CHAPTER 5

DISCUSSION OF THE MAPS AND PROFILES

5.1. ENCLOSURES 1 AND 2. STRUCTURAL ZONES

Enclosure 1 presents the structural profiles, and Enclosure 2 the maps of the present area. Index maps and profiles show the locations of maps, the relations between the Mischabelrückfalte, the Vanzone Antiform and the present area, and they summarize the major structural features. The maps and profiles of the Saas area (Map 3), the eastern part of the Furgg zone (Map 4), and the eastern portion of the Antrona Syncline (part of Map 5) were initially on a 1:10,000 scale and have been reduced to 1:25,000. The remaining areas were investigated on 1:25,000 maps.

The profiles on Enclosure 1 are perpendicular to the F_2 fold axes. Exceptions are the Alpe Pasquale and the Moncucco zone, where the F_2 folds have variable orientations and the profiles are perpendicular to the F_3 fold axes, which are nearly parallel to the normal orientation of the F_2 fold axes. With these exceptions, all profiles refer principally to the F_2 folds because F_2 is the most important folding phase, and the axial direction of the F_2 folds is reasonably constant over the greater part of the area. The F_3 and F'_3 folds are also shown, but quite a few F_1 folds are parallel to the plane of the profile and had to be omitted.

The F_2 folds are viewed in their (westerly) direction of plunge, and each profile lies west of the profile below it, with the exception of the area marked B (Enclosure 1) in the Camughera zone where the F_2 folds plunge shallowly east.

Structural zones. – Structural zones 1 to 4 (Fig. 5-1) are identical with the limbs of the major F_2 folds. Structural zone 1 is a small portion of the inverted limb of the Mischabelrückfalte, and its southern boundary is the axial-plane trace of the Mittaghorn Synform. Structural zone 2 is an area of non-inverted S_1 between the axial-plane traces of the Mittaghorn Synform and the Trifthorn Antiform. The latter's axial-plane trace is the boundary with structural zone 3, where inverted S_1 predominates. Structural zone 3 comprises the greater part of the Saas area, and its southern boundary lies south of the Furgg zone. East of the Saas area, where its southern boundary is formed by the axial-plane trace of the Antrona Syncline, it contains the Portjengrat zone, most of the eastern part of the Furgg zone, and the northern limb of the Antrona Syncline. Structural zone 3 is continuous from the Saastal to the Simplon–Centovalli Fault (see 5.3.3).

Structural zone 4 is the non-inverted southern limb of the Antrona Syncline, and it includes a small portion of the eastern part of the Furgg zone. Its boundary with structural zone 3 is the axial-plane trace of the Antrona Syncline, and from structural zone 5 (see below) it is separated by the contact between the Antrona amphibolites and the Camughera zone.

Structural zones 5 and 6 (Fig. 5-1) in the Camughera–Moncucco Complex have been defined according to the orientation of S_1 in the F_3 Brevettola Antiform. Structural zone 5, corresponding approximately to the Camughera zone, is the closure of the Brevettola Antiform. The northern contact of the ultramafic body of Montescheno is the boundary with structural zone 6, and structural zone 5 therefore includes a portion of the Moncucco zone. Structural zone 6 corresponds to the remaining part of the Moncucco zone, and it lies in the southern, overturned limb of the Brevettola Antiform. It is separated by the contact between the Moncucco zone and the Antrona amphibolites from structural zone 7, which comprises the

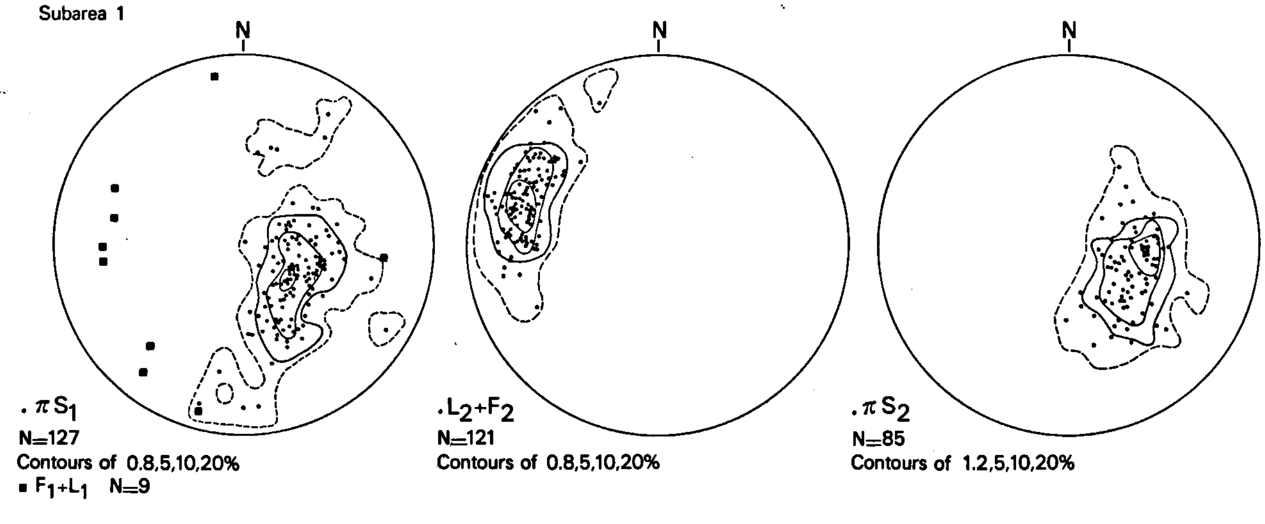


Fig. 5-2. Distribution of planar and linear structural elements in subarea 1 (for its boundaries, see Enclosure 2). Counting circle 1%.

thin layer of Antrona amphibolites and the Monte Rosa zone.

5.2. SAAS AREA (MAP 3)

5.2.1. *Structural zones 1 and 2*

Structural zone 1. – Bearth (1964) and Zwart (in prep.) demonstrate that it belongs to the inverted limb of the Mischabelrückfalte, and this is confirmed in the present paper. In the area of F_3 folding west of the Mälligagletscher, F_2 folding is intensive, and according to Nieuwland (1975), the axial-plane trace of the Mittaghorn Synform continues there. This continuation does not figure on Enclosures 1 and 2.

Fold 1 – F_2 (Mittaghorn Synform). – This synform was described by Güller (1947) and Bearth (1964, 1967) who considered it as the frontal hinge of the Ophiolite Nappe, and which is therefore actually an F_2 fold (see also 1.3). Its closure is well-exposed in the southeastern face of the Mittaghorn. The Mittaghorn Synform is a disharmonic fold, and its properties have been previously described in 3.3.1–2 and in 3.4.1. Its axial plane is contained between the two intercalations of calcareous schists east of Saas Fee, that before F_2 were probably situated at an identical stratigraphical position. These intercalations disappear under a cover of moraine and ice, and it is therefore not certain that they were initially one layer.

Faulting parallel to S_1, associated with folding of late but uncertain age, took place in the calcareous schists, and good examples of it occur at 1625 and 2400 m in the northern intercalation. Folds adjoining the faults die out at some distance from the fault plane, and when they are present on either side of the fault, the fold profiles do not match. It appears that the folds terminate laterally in a fault. Occasionally S_1 on one side of the fault makes an angle of up to 25° with the fault plane that is parallel to S_1 on the other side. The fault itself may contain irregularly shaped carbonate lenses and be slightly deformed. These structures are possibly of F_3 age, because of the association between folds and dislocation surfaces and the box-like geometry of some of the folds (see 4.1). Similar structures, associated with the 'rétrocharriage' phase in the Western Alps, were described by Caron (1973).

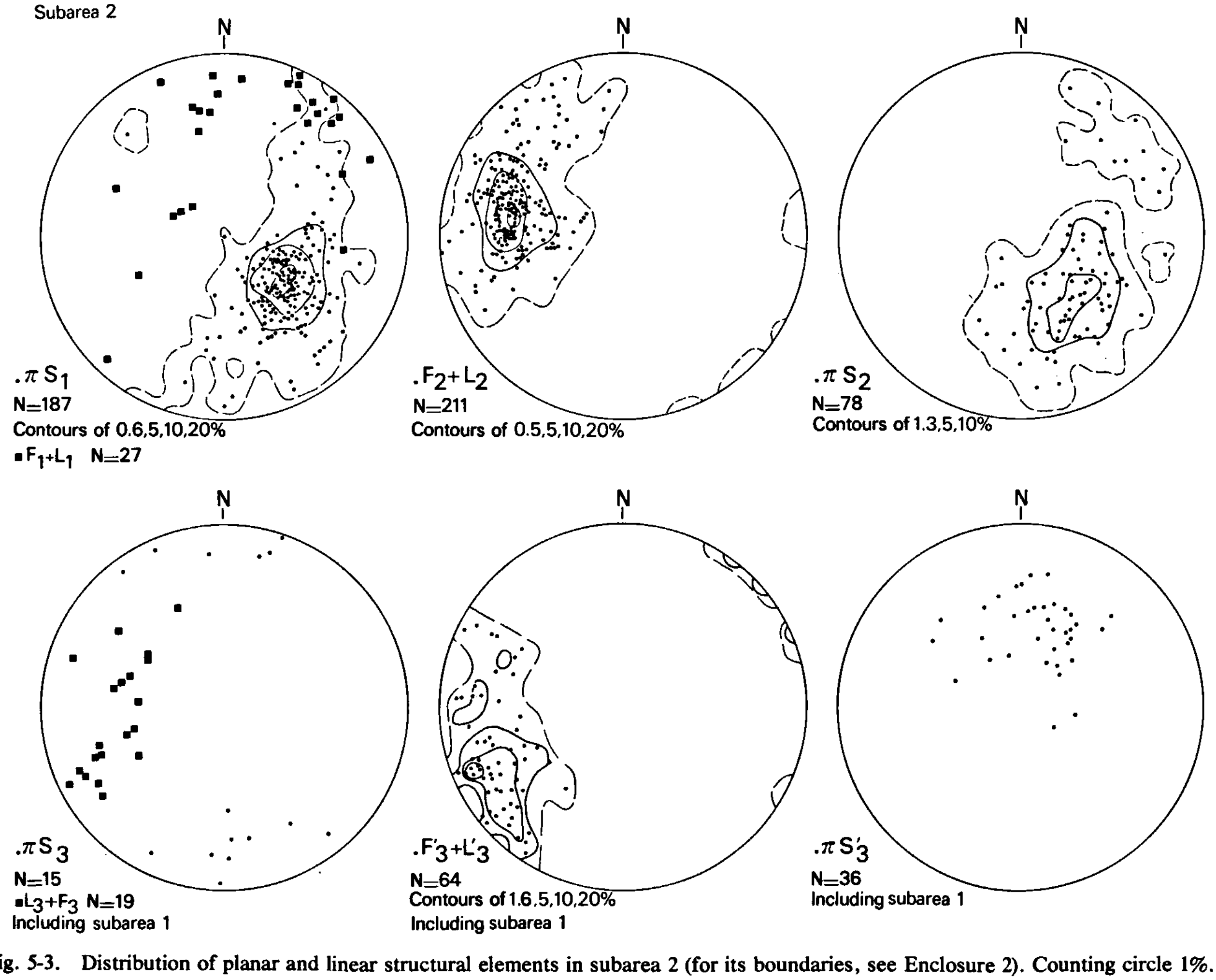


Fig. 5-3. Distribution of planar and linear structural elements in subarea 2 (for its boundaries, see Enclosure 2). Counting circle 1%.

Adjoining the contacts between the calcareous schists and the paragneisses, and in particular both contacts of the northern intercalation, occur slices of paragneiss enclosed by calcareous schists and vice versa. No relation with F_2 or later folding could be established, and these structures are likely to have been formed during the nappe movements. Nappe movements and F_1 folding are responsible for the structural complications in the southern intercalation of calcareous schists, where it terminates between Saas Fee and the Saaser Vispa. The enclosures do not show the geology in full detail.

In the western slope of the Trifthorn, glacial erosion has cut through the axial plane of a first order parasitic F_2 fold. This axial plane disappears in all directions below the topographic surface, and the axial-plane trace consequently frames a structural window in the non-inverted limb of the fold through which the inverted limb can be observed.

Fold 2 – F_2 (Trifthorn Antiform). – The axial-plane trace of the Trifthorn Antiform is well-exposed between Saas Fee and the Vispa, and in the opposite slope it has been slightly refolded by F_3 folds. It is contained between the recumbent calcareous schists capping the Grundberg, and a thin layer of Triassic rocks (Bearth, 1957a, map explanation) descending in the Steintälli, and it is well-exposed below the Trifthorn from where it can be traced into the ridge separating the Steintälli and the Rottal. The Trifthorn Antiform becomes progressively tighter from the Rottal towards Saas Fee, and it is another disharmonic fold with limbs that are convex towards the axial plane, of the kind described in 3.3.1 and to which the Mittaghorn Synform also belongs. The Mittaghorn Synform and the Trifthorn Antiform tighten in opposite directions (Enclosures 1 and 2), and their change in geometry is accompanied by a similar change in the geometry of their parasitic folds.

5.2.2. *Structural zone 3*

Folds 3 and 4 – F_2. Their relations with the Trifthorn Antiform. – The axial-plane trace of fold 4 passes from orthogneiss into paragneiss, and this presumably occurs below the Steintälli, where the contact is covered by moraine.

The shallow to moderate dip of S_1 north of the Trifthorn Antiform does not reappear south of it, and antiform 4 is much more difficult to locate and less well-defined than antiform 2. Antiform 2 is therefore presented here as the closure of the Trifthorn Antiform. However, it might well be that their relative importance is not much different and that both are parasitic folds in the closure, separated by synform 3.

Fold 5 – F_2. – Presents no particulars.

Folds 6, 6′, 7 and 7′ – F_2. – Folds 6 and 7 (together forming the Zwischbergen parasitic fold) could well be continuous with folds 6′ and 7′ at the end of the Almagellertal, but this could not be demonstrated because of poor exposure. The linear orthogneisses in the hinges and in the short limb of the Zwischbergen parasitic fold form the high part of the Dri Hörlini ridge. In the Mesozoic rocks at the Zwischbergen pass, cm–dm sized F_3 folds occur in the closures of folds 6 and 7.

Fold 8 – F_2?. – It is exposed in the Rottalbach between 2510 and 2580 m altitude, and its main characteristic is the high angle between the westerly plunging L_2 mineral lineation and the northeasterly plunging fold axes. Fold 8 has folded a schistosity parallel to a compositional layering and it contains an axial-plane schistosity; these features form evidence for an F_2 fold but in the field no explanation was found for the divergence between the fold axes and L_2.

Should fold 8 have an F_1 age, the northeastern plunging fold axis would correspond to the normal orientation of F_1 folds in that area, and the angle with L_2 is explained. In that case, the axial-plane schistosity must be S_1 and the folded schistosity an older plane. A pre-S_1 schistosity plane has, however, not been found in

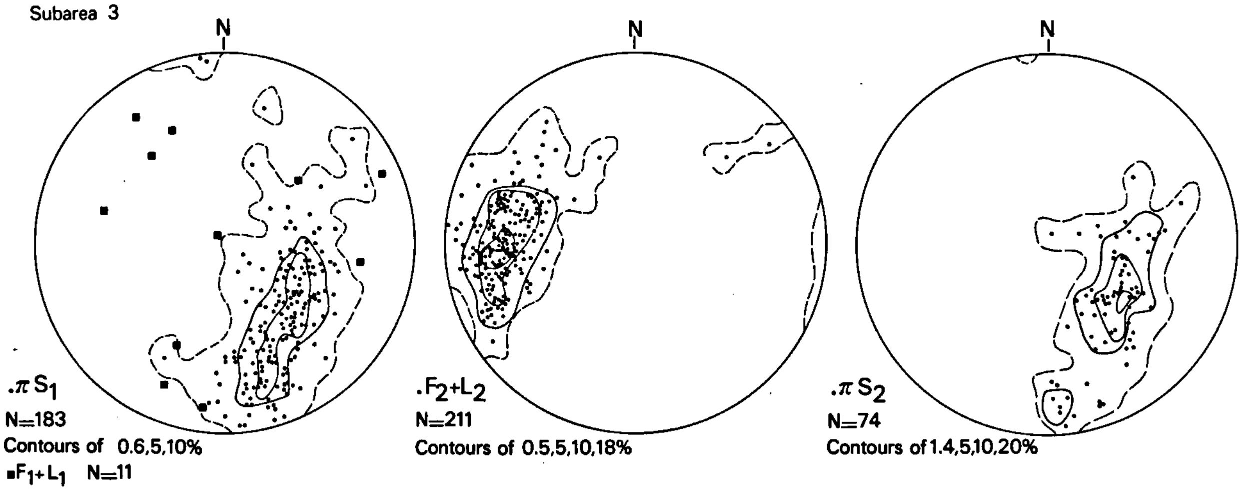


Fig. 5-4. Distribution of planar and linear structural elements in subarea 3 (for its boundaries, see Enclosure 2). Counting circle 1%.

the Saas area and for this reason fold 8 is not likely to be F_1. It is tentatively assigned to F_2, though with some reservation because the reason for its abnormal features is unknown and no other F_2 folds of this kind were found. It may have been a relatively open F_1 fold, that was subjected to renewed deformation by F_2.

Fold 9 – F_2. – It has folded a syn-nappe movement zone of calcareous schists and prasinite lenses, traceable for a short distance only.

Fold 10 – F_1 refolded by F_3. – Presents no particulars.

Folds 11 and 12 – F_2. – Synform 11 is covered by scree and could not be located in the Portjengrat; antiform 12 might be situated below the summit of the Portjengrat or south of it, where S_1 curves into homogeneous orthogneisses that could be linear gneisses concealing an antiform. F_2 folds with a moderate, northeasterly plunge occur between these hinges, and this could explain why folds 11 and 12 were not seen, since they would plunge away from the observer. This is an area of intensive F_1 folding.

The formation of the steep dip of S_1 in the Portjengrat. – F_3 folds are absent between the F_2 folds 7 and 14 in the Portjengrat, where steeply dipping inverted S_1 prevails, but they reappear in shallowly dipping, non-inverted S_1 between the F_2 folds 14 and 15 (see below). It has been assumed that the formation of the F_3 folds depends on the angle between their axial planes and the surfaces to be folded (see also the occurrence of the F_3 folds in the Zwischbergenpass area). A small angle is not favourable for the formation of parallel folds but would tend to produce movements along the S_1 schistosity or the plane of the compositional layering instead. Evidence for such movements could be the changing orientation of the axial planes of the F_2 folds contained by micaschists, where the folds adjoin the contact with orthogneisses. This change is not related to the dying out of the folds, and it was occasionally encountered in steeply dipping S_1 in the Portjengrat.

At present, the orientation of S_1 between folds 7 and 14 is almost the same as the orientation of the axial planes of F_3 folds elsewhere. Considering the remarks made previously, the absence of F_3 folds between folds 7 and 14 can be explained when it is accepted that the steep dip of S_1 existed before F_3. This implies that the steep dip of S_1 in the Portjengrat results from F_2 folding. It is believed here that the steep dip of S_1 in the Portjengrat, and in general, the prevailing steep dip of S_1 in structural zone 3 have been produced during F_2 deformation, and forms the more steeply dipping, inverted limb of the Trifthorn Antiform, in contrast to structural zone 2, the less steeply dipping non-inverted limb. F_3 and F'_3 folds are frequent in structural zone 2, demonstrating the regionally different orientation of S_1 in both limbs before F_3 folding took place.

Folds 14 and 15 – F_2, fold 18 – F_3. – Folds 14 and 15 enclose an area of F_2–F_3 refolding. On the northeastern side of the tongue of the Rotblattgletscher, the axial-plane trace of synform 14 is covered by moraine, and the axial-plane trace of antiform 15 passes from layered orthogneisses into massive orthogneisses below the Mittelrück, folding the contact to an almost recumbent antiform. The axial planes 14A and 15A of a lower order parasitic F_2 fold meet near the glacier where it dies out. Local F_3 refolding is caused by the F_3 fold 18, less prominent here than at the other (southwestern) side of the glacier.

Fold 16 – F_2. – The dip of its axial plane decreases downwards, and the folds 15 and 16 are believed to cancel out at depth, for their axial planes are converging.

Fold 17 – $\tilde{F}_2$. – The axial plane of this antiform and the axial plane of fold 16 are 350 m apart on the Italian side of the Mittelrück. They approach each other towards the summit where they are no more than 100 m apart and it is likely that they meet at a higher level.

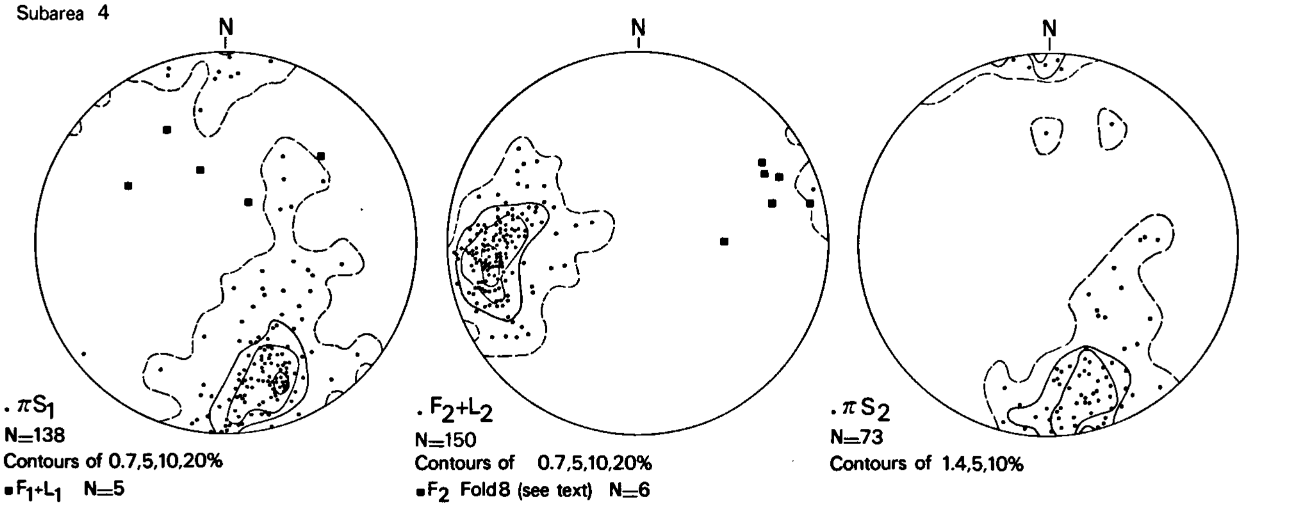


Fig. 5-5. Distribution of planar and linear structural elements in subarea 4 (for its boundaries, see Enclosure 2). Counting circle 1%.

Fold 18 – F_3. – See 4.1. In the ridge between the Känzelti and the Sonnighorn, F_2 folds are numerous in the orthogneisses, but their asymmetry is uncertain. Just west of the Sonnighorn, the axial planes of F_2 folds pass from orthogneisses into paragneisses, and the contact between both rock units has been folded into symmetrical close folds with an almost vertical enveloping surface. The ridge contains another F_3 fold about 200 m northwest of the contact; it is approximately of the same size as fold 18 (this fold is not shown on Enclosure 1).

Fold 19 – F_2. – The orthogneisses north of this synform lie in inverted position, which is demonstrated by folds below the Känzelti (they are best seen from the slopes opposite Furggu). Orthogneisses in non-inverted position lie close to the Furggtal valley-floor and contain isoclinal F_2 folds whose asymmetry is difficult to see. Intercalations of paragneiss show the effect of refolding by cm-sized F_3 folds. Fold 19, which on Swiss territory synformally folds the whole orthogneiss body, loses its importance towards the east where on Italian territory the lower contact of the orthogneiss body lies in an inverted position (see below). this is shown by F_2 folds that have deformed the contact below the Sonnigpass.

Folds 20, 21 and 21A – F_2. The lower contact of the Mittelrück orthogneiss body. – The lower contact of the orthogneiss body that contains the Mittelrück, changes from its inverted position on Italian territory and on the higher northeastern slopes of the Furggtal into a non-inverted position lower in the Furggtal. This goes to show that the axial plane of an antiform (maybe the equivalent of fold 21) must leave the orthogneisses between the Sonnighorn and the Furggtal, and enter the paragneisses south of this orthogneiss body, where it probably is the antiform in the thin layer of orthogneiss shown on the section. Enclosure 1 shows that this region and the area of fold 22′ have been strongly affected by distortion due to divergence of the F_2 fold axes, introducing considerable inaccuracy in the location and shape of the lower contact of the Mittelrück orthogneiss body.

Folds 22 and 22′ – F_2. – These folds, believed to be continuous, are the antiforms nearest to the Furgg zone.

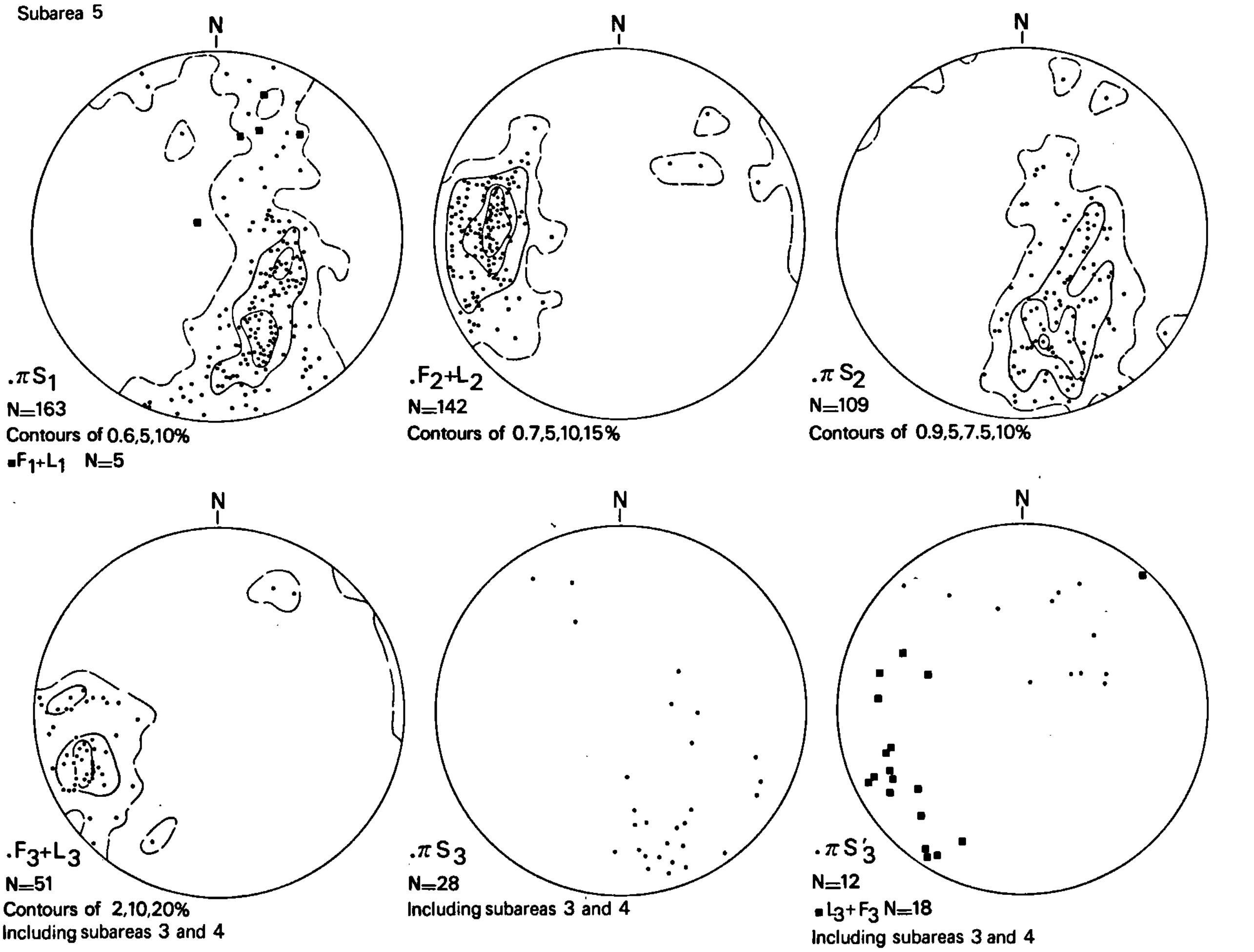


Fig. 5-6. Distribution of planar and linear structural elements in subarea 5 (for its boundaries, see Enclosure 2). Counting circle 1%.

Fold 22′ is located in a zone of isoclinal folding of more than 200 m width in the Furggtal. Its axial plane could not be located exactly, but F_2 folds with the asymmetry of the inverted limb in the Furgg zone itself, indicate the existence of this antiform. The zone of isoclinal folding must also contain a synform, confined between the antiform in the thin layer of orthogneiss (see above) and antiform 22′. It could be the equivalent of fold 21A.

5.2.3. *Stereograms*

For the presentation of the structural data in lower hemisphere equal-area projection, the Saas area has been subdivided into subareas 1–5 whose boundaries are shown on Enclosure 2. The selection of the subareas demonstrates a small regional change in orientation of the F_2 folds, and the steepening of S_1 from the northern and western parts of the Saas area towards the Portjengrat (Figs. 5-2 to 5-6).

The distribution of the structural elements in the five subareas shows little difference. The results of post-F_2 folding are most apparent in subareas 3 and 4, but only subarea 3 is affected slightly more than on a strictly local scale. The importance of F_2 folding is shown by the distribution of poles to S_1. Due to the choice of subareas, the inverted and non-inverted limbs of major F_2 folds cannot be recognized in the diagrams.

The predominant northerly to northeasterly plunge of the F_1 fold axes reflects the orientation of numerous folds in the central part of the Portjengrat zone, and is not representative for the Portjengrat zone as a whole.

5.2.4. *The southern boundary of structural zone 3; evidence for the continuation of the Antrona Syncline in the Monte Rosa Nappe s.s.*

The closure of the Antrona Syncline is, among other areas, easily observable in the eastern part of the Furgg zone (see 5.3.1–4) from where it is believed to continue into the Monte Rosa Nappe s.s. The arguments for this are:

a. The Antrona Syncline cannot be contained by the Furgg zone in the Upper Val Loranco, since F_2 folds with the same asymmetry occur on both sides of the Furgg zone. They are parasitic on a synform in the south, that must be located within the Monte Rosa Nappe s.s.

b. The axial plane of the Antrona Syncline was seen to enter the frontal part of the Monte Rosa Nappe s.s. above the Bacino Alpe dei Cavalli.

c. At the locality called Kehrenrück, in the northeastern slopes of the Furggtal and south of the Furgg zone, F_2 folds in shallowly dipping orthogneisses show that S_1 is in non-inverted position. Between the Kehrenrück and the steeply dipping Furgg zone, other F_2 folds indicate that S_1 has been inverted and therefore a large synform is present; its axial-plane trace lies approx. 700 m southeast of the Furgg zone. This proves the existence of an F_2 synform in the Monte Rosa Nappe s.s., and it is probably a part of the Antrona Syncline.

d. It has been previously supposed that the Monte Rosa Nappe s.s. should contain an important fold. Blumenthal (1952) pictures the Monte Rosa Nappe s.s. as a recumbent fold, and the internal geology of the nappe suggested the same to Bearth (1958). He, however, dismissed the idea because in the frontal part of the nappe, in the Val Loranco where the hinge should lie, old structures (pre-Alpine metamorphic and intrusive rocks; remark by this author) have been preserved. This cannot be used as an argument against the model presented here, for pre-Alpine rocks can be folded by Alpine folds.

e. Bearth (1952a) differentiated the Stelli zone in the frontal part of the Monte Rosa Nappe s.s., and defined it as a zone of strong Alpine deformation that resembles the Portjengrat zone. The Stelli zone contains the synform mentioned under (c), and it is believed to be the area where the closure of the Antrona Syncline is located.

The axial-plane trace of the Antrona Syncline has for these reasons been tentatively continued towards the west, where it reaches the Saastal close to Mattmark.

5.2.5. *Regional importance of the Trifthorn Antiform (1); its relation to the Antrona Syncline in the Mattmark area*

F_2 folds with the asymmetry corresponding to an inverted limb predominate in the lower western slopes of the Saastal, whereas the augengneiss of Saas Fee and the overlying Mesozoic rocks are in non-inverted position. Therefore, the Trifthorn Antiform continues in the paragneisses in the western slopes of the Saastal where it becomes tighter and gradually changes its orientation to a reclined fold.

Northwest of Mattmark, the Furgg zone (structural zone 3) meets the Mesozoic cover of the Portjengrat zone and the augengneiss of Saas Fee (both structural zone 2). As a consequence, an antiformal structure is present. This is substantiated by F_2 folds in the Furgg zone below the Allalingletscher, indicating an inverted limb, and by others in Mesozoic rocks south of the Mittaghorn where they indicate a non-inverted limb. Though the axial-plane trace of the Trifthorn Antiform could not be followed through the inaccessible western slopes of the Saastal, there is no doubt that it continues as a reclined fold to Mattmark. Because of its almost recumbent attitude, the axial plane of the Trifthorn Antiform swiftly approaches the proposed continuation of the axial-plane trace of the Antrona Syncline. It is supposed that they meet near Mattmark, in an area of extremely intense F_2 folding in the slopes northwest, west and southwest of the lake.

These axial planes enclose the inverted S_1 of structural zone 3, that disappears where they meet. When this happens, the Trifthorn Antiform and the Antrona Syncline terminate, and the non-inverted S_1 of structural zone 2 becomes continuous with structural zone 4, the non-inverted southern limb of the Antrona Syncline. The meeting of these axial planes is a structural necessity if the largest portion of the Monte Rosa Nappe s.s. and covering rocks are to be in non-inverted position, as is generally considered to be the case. In the model suggested here, only a small frontal part of the Monte Rosa Nappe s.s. has been inverted by F_2.

5.2.6. *Regional importance of the Trifthorn Antiform (2); its relation to the Balmahornrückfalte*
The Trifthorn Antiform is transitional between the non-inverted structural zone 2 and the inverted structural zone 3, which extends from the Saastal to the Simplon–Centovalli Fault (see 5.3.3). Structural zone 3 is the northern limb of the Antrona Syncline and at the same time the southern limb of the antiformal Balmahornrückfalte (Te Kan Huang, 1935; Blumenthal, 1952). It is also the southern limb of the Trifthorn Antiform. This makes it very likely that the Balmahornrückfalte and the Trifthorn Antiform are the same fold, a possibility that has been investigated by Nieuwland (1975) who found that the northeastern prolongation of the Trifthorn Antiform is contained in the spur between the Weissmies and the Zwischbergenpass, and is continuous with the closure of the Balmahornrückfalte.

This confirms that both folds are the same structure, and in subsequent papers the name Balmahornrückfalte should for priority reasons replace the name Trifthorn Antiform. It has not been possible to do so here, as the Enclosures and the text were in too advanced a stage of preparation when Nieuwland's results became known.

5.2.7. *Deformation of the nappe boundaries by F_2 folds. The Saas Fee-Zwischbergen boundary*
A glance at Enclosures 1 and 2 will show that there are many instances of folding of the nappe boundaries by F_2 folds, in the Saas area as well as in the area yet to be described. There is no need to go into this in detail, except for the Saas Fee–Zwischbergen boundary because of its importance as the formerly disputed boundary between the Monte Rosa Nappe s.l. and the Bernhard Nappe. The Saas Fee–Zwischbergen boundary separates the Bernhard Nappe from the Portjengrat zone, and as it crosses structural zones 1, 2 and 3, it is partly inverted and partly in non-inverted position. The boundary follows the two layers of Mesozoic calcareous schists east of Saas Fee, believed to have been situated at an identical stratigraphical level after F_1. A small part of which is at present the southern layer was later to be folded by the Trifthorn Antiform to obtain a steeply dipping attitude in the Steintälli, where it has been described by Bearth (1957a). From this occurrence of Mesozoic calcarous schists, the boundary has been traced along the orthogneiss lens in the Steintälli, where remnants of calcareous schists adjoin its northern contact. Its present structural position shows that this lens must have been displaced parallel to S_2 during F_2 deformation, demonstrating that movements parallel to S_2 introduce complications in the reconstruction of this boundary. It then continues from here along the orthogneiss lens in the Rottal, which has been folded by fold 3, and meets the contact of the orthogneisses and paragneisses in the northern slopes of the Almagellertal, becoming inverted as the result of fold 4. The boundary exposed in the Rottal, towards the Zwischbergenpass is believed to continue along this contact, which is not where Mesozoic rocks adjoin the orthogneisses (Geol. Atlas der Schweiz, Blatt Saas).

Bearth (1957a, map description) suggests that east of the Grundberg the boundary lies north of the connection suggested here, but it is believed that folds 3 and 4 have displaced it towards the south. The boundary suggested by Bearth would be possible if there were a zone of thrusting with numerous individual thrust planes rather than a single plane. Though the interlayered basement and cover rocks close to the Zwischbergen could be evidence for this, elsewhere little of the kind was found, and the nappe boundary has therefore been represented as a single plane.

From this it follows that the Saas Fee–Zwischbergen boundary can be traced between Saas Fee and Zwischbergen, with the exception of a moraine-covered stretch in the Rottal, and that notwithstanding having been folded by F_2 it can be considered as an essentially continuous boundary between the Bernhard Nappe and the Monte Rosa Nappe s.l.

5.3. THE EASTERN PART OF THE FURGG ZONE, AND THE ANTRONA SYNCLINE (MAPS 4 AND 5)

5.3.1. *Structural zone 3 in the Upper Val Loranco (subarea 6)*
This subarea is partly featured on the detailed maps, and it comprises the Portjengrat zone east of the Pizzo d'Andolla, and the Furgg zone, both until the eastern termination of the Monte Rosa Nappe s.s. whose frontal part is also included with this subarea.

F_2 folds in the Portjengrat zone are in all respects similar to those from the Saas area, but subvertically plunging F_2 folds occur in the Furgg zone near Alpe Camasco (Fig. 3-1B). In the Monte Rosa Nappe s.s., south of this locality, F_2 folds have a moderate to steep northwesterly plunge which towards the east becomes a shallow western plunge. They are parasitic on a synform in the south, and S_1 is inverted. Wetzel (1972) claims that the Furgg zone is an isoclinal synform, but should this be true it cannot have F_2 age, because F_2 folds have the same asymmetry in the Portjengrat zone as in the frontal part of the Monte Rosa Nappe s.s. Faulting has occurred in the middle ridge descending from the Cime di Pozzuoli into the Upper Val Loranco. Between this and the eastern ridge, extensive refolding of F_1 by F_2 folds occurs close to the Cime di Pozzuoli where their axial directions make a high angle.

5.3.2. *Structural zone 3: the eastern part of the Furgg zone (subarea 7)*
This is the continuation of subarea 6 east of the frontal part of the Monte Rosa Nappe s.s., and comprises the Portjengrat zone and the Furgg zone until the axial-plane trace of the Antrona Syncline, the southern limit of inverted S_1 of structural zone 3 (Fig. 5-1).

Folds 23 and 24 – F_2. – Synform 23 (already observed by Blumenthal, 1952) and antiform 24 present no particulars.

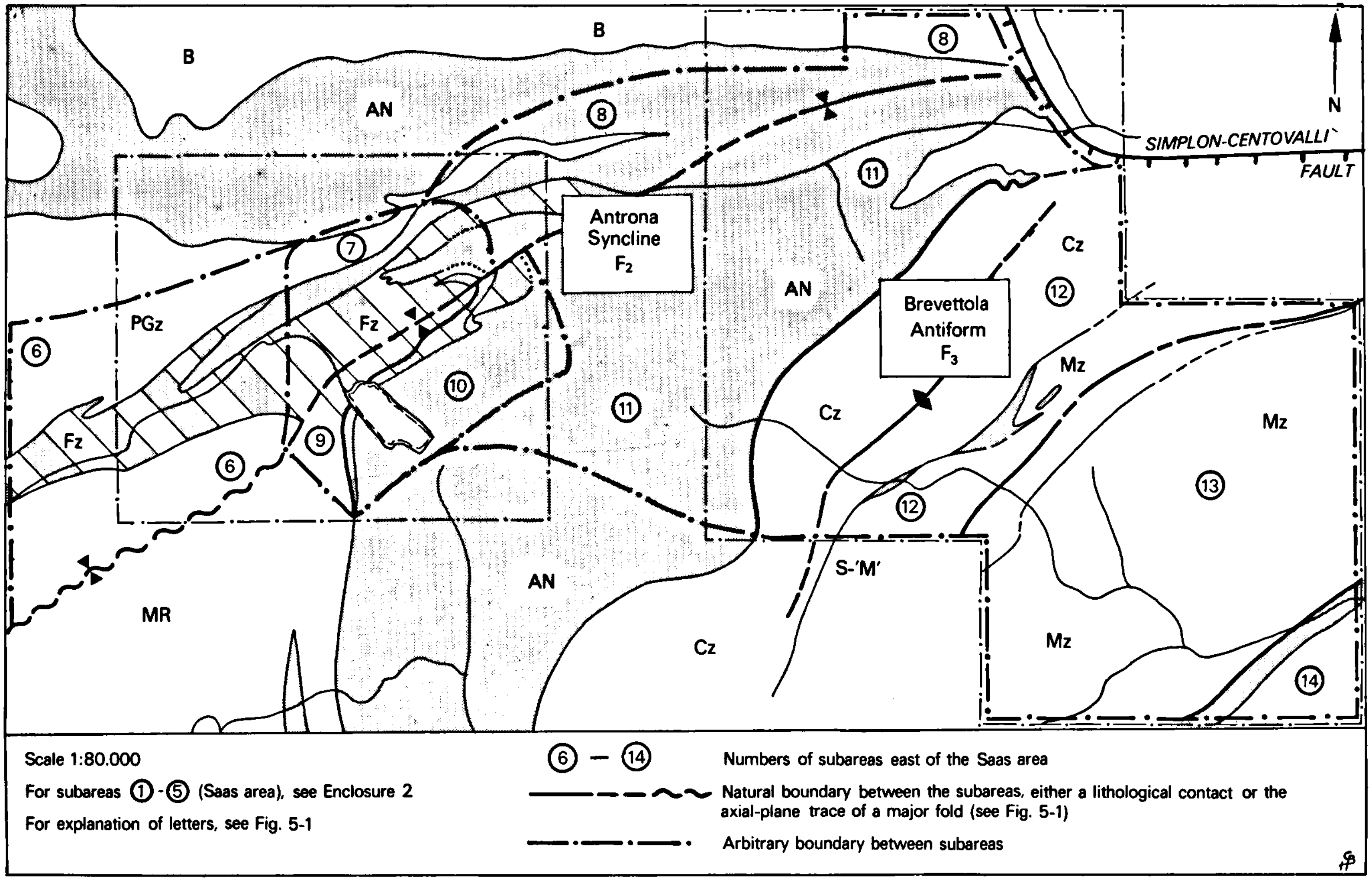
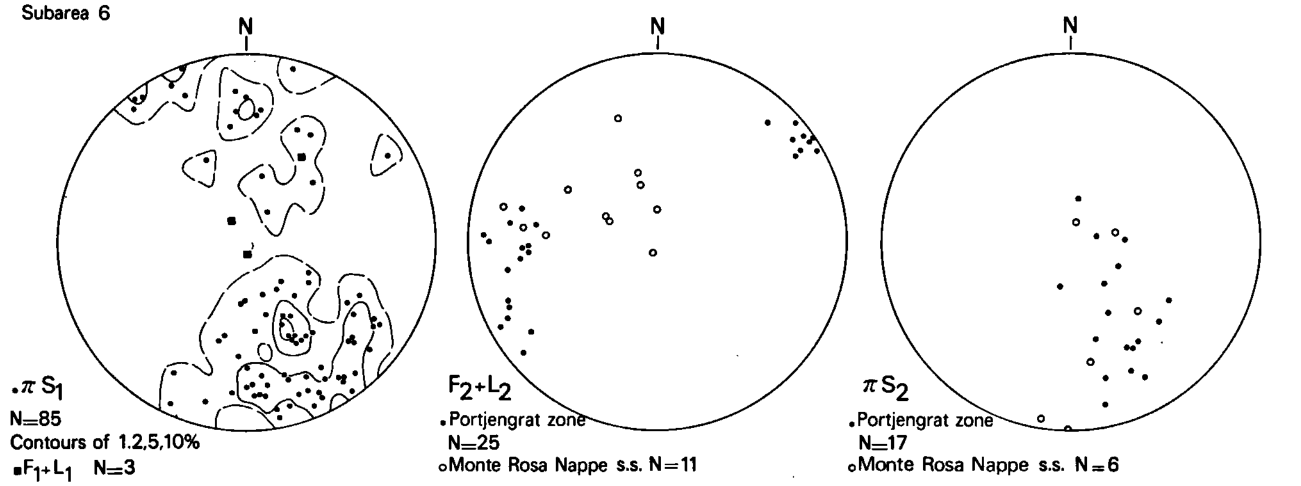


Fig. 5-7. Location map of subareas 6–14, east of the Saas area.



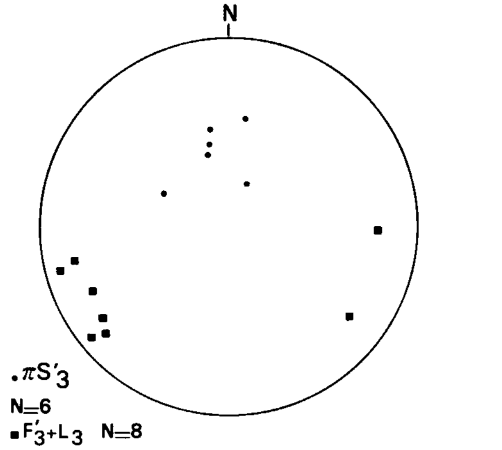


Fig. 5-8. Distribution of planar and linear structural elements in subarea 6 (for its boundaries, see Fig. 5-7). Counting circle 1%.

Fold 25 – F₂. – Is particularly well-exposed in the northwestern slopes of the Monte della Preja, where it contains many lower order parasitic folds and frequent F_1–F_2 Type 2 (Ramsay, 1967) interference patterns.

Fold 26 – F_2 (Antrona Syncline). – Its closure in the eastern slope of the Monte della Preja (see also Wetzel, 1972) is formed by two moderately plunging synforms, with a distance of approximately 30 m between their axial planes, and connected by a broken-through antiform. The Furgg zone schists and gneisses that contain this closure have a faulted contact with the serpentinites north of them. These serpentinites possess a local S_2 schistosity with vertical fold axes, and though these folds have an exceptionally steep plunge, a moderate to steep plunge is common in the eastern part of the Furgg zone in contrast to the Portjengrat zone where the F_2 folds have a shallow to moderate plunge. A moderate to steep plunge occurs also in the adjoining subareas 9 and 10, and it is a characteristic property of the eastern part of the Furgg zone and the adjoining part of the Antrona Syncline (compare the stereograms of subareas 7, 9 and 10 (Figs. 5-9, 5-11, 5-12) with those of subareas 6 and 8 (Figs. 5-8 and 5-10)).

Some moderate to shallow northeasterly plunging F_2 folds occur in alignment parallel to the general strike of

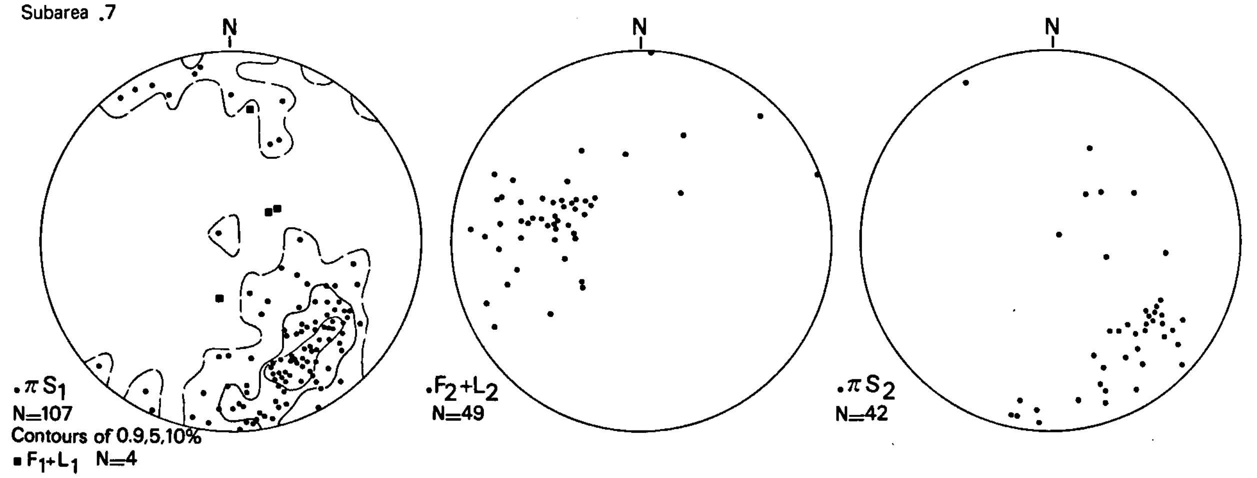


Fig. 5-9. Distribution of planar and linear structural elements in subarea 7 (for its boundaries, see Fig. 5-7). Counting circle 1%.

S_1 in the northwestern slopes of the Monte della Preja. It is not known what caused their different orientation, though it may have been faulting. Observed faults are parallel to S_1 but they could change the orientation of F_2 folds by rotation of the walls of the fault relatively to each other. The existence of faults in this subarea is confirmed by Bearth (1956a) and Wetzel (1972) who call them mylonite zones and suppose that they are related to the Simplon–Centovalli Fault. Wetzel (1972) states that they are common in the Val Loranco but lose their importance towards the west. These faults were not considered in the profiles, for their displacement could not be established.

In a general way, folds 25 and 26 conform to the folds shown on the profiles of Blumenthal (1952).

5.3.3. *Structural zone 3: the eastern part of the Antrona Syncline (subarea 8). The extent of structural zone 3*

This subarea consists of steeply to shallowly dipping (the latter resulting from downslope gravitational movements) inverted and non-inverted S_1, north of the closure of the Antrona Syncline.

Fold 26 – F_2 (Antrona Syncline). – Its closure is found at approx. 1200 m altitude in the small stream between Colorio and Cresta in the northern slopes of the Val Bognanco. Here, refolding of F_2 folds by F'_3 folds occurs in a good section through first order parasitic F_2 folds at 1320 m altitude, west of the stream. Upslope and west of this section the F_2 folds have been refolded by open folds with tens of metres amplitude and with curved axial planes. The direction of tectonic transport of these folds is always downslope, and after Bearth (1956a) they are believed to result from uplifting of the northern part of the Antrona Syncline along late faults in the Val Bognanco, which are associated with the east-west directed portion of the Simplon–Centovalli Fault.

In the section between Pizzanco and Alpe Oriaccia, Bearth (1956a) described a synformal closure above Alpe Pranzano. He considers it as the closure of the Antrona Syncline, and his opinion is adhered to in this paper.

Folds 27 and 28 – F_2. – Present no particulars.

The extent of structural zone 3. – Minor folds in the Verosso gneisses (zone of Mischabel–Siviez, Bernhard Nappe) show that these rocks structurally belong to the northern limb of the Antrona Syncline. This relation is also suggested by asymmetrical folds in the profiles of Blumenthal (1952), situated between the Cima d'Azoglio and the Antrona amphibolites.

Structural zone 3 terminates in the east, where the Antrona Syncline meets the Simplon–Centovalli Fault and its northern limb disappears (Amstutz, 1954). Enclosure 1 and the preceding descriptions have shown that structural zone 3 is characterized by inverted or subvertically dipping non-inverted S_1 and first order parasitic folds whose asymmetry indicates a synform in the south. At the Alpe Vallaro and in the Portjengrat zone south of Alpe Campo the same observations were made.

Therefore it is concluded, that structural zone 3, the northern limb of the Antrona Syncline, extends continuously from the Simplon–Centovalli Fault to the Saaser Vispa.

5.3.4. *Structural zone 4: the eastern part of the Furgg zone (subarea 9)*

Structural zone 4 is the southern, non-inverted limb of the Antrona Syncline, and subarea 9 consists of the portion of the Furgg zone that is enclosed by the axial-plane trace of the Antrona Syncline and the serpentinites in the Vallone Pasquale. It contains isoclinal F_2 folds in calcareous rocks intercalated with gneiss in the eastern slopes of the Monte della Preja. These are too small to be shown on the Enclosures.

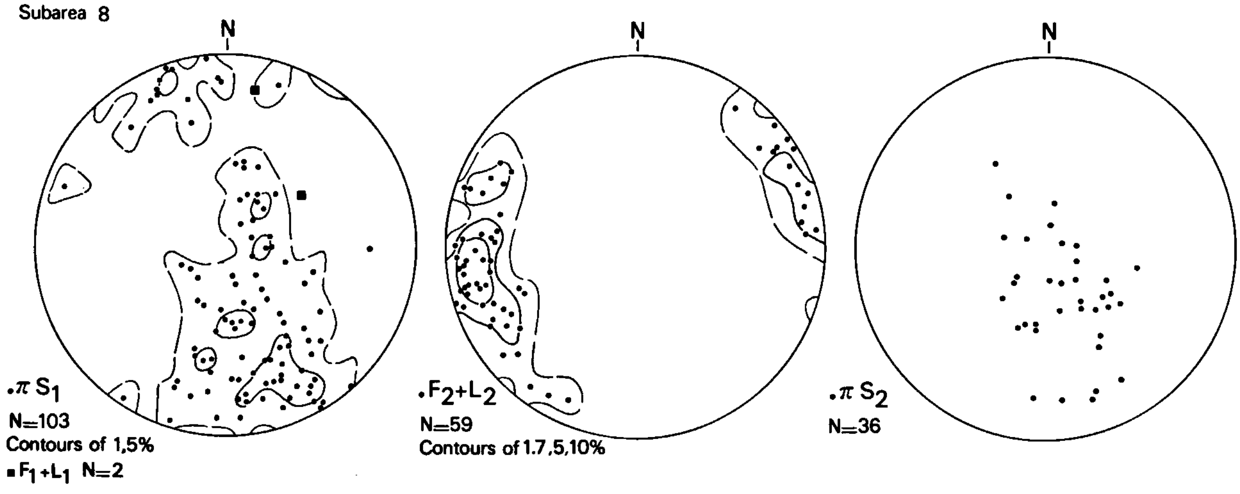


Fig. 5-10. Distribution of planar and linear structural elements in subarea 8 (for its boundaries, see Fig. 5-7). Counting circle 1%.

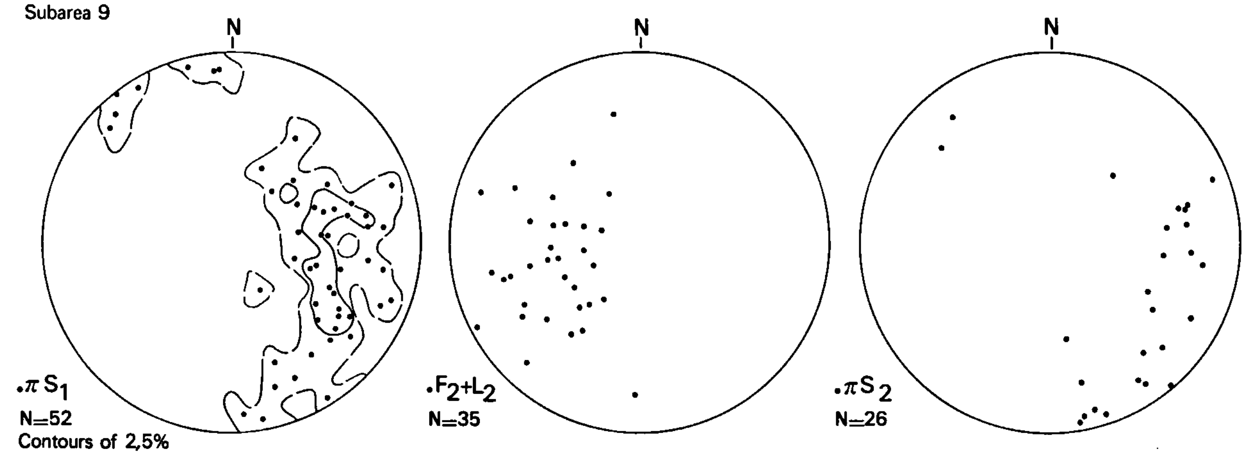


Fig. 5-11. Distribution of planar and linear structural elements in subarea 9 (for its boundaries, see Fig. 5-7). Counting circle 1%.

A comparison of the stereograms (Figs. 5-9, 5-11) of subarea 7 and 9 shows the different orientation of S_1 in the northern and southern limbs of the Antrona Syncline in the eastern part of the Furgg zone. The moderate to steep plunge of the Antrona Syncline in subareas 7 and 9 brings the greater part of the Furgg zone above the topographical surface, and consequently the Furgg zone has almost completely disappeared east of the Monte della Preja.

There is a little F'_3 refolding of the axial-plane trace of the Antrona Syncline near the Alpe del Gabbio.

5.3.5. *Structural zone 4: the Alpe Pasquale area (subarea 10)*

The boundary with subarea 9 is formed by the contact between the Furgg zone gneisses and the serpentinites near Alpe and Vallone Pasquale. The contact has been faulted in the southeastern spur of the Monte della Preja, and above the Bacino Alpe dei Cavalli, but between these locations the fault disappears into the serpentinites. The fault plane above the lake contains lenses of calcareous schists with sizes between some dm and 6 m, and drag along the fault plane has changed the orientation of S_1 in its vicinity. Between 1800 and 2150 m altitude the southern contact of the serpentinites is probably faulted too, but structurally the serpentinites belong to subarea 10, for they contain subvertical F_2 folds.

The dip of S_2 increases from Alpe Cheggio to Alpe Pasquale, and its changing orientation describes an antiformal monocline of F'_3 age. The asymmetry of minor F'_3 folds, occurring exclusively at the Alpe Pasquale, is parasitic on this monocline.

Numerous subvertical F_2 folds occur at the Alpe Pasquale south of the huts, in Furgg zone schists south of the serpentinites, and around the small lake at 2345 m, in the Antrona amphibolites. On the other hand, F_2 folds with slight to moderately plunging axes occur in the eastern part of the Alpe Pasquale, where they have been refolded by F_3 folds. Folds of the two generations have

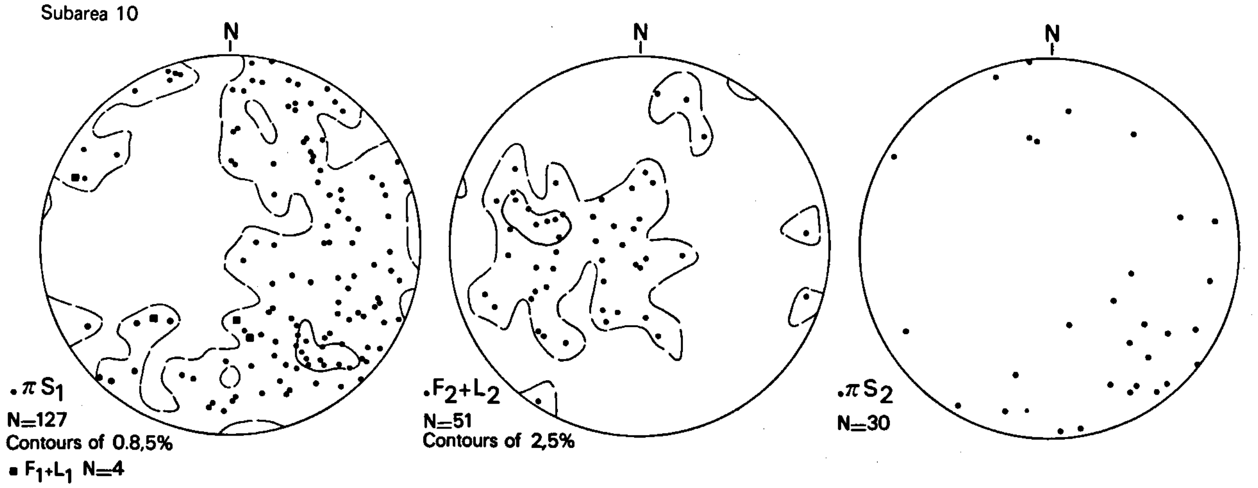
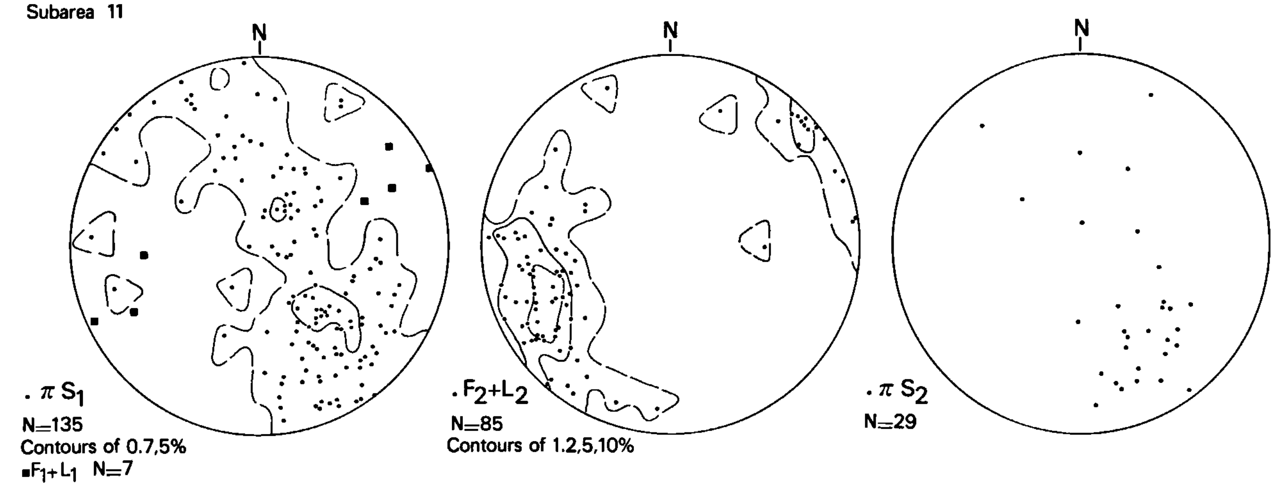


Fig. 5-12. Distribution of planar and linear structural elements in subarea 10 (for its boundaries, see Fig. 5-7). Counting circle 1%.



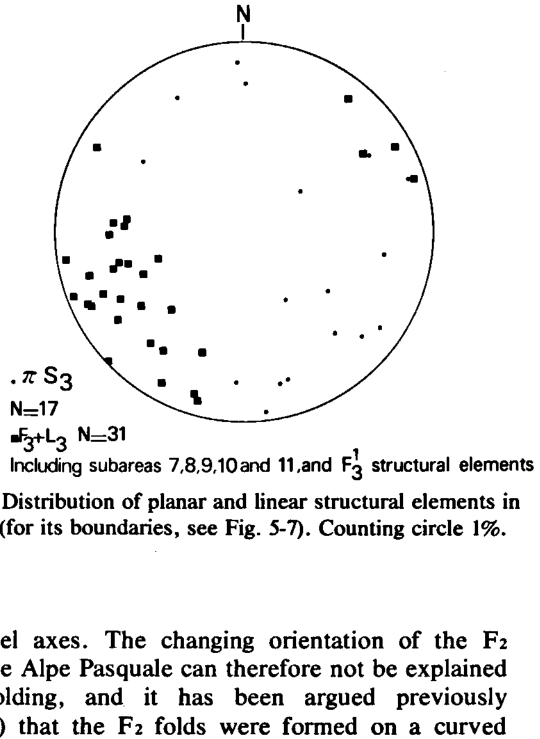


Fig. 5-13. Distribution of planar and linear structural elements in subarea 11 (for its boundaries, see Fig. 5-7). Counting circle 1%.

semiparallel axes. The changing orientation of the F_2 folds at the Alpe Pasquale can therefore not be explained by F_3 folding, and it has been argued previously (see 3.5.4) that the F_2 folds were formed on a curved surface. This deviation from normal conditions, apparent on a regional scale from the relatively steep plunge of many F_2 folds in the eastern part of the Furgg zone and adjoining portions of the Antrona Syncline (see 5.3.7), could have been induced by the nearby presence of the Monte Rosa Nappe. It may have imposed an aberrant orientation of the compositional layering and S_1 by behaving as a more resistant rock unit during F_1 deformation.

5.3.6. *Structural zone 4: the remainder of the southern limb of the Antrona Syncline (subarea 11)*

Minor F_2 folds were less frequently found than in the preceding subareas, but this part of the Antrona Syncline was not investigated in great detail, and there are likely to be more folds than were actually mapped.

Folds 29 and 30 – F_2. – Together they are a first order parasitic F_2 fold. Numerous lower order parasitic folds were found in the closure of fold 29 (Fig. 3-6A).

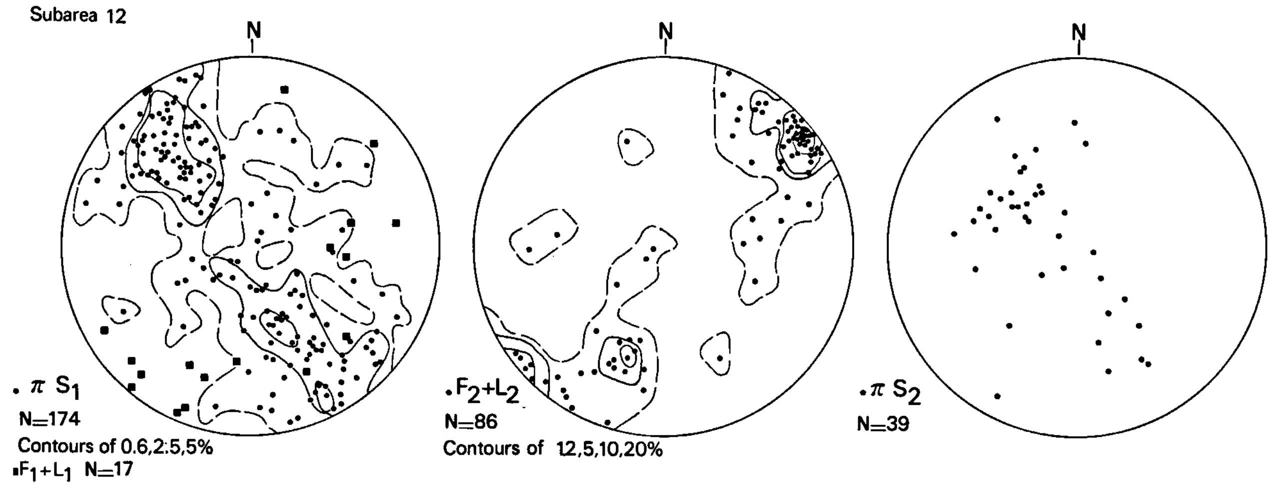
Other parts of subarea 11. – The area around Alpe Gianoli (at the centre of the inset 'Antrona Syncline F_2' in map 2) does not show a consistent asymmetry of the F_2 folds, and it is probably situated near the closure of the Antrona Syncline.

The section along the stream descending from Alpe Curtit shows that between structural zone 5 and the T. Bogna, large F_2 folds are absent. East of that stream and close to Bognanco Terme occurs an area of metasediments of unknown age and uncertain structural position, the 'Bognanco-Keil' (Bearth, 1939). For details, see Bearth (1939), Blumenthal (1952) and Amstutsz (1954).

5.3.7. *Stereograms*

See Figs. 5-8 to 5-13. The steeper plunge of the F_2 fold axes and L_2 in subareas 7, 9 and 10, of the eastern part of the Furgg zone and adjoining portions of the Antrona Syncline, appears clearly when their distributions are compared with those of subareas 8 and 11, covering the remainder of the Antrona Syncline. The distributions of the poles to S_2 have been affected by post-F_2 faulting in subarea 7, and refolding by F_3 in subareas 9, 10 and 11. Subarea 10 moreover contains the subvertical folds at the Alpe Pasquale, and subarea 11 includes the Passo del Fornalino area where F_4 folding has been active.

The distribution of poles to S_1 forms a great-circle girdle in most subareas, and the orientation of the girdle axis changes from the eastern to the western part of the Antrona Syncline, corresponding to the different orientation of the F_2 folds.



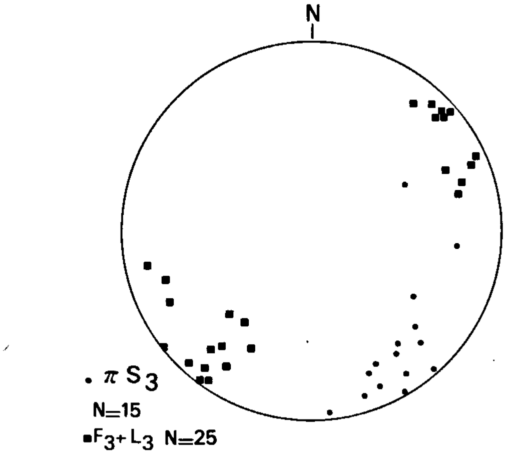


Fig. 5-14. Distribution of planar and linear structural elements in subarea 12 (for its boundaries, see Fig. 5-7). Counting circle 1%.

5.4. THE SECTION THROUGH THE CAMUGHERA-MONCUCCO COMPLEX

5.4.1. *Structural zone 5: the Camughera zone and part of the Moncucco zone (subarea 12)*

This structural zone is defined as the closure of the F_3 Brevettola Antiform (fold 33). It contains an important first order parasitic fold on the Antrona Syncline, the Camughera parasitic fold that is formed by closures 31 and 32.

Folds 31 and 32 – F_2. – Antiform 31 is located in the augengneisses below the summit of the Camughera and at Alpe Saudera, occurring in an area with tight and isoclinal folds of uncertain asymmetry. It is tighter than synform 32, and together they are an example of adjoining fold closures with dissimilar geometry. Synform 32 is very well exposed in the Val Brevettola where Blumenthal (1952) correctly interpreted it as a large synformal fold, with its closure in the north. It might form the continuation of the so-called transversal fold of Antronapiana (Amstutz, 1954), a synformal fold in the Camughera zone near Antrona–Schieranco.

The present thickness of the Camughera orthogneisses (approx. 1200 m in the closure of the Brevettola Antiform) is believed to be more than twice their average thickness after F_1 folding, estimated to be between 400 and 600 m, because of the Camughera parasitic fold contained in the orthogneisses. This explains the apparently rapid disappearance of the orthogneisses in the Val Bognanco, which is schematically indicated on the enclosures since their distribution is not accurately known. The changing thickness of the Camughera zone between the Val Bognanco and the Val Antrona (Blumenthal, 1952; Bearth, 1956a) may also be explained by variations in the apparent thickness, as a result of F_2 folding and the intersection of folds with the topographic surface.

It is believed that thinning of the Camughera orthogneisses in a northerly direction, as a consequence of their disappearance in the Val Bognanco, caused antiform 31 to fold a thinner orthogneiss body than synform 32, and that this produced the different geometry of both closures.

Fold 33 – F_3 (Brevettola Antiform). – The changing orientation of S_2 over a distance of approx. 3 km in the Val Brevettola defines an antiformal F_3 fold that has been called the Brevettola Antiform. First order parasitic F_3 chevron (Turner & Weiss, 1963) folds cause the Camughera augengneisses to disappear south of the intersection between its axial-plane trace and the T. Brevettola. Their age is demonstrated further down the river by F_2 folds with vertical axial planes, contrasting with the recumbent S_2 in the closure of the Brevettola Antiform (see also Laduron & Merlyn, 1974). A second level of orthogneiss encloses layers of amphibolite and biotite schists that may be hinges of F_1 folds. This orthogneiss is overlain by quartz-mica schists belonging to the Salarioli-'Mulde', that have been strongly and irregularly deformed by folds with subhorizontal and subvertical axial planes. The enveloping surfaces, and at times S_1 itself, have a moderate to steep southerly dip. From here until the bridge at Alpi Sogno, no more F_3 folds were found, but the axial planes of F_2 folds and S_1 maintain a southwesterly dip. Southeast of the bridge, southeasterly dipping S_1 and S_2 prevail until close to the ultramafic body of Montescheno.

In the section from the Camughera towards the ultramafic body, a southeasterly dip of S_2 occurs in the spur towards the Passo del Pianino, and the prevailing northwesterly dip of S_1 has been caused by a high angle between S_1 and S_2, resulting from F_2 folding. Between Alpi Sogno and the ultramafic body few exposures occur, and many of these are not in situ. Here, however, refolding of F_2 by F_3 is relatively frequent, and a southeasterly dip prevails until close to the ultramafic body (locally also southeast of it), where S_1 has been finally rotated into the overturned position of structural zone 6, south of the closure of the Brevettola Antiform.

Continuation of the axial plane of the Brevettola Antiform towards the northeast. – Limited exposure on the northeastern side of the Camughera makes it difficult to trace the axial plane of the Brevettola Antiform towards the northeast. In the R. di Molezzano close to the Alpe Carbone, open F_3 folds with wave-lengths of some tens of metres and amplitudes of a few metres are parasitic on an antiform in the northwest. Changing orientations of S_1 higher in the small valley of the R. di Molezzano indicate the existence of other F_3 folds whose closures were not observed; they appear to have the same asymmetry as the folds mentioned before. The axial-plane trace of the Brevettola Antiform is directed towards the curve in the Simplon–Centovalli Fault trace (Amstutz, 1954; Bearth, 1956a, 1956b; Hunziker, 1970) east of Bognanco Terme (Fonti), suggesting that this curve could be a result of F_3 folding. Though some folds with F_3 properties were actually found east of Bognanco Terme, the existence of a fold hinge could not be demonstrated, chiefly because of intensive kink folding associated with the Simplon–Centovalli Fault. No minor F_3 folds were found in the Monte Leone Nappe.

The other possible continuation of the axial-plane trace, in the southern slopes of the Val Bognanco, should pass north of the exposures in the R. di Molezzano, and subsequently curve towards the east. This possibility has been checked in the section between Vagna and Domodossola, where a major F_3 fold is not likely to exist but the structural geology has been complicated by kink folds and faults associated with the Simplon–Centovalli Fault.

When the axial-plane trace of the Brevettola Antiform is tentatively continued as a straight line through the bend in the Simplon–Centovalli Fault, it meets the Val Divedro (Val Diveria) near the locality called Enso, where a large antiformal hinge (400–500 m exposed height) can be seen. This hinge forms the connexion between the Antigorio Nappe and its 'root-zone'. It seems possible that investigations there would reveal a continuity with the Brevettola Antiform, though Milnes (1974) does not report F_3 folding from this location. The present author found, near the closure at Enso, and along the Sempione main road, subvertical minor folds on cm–dm scale, evidently predating the closure and having folded a schistosity plane. On this evidence a post-F_2 age in the local series of events can be attributed to the closure at Enso.

Continuity between the Brevettola Antiform and the Vanzone Antiform in the southwest seems certain; it will be discussed in 7.3.

5.4.2. *Stereograms*

See Fig. 5-14. The plunge of the F_3 fold axes and L_3 varies between shallowly easterly and moderately southwesterly, and the F_3 axial planes dip steeply northwest. The effect of F_3 folding on the distribution of S_2 is shown by the southeasterly dip of most S_2 planes; a northwesterly dip of S_2 occurs only in the northernmost part of structural zone 5. The distribution of F_2 fold axes and L_2 contains two maxima, of which the larger represents their orientation in the Camughera gneisses, and the smaller shows the orientation of F_2 linear elements in the remainder of structural zone 5, southeast of the Camughera gneisses. The angle between the F_2 and F_3 axial directions is small, but the observations in the area southeast of Alpi Sogno where a moderate angle between the F_2 and F_3 axial directions exists, do not figure in the stereogram because many exposures are not in situ.

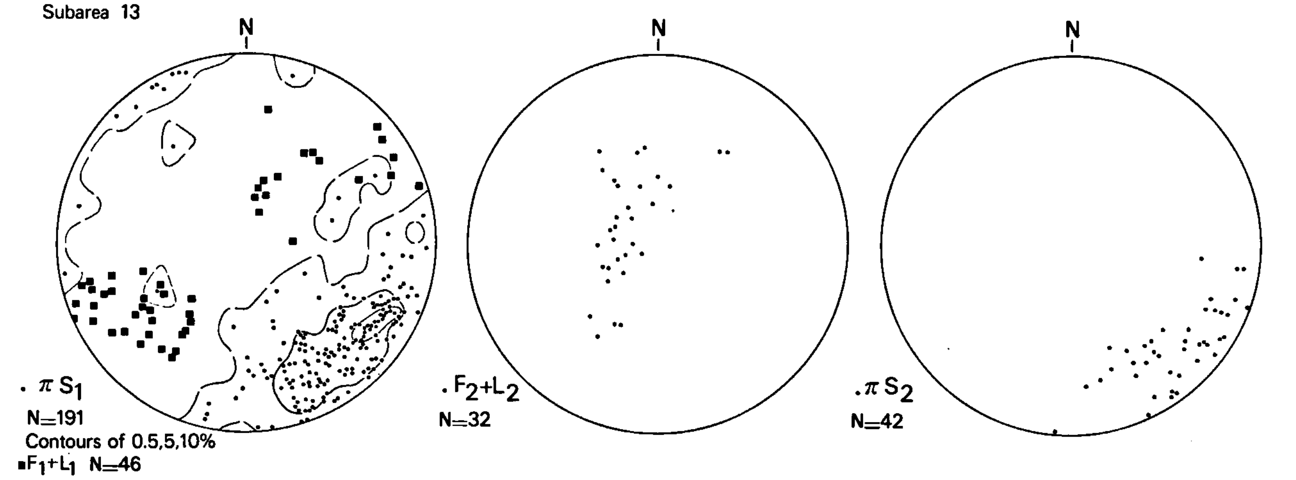
The distribution of poles to S_1 forms a great-circle girdle due to the combined effects of F_2 and F_3. The two maxima are caused by the distribution of the observations over two areas, namely the Val Brevettola with a prevailing southeasterly dip, and the Camughera gneisses where a northwesterly dip prevails.

5.4.3. *Structural zone 6: the remainder of the Moncucco zone (subarea 13)*

This structural zone is the overturned southern limb of the F_3 Brevettola Antiform (see 5.4.1), and comprises the Moncucco zone between structural zone 6 and the Antrona amphibolites.

The axes of F_1 and F_2 folds intersect at high angles in the river section south of the ultramafic body of Montescheno. Their relations are reversed when compared with the Portjengrat zone and the eastern part of the Furgg zone (see 3.1), for the F_1 folds have a shallow to moderate plunge, and the F_2 folds a steep to subvertical plunge. This relationship occurs frequently in structural zone 6. In micaschists adjacent to the ultramafic rocks, F_2 rootless folds occur in lenses of amphibolite of 2–5 cm thickness, and in the micaschists themselves S_2 is a typical transposition foliation (Turner & Weiss, 1963). S_2 is less conspicuous in adjoining amphibolites. In the micaceous gneisses and micaschists between both levels of ultramafic rock, the schistosity is at least locally S_2, since infrequent vertical F_2 folds contain a penetrative axial-plane schistosity.

Folds 34 and 34' – F_1. – F_1 folds and angular relationships between the compositional layering and S_1



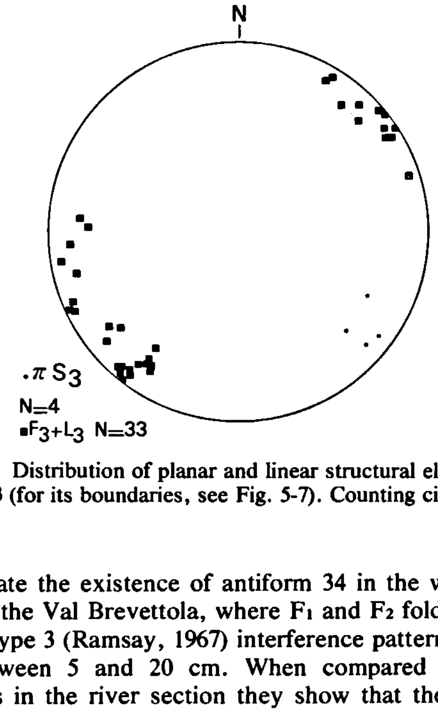


Fig. 5-15. Distribution of planar and linear structural elements in subarea 13 (for its boundaries, see Fig. 5-7). Counting circle 1%.

demonstrate the existence of antiform 34 in the western slopes of the Val Brevettola, where F_1 and F_2 folds have formed Type 3 (Ramsay, 1967) interference patterns with sizes between 5 and 20 cm. When compared to the exposures in the river section they show that the angle between the F_1 and F_2 axial directions is variable. Fold 34, though it has not been found in the river section, is believed to continue in the orthogneisses in the opposite (eastern) slopes of the Val Brevettola. It is likely to be continuous with fold 34′ in the same body of orthogneisses, whose southern limb and the hinge zone were found along the road from Vagna to Alpe Lusentino, and where the characteristic L_1 lineation (see 2.5) is very frequent.

The presumed axial-plane traces of folds 34 and 34′ separate both levels of ultramafic rock, which therefore could have been a single body folded by F_1, and the maps presented by Bearth (1956a) suggest an F_1 fold with a boudinaged southern limb. This could not, however, be satisfactorily demonstrated, for its supposed closure (in the Testa dei Rossi, see Bearth, 1956a) lies west of the area studied. The boudinaged southern limb presents a special problem, since Bearth (1956a) claims that the pegmatite of the Cava di Mica i Mondei has filled the space between two boudins. This pegmatite has a Rb/Sr radiometric age of more than 198 m.y. (determined by Ferrara et al. (1962) on coarse pegmatitic muscovite), and boudinage therefore would have to be of pre-Alpine age.

The sequence of amphibolites, hornblende gneisses and subordinate micaschists containing the southern level of ultramafic rocks, has not been mapped in sufficient detail to completely unravel the structural geology, but some major relations could be established.

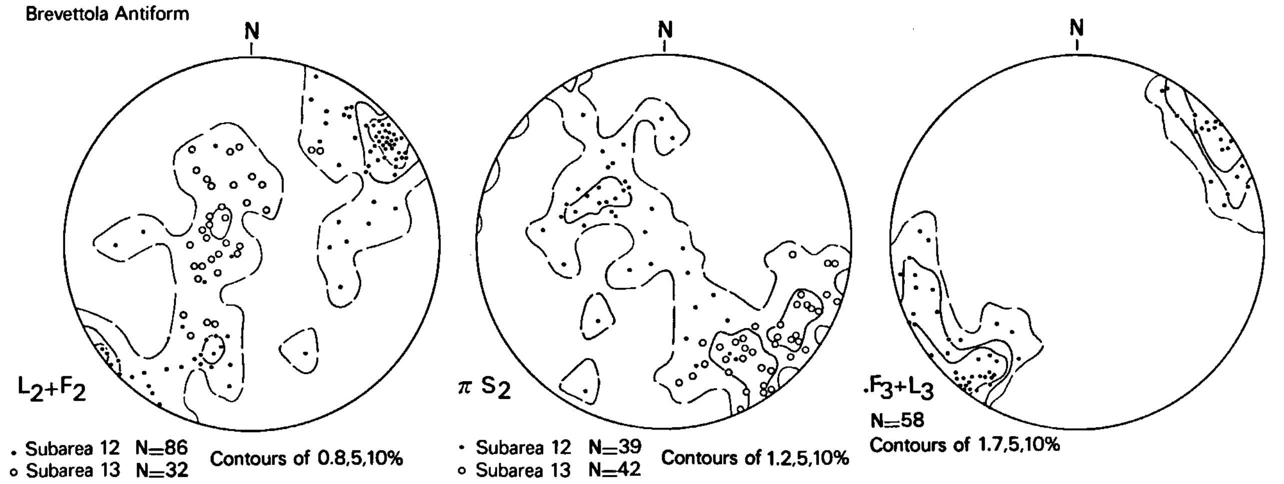
Fold 35 – F_2. – Its northern limb is exposed in the T. Brevettola river section south of the ultramafic body of Montescheno, and its southern limb is visible in the stream descending from the Cava di Mica area and in the Val Brevettola section west of Aulamia. Its plunge steepens in the direction from the stream, descending from the Cava di Mica to the Val Brevettola. This change in orientation has not been caused by F_3, for apart from the L_3 crenulation lineation, F_3 folds are almost absent. It is believed instead that the F_2 fold axes were curvilinear from the outset, either because the F_2 folds deformed non-planar surfaces or because of heterogeneous strain in the axial plane (Hobbs, 1965).

The different direction of plunge of the F_1 fold axes and L_1, north and south of the axial plane of fold 35, is interpreted as the result of the angle between the F_1 and F_2 axial directions. Though this angle is variable, the generalized relationships are as follows (directions of plunge are indicated):

a. Northern limb of fold 35: L_3 shallow plunge (SW); L_1 and F_1 moderate plunge (W–SW); L_2 and F_2 steep and variable plunge.

b. Southern limb of fold 35: L_3 shallow plunge (SW); L_2 and F_2 steep and variable plunge; L_1 and F_1 moderate to steep plunge (NE).

These relationships are consistent with the refolding of F_1 folds and L_1 around the closure of a steeply plunging F_2 fold. When deformation by flattening of the F_2 folds took place after their formation, the angle between the



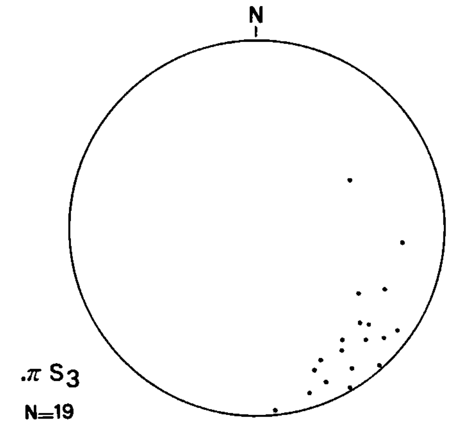


Fig. 5-16. The effect of F_3 refolding on F_2 age structures in the Brevettola Antiform. Subarea 12 covers the northwestern limb and the closure of the Brevettola Antiform, and subarea 13 its southeastern limb. The orientation of F_3 fold axes, L_3 lineations and S_3 axial planes in the two subareas is also shown. Counting circle 1%.

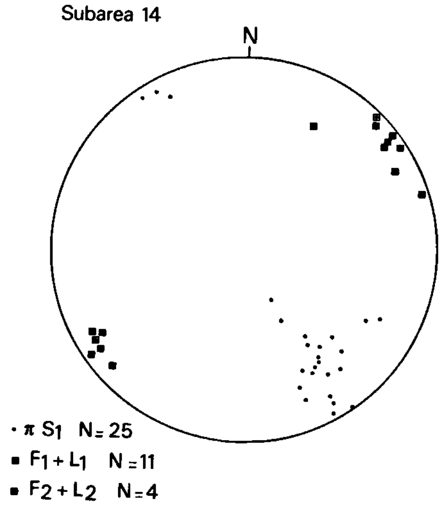


Fig. 5-17. Distribution of planar and linear structural elements in subarea 14 (for its boundaries, see Fig. 5-7). Counting circle 1%.

F_1 and F_2 fold axes will be variable (Ramsay, 1967). This could be a contributing factor in the changing angle between F_1 and F_2 in the Moncucco zone.

5.4.4. *Stereograms*

See Fig. 5-15. The orientation of the F_3 folds and the L_3 lineation in structural zone 6 corresponds to structural zone 5. The distributions of the second generation structures show the steep to vertical plunge of the F_2 fold axes and L_2, which together with the steep dip of S_2 are characteristic for this structural zone.

The poles to S_1 have been spread by a slight regional change in the strike of S_1, not related to F_2 or F_3. The orientation of the F_1 fold axes and L_1 may be close to the orientation of L_3; the moderate to steep northeasterly plunge of L_1 and F_1 are poorly represented in the diagram, since measurements in the northern limb predominate.

5.4.5. *Discussion of F_3 and F_2 folding in the Camughera and Moncucco zones*

F_3 folding. – The distributions of F_2 fold axes, L_2, and poles to S_2 from structural zones 5 and 6 in Fig. 5-16 illustrates the post-F_2 age of the F_3 Brevettola Antiform. There is an aberrant factor in the distribution of the F_2 fold axes and L_2. In the Camughera zone, F_2 and F_3 fold axes are nearly parallel, and F_3 folding therefore could not strongly deviate the orientation of the F_2 fold axes. In the Moncucco zone, where the orientation of the F_3 folds remains the same (compare Fig. 5-14 with Fig. 5-15), the F_2 and F_3 fold axes enclose a large angle, and a moderate angle is present east of Alpi Sogno. This implies that, in the case of F_2 fold axes in the Moncucco zone being initially parallel to those in the Camughera zone, some event before F_3 has changed their orientation. This could have been faulting, and the vicinity of the Simplon–Centovalli Fault, which was active after F_2 (see 7.5 and 6.11), suggests that this is not altogether impossible.

Another explanation for the aberrant F_2 fold axes in the Moncucco zone could be that their orientation at the time of their formation was different from the orientation of the F_2 fold axes in the Camughera zone. Important differences between the F_2 folds in the Moncucco zone, and in the remainder of the present area have been found (see before). Also the asymmetry of the F_2 folds in the Moncucco zone cannot be related to the Antrona Syncline. Moreover, F_2 folds in the Moncucco zone tend to be a better approximation of Class 2 (Ramsay, 1967) folds than elsewhere, and the time–temperature curve of the present area (Fig. 6-8, see also 6.11) suggests that F_2 folding in the Moncucco zone could have occurred a little earlier than in the Saas area. This indicates that the F_2 folding phase in the Moncucco zone had some properties of its own, and that it is not impossible for the initial orientation of the fold axes to have been different from the regionally prevalent orientation.

This explanation does not invalidate the other, and F_2 folding and later faulting in the Camughera–Moncucco Complex would have to be investigated with this particular problem in mind.

5.4.6. *Structural zone 7 (subarea 14)*

The structures in the Monte Rosa zone, most of them presumably F_1 folds, have been described previously (see 2.3 and 2.5). The stereogram (Fig. 5-17) shows the close correspondence in direction of the fold axes and lineations of F_1 and F_2 age.

CHAPTER 6

RELATIONS BETWEEN THE FOLDING PHASES AND THE GROWTH OF METAMORPHIC MINERALS

6.1. SUMMARY OF THE ALPINE METAMORPHIC HISTORY IN THE PENNINE ALPS

Two Alpine metamorphic events have been distinguished in the Pennine Alps. The first, whose duration is unknown (Bearth, 1974), began in the Late Cretaceous approx. 80–100 m.y. before present (Dal Piaz et al., 1972; Bocquet et al., 1972; Bearth, 1974); Dal Piaz et al. (1972) call it the Eoalpine metamorphic event. In the Ophiolite zone of Zermatt–Saas Fee it is characterized by high pressure/low temperature mineral assemblages with omphacite, jadeite, garnet, glaucophane, chloritoid, kyanite, colourless micas (muscovite, phengite and paragonite) and talc (Bearth, 1974). It is believed that this metamorphic event consisted of more than one stage, for most of these minerals exist in several generations, and glaucophanitic rocks are formed later and at the expense of eclogite (Bearth, 1964, 1974).

Intensive metamorphism during the second Alpine metamorphic event is the reason why, in the Antrona amphibolites and other mafic rocks in the present area, the first Alpine metamorphic event is much less evident than in the zone of Zermatt–Saas Fee (Bearth, 1958, 1967; Wetzel, 1972). Eclogite was found in the epidote-amphibolite lens of the Wysstal (Bearth, pers. comm.). At the Alpe Pasquale, transformed eclogites contain a coarse layering formed by different relative proportions of glaucophane, omphacite pyroxene, garnet and colourless mica. Wetzel (1972) reported similar rocks from the Furgg zone.

Little is known about the first Alpine metamorphic event from the Monte Rosa Nappe s.s. and other basement rocks of the present area. In the following, relationships between mineral growth, the F_1, F_2 and F_3 folding phases, and the second Alpine metamorphic event will be considered.

The absolute age of the change in metamorphic conditions is not yet known, but many authors believe that an increase in temperature follows the emplacement of the nappes (Bearth, 1967; Niggli, 1970; Dal Piaz et al., 1972; see also Ayrton & Ramsay, 1974, and Zwart, in prep.) and marks the beginning of the second Alpine metamorphic event. Dal Piaz et al. (1972) believe that the geoisotherms rose as a consequence of the emplacement of the nappes, initiating the second Alpine metamorphic event (they call it the Alpine metamorphic event) that culminated approx. 38 m.y. before present (Hunziker, 1969, 1970). Bearth (1962, 1974) calls it the 'Lepontine phase', for it corresponds in time with the temperature culmination within the Lepontine Nappes.

This metamorphic event produced the albite–oligoclase isograde that Bearth (1958) mapped in and around the Monte Rosa Nappe s.s. (Fig. 1-3). The trajectory of the isograde has been disputed (see 6.2.1).

The time relations between the growth of metamorphic minerals and the folding phases are shown in Fig. 6-1. At the end of this chapter (Fig. 6-8), the folding phases will be correlated with the time–temperature curves of the present area.

6.2. PLAGIOCLASE

6.2.1. *The albite–oligoclase isograde. Composition of the plagioclase*

Bearth (1958) has drawn the isograde along the first appearances of pure oligoclase, and he comments on the frequency of inversely zoned plagioclases (core: albite, rim: oligoclase) near it. Some of these occur on the high-temperature side of the isograde, whereas others are found on its low-temperature side. This isograde is located where higher temperatures persisted, at least for the time during which the oligoclases formed, and therefore it does not accurately mark the change in conditions. The true isograde is more likely to be situated on the low-temperature side of the isograde proposed by Bearth.

Bearth's isograde passes through the Furgg zone in its easternmost part, but Wetzel (1972) claims that it should

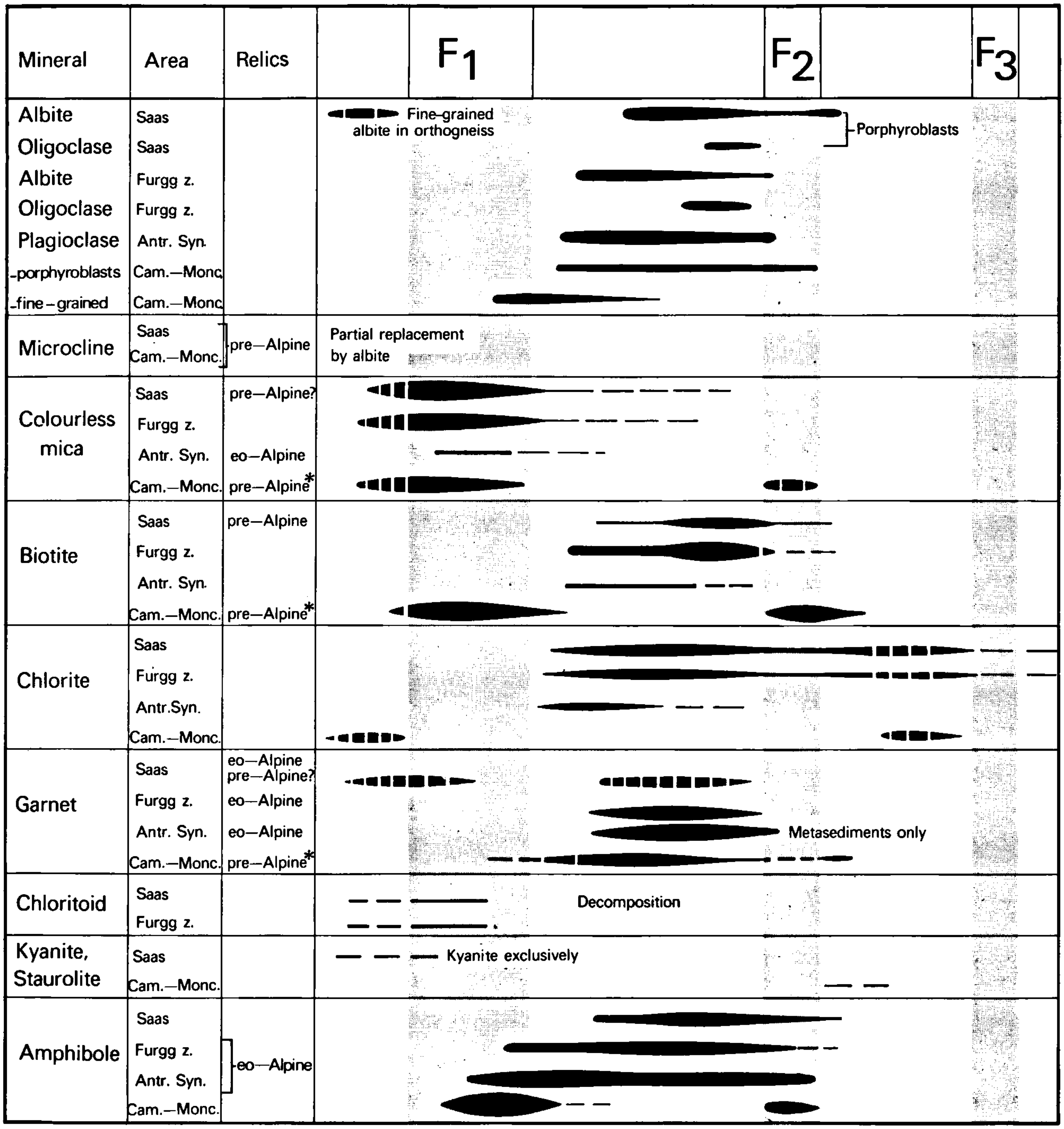


Fig. 6-1. Time relations between the folding phases and the periods of growth of metamorphic minerals. The asterisk refers to minerals from the pegmatite of the Cava di Mica 'I Mondei'.

cross the Furgg zone in the Upper Val Loranco (close to the Portjengrat), where he found plagioclases with albite cores and oligoclase rims in carbonate-bearing rocks. Wetzel locates his isograde at the first regional appearance of oligoclase rims around an albite core, and concludes, while taking the albite–oligoclase isograde as the boundary between the greenschists and amphibolite facies, that the Furgg zone east of the Upper Val Loranco belongs in the lower-temperature part of the amphibolite facies.

Wenk & Keller (1969), using their hornblende–plagioclase An 17 isograde as the boundary between amphibolite and greenschist facies conditions, locate it east of the Furgg zone. They too, consider the maximum

An content of a given plagioclase as the most sensitive indicator for the highest temperature that was reached.

In paragneisses and micaceous intercalations within orthogneisses of the Saas area, and notably in the Almagellertal, inversely zonar plagioclases have a core of albite (An_{0-5}) surrounded by a thin rim of oligoclase (approx. An_{25}). The transition may be gradual but frequently a marked change in refractive index indicates the difference in composition (see also Dal Piaz, 1966). No evidence for interrupted plagioclase growth was found, and it is believed that the plagioclases grew during a period of increasing temperature (Dal Piaz, 1966) that just before the F_2 folding phase (see 6.2.2) superseded the albite-oligoclase isograde. A part of the Saas area has therefore been subjected to amphibolite facies metamorphism, though for a short time only. Unfortunately, with present sampling, the distribution of these porphyroblasts over the Saas area does not indicate where an isograde could be drawn. Albites with An_{0-7} predominate in the Saas area.

The Antrona Syncline lies in amphibolite facies territory, but Bearth (1958) exempted the area northwest of a line connecting the Monte della Preja with the end of the Portjengrat zone. That line conforms approximately to the isograde on the maps of Wenk & Keller (1969). Bearth (1958) found oligoclase and andesine, or more basic plagioclase in the Antrona Syncline; Wenk & Keller (1969) found albite and plagioclase An_{18-20} or $An_{0-5/32}$ in the amphibolites adjoining the Bacino Alpe dei Cavalli. The present author prefers Wetzel's (1972) isograde, according to which the Antrona Syncline would be amphibolite facies territory.

The Camughera–Moncucco Complex lies completely in amphibolite facies territory (Bearth, 1958; Wenk & Keller, 1969). Bearth (1958) found oligoclase, andesine or more basic plagioclase; Wenk & Keller (1969) measured An contents varying between 17–38% and 27–42%.

From the Monte Rosa zone, Reinhardt (1966) described plagioclases with an An content between 0–5% and 15%, with oligoclase occurring more frequently than albite.

6.2.2. *Period of growth*

In the Saas area, the albite porphyroblasts contain frequent inclusions of quartz, colourless mica, chlorite, garnet and opaque matter, and occasional inclusions of clinozoisite, actinolitic amphibole, chlorite, biotite, calcite and titanite. Many of the porphyroblasts have the inclusions concentrated in the core, which is surrounded by a small rim of clear, inclusion-free albite. This could be the effect of a temperature increase during growth of the porphyroblasts, for the oligoclase rims do not contain inclusions either.

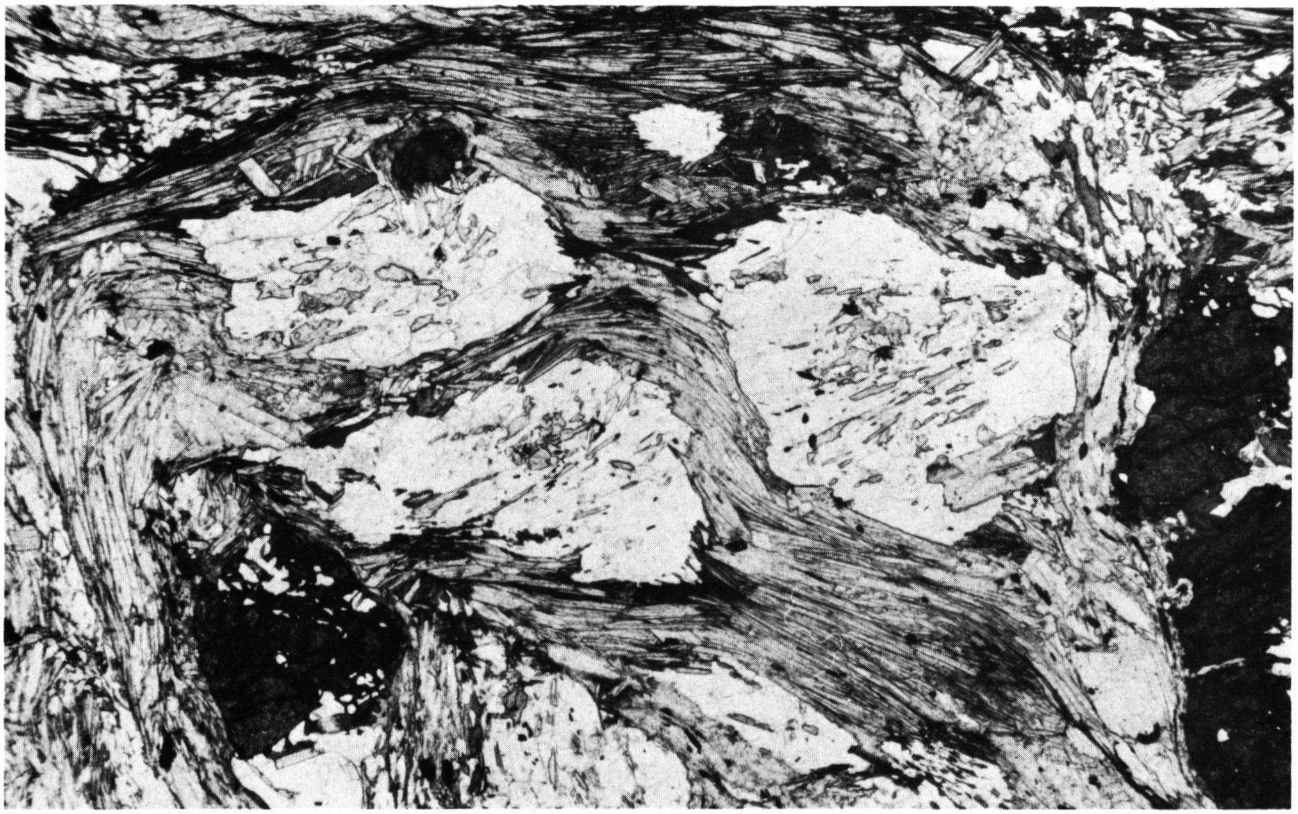


Fig. 6-2. Albite porphyroblast with planar S_i, rotated by F_2 folding. (Paragneiss of the Portjengrat zone, Almagellertal). 10×, parallel polarizers.

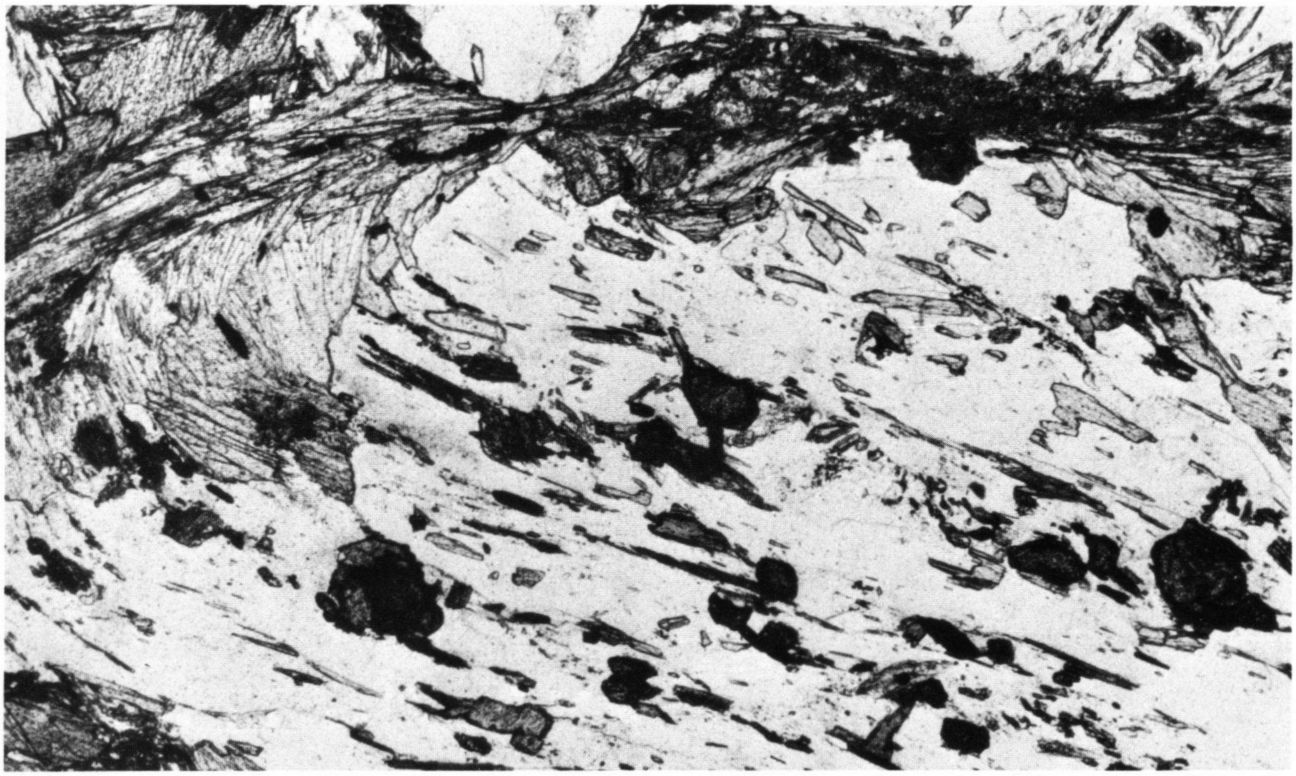


Fig. 6-3. Rotated albite porphyroblast with planar S_i, showing a trace of synrotational growth (F_2 age) at its left-hand boundary. (Paragneiss of the Portjengrat zone, Almagellertal). 30×, parallel polarizers.

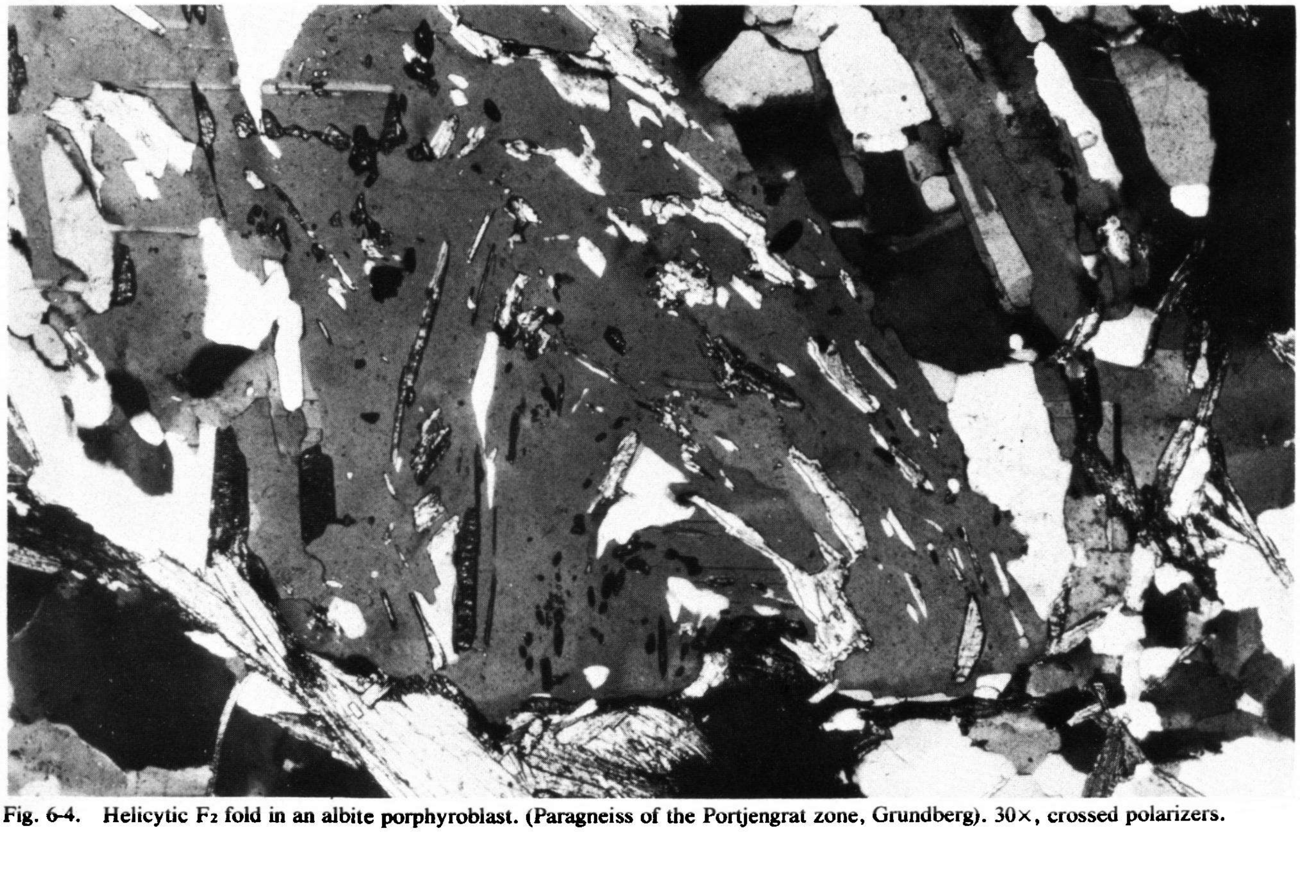


Fig. 6-4. Helicytic F_2 fold in an albite porphyroblast. (Paragneiss of the Portjengrat zone, Grundberg). 30×, crossed polarizers.

The porphyroblasts occur preferentially in mica-rich layers where they replace colourless mica (Bearth, 1952a), and in smaller quantity probably chlorite also. This replacement leaves remnants of colourless mica, elongate parallel to (001) and oriented in S_1, inside the porphyroblasts to form a planar S_i derived from the preferred orientation of the micas in S_1 (Figs. 6-2, 6-3). This internal fabric may be substantiated by thin layers of opaque matter, or one or more of the minerals mentioned previously with an initial preferred orientation parallel to S_1. When no later deformation has taken place, S_i in the porphyroblasts is continuous with external S_1.

The planar S_i is very frequent, most albites having grown between F_1 and F_2. During F_2 folding, many porphyroblasts have been rotated relatively to S_1, but in some F_2 folds the continuity between S_i and S_1 has been preserved. In that case, the porphyrobasts could also have grown after F_2, but the presence of albites with planar S_i in the hinges of F_2 folds demonstrates that this is not so; a non-planar S_i would have been expected there when the porphyroblasts were of post-F_2 age.

There is also evidence for quantitatively less important, later growth of albite. Some albites possess a straight S_i that becomes curved at their margins, without there being evidence for deformation of the porphyroblast; these crystals have, however, been rotated by F_2, and their growth began post-F_1, and lasted until early syn-F_2. Other porphyroblasts with this structure have deformed margins and predate F_2. In a very few cases, albite porphyroblasts contain helicitic F_2 folds defined by inclusions of colourless mica (Fig. 6-4). Other evidence for post-F_2 growth of albite are the porphyroblasts that grew in the hinges of F_2 folds at Hannig. These do not contain mica inclusions, but the fact that they are undeformed is proof for their post-F_2 age. Numerous early porphyroblasts consist of a small number (two to four) of domains with a slightly different extinction position and not related to albite twinning (see also Bearth, 1952a). This structure probably results from F_2 deformation.

In the orthogneisses, fine-grained xenoblastic albites and some recrystallized albites (Wilson, pers. comm.) occur that do not show features from which their age could be derived.

Conclusion. – In the paragneisses and in micaceous orthogneisses of the Saas area, growth of albite porphyroblasts took place between F_1 and some moment after F_2 but before F_3, with major growth between F_1 and F_2. This event may not be limited to the present area; in the eastern part of the Bernhard Nappe, Vallet (1950) describes albites with planar S_i that were subjected to later rotation, and Bearth (1952a) describes them from the Monte Rosa Nappe s.s. The inversely zonar porphyroblasts from the present area have been rotated by F_2, demonstrating that F_2 took place after the temperature culmination. The porphyroblasts that grew post-F_2 are albites.

This period of growth is thought to apply also to the Mesozoic rocks, though albite porphyroblasts are infrequent. Bearth (1967) describes rotated albites with or without an internal fabric from Mesozoic rocks outside the present area. In the amphibolite lens of the Wysstal, albite forms skeletal crystals between hornblende, garnet and pyroxene in transformed eclogitic rocks (see also Wetzel, 1972). According to Bearth (1967) this is the beginning of the crystallization of albite in this rock-type.

In the eastern part of the Furgg zone, comparable structural relationships indicate a similar period of growth of plagioclase. Some of the plagioclases show two stages of growth, when an irregularly shaped albite core filled with inclusions of colourless mica is enclosed by an inclusion-free oligoclase (An_{23}) porphyroblast with roundish outlines. There commonly is a well-defined difference in refractive index between the core and the rim, and the core sometimes forms the smaller part of the total volume. Biotite is a frequent inclusion.

In the metasediments of the Antrona Syncline, plagioclase grew during the same period, but evidence for post-F_2 growth is lacking.

The age of plagioclase growth in the Antrona amphibolites is similar. Plagioclase occurs as porphyroblasts and elongate aggregates of small crystals, and many of these have a preferred dimensional orientation in S_1. Their grain boundaries are imposed by the (110) planes of amphibole, and the preferred orientation was probably imposed upon the plagioclase by a pre-existing microstructure of amphibole oriented in S_1, inhibiting grain growth of the plagioclase in the direction perpendicular to S_1, in much the same way as the boundaries of micas influence the shape of quartz grains (Wilson, pers. comm.). This implies that the growth of amphibole predated or partly predated plagioclase growth. Occasional plagioclase porphyroblasts with a planar S_i formed by inclusions of idiomorphic amphiboles, of smaller size than the surrounding amphiboles, are evidence for this. The porphyroblasts frequently contain pericline twins and are also strongly undulose. Considering that the F_3 folding phase has not been important in the Antrona Syncline, it is believed that the undulatory extinction derives from F_2 folding, and that the plagioclase growth generally ceased before F_2. Evidence for early syn-F_2 growth is, however, supplied by the S_2 in the T. Asinera (see 3.5.4).

In the Camughera–Moncucco Complex, few porphyroblasts with an internal fabric were found, but those that were possess either a planar S_i, indicating growth after F_1 and before F_2, or a curved S_i when the porphyroblasts grew syn-F_2. The majority of the plagioclases are fine-grained and equigranular. In the Moncucco zone a polygonal microstructure may be found, that was formed between F_1 and F_2, for F_2 deformation affected the individual grains. A well-developed polygonal microstructure occurs in the Cava di Mica 'i Mondei' pegmatite (Wilson, pers. comm.).

The fine-grained plagioclases may have grown syn-F_1

and/or post-F_1 but before F_2, because they were subject to deformation during this phase. The grain boundaries of fine-grained plagioclase are frequently formed by the (001) planes of mica, but it is not sure if this can be interpreted in terms of relationship. The dimensional orientation of coarser plagioclase grains parallel to S_1 and/or L_1 in orthogneisses make it likely that plagioclase could grow simultaneously with F_1. In the Camughera–Moncucco Complex, recrystallization of plagioclase after F_2 is frequent, and one case illustrates this very clearly, namely an aggretate of plagioclase with low-angle grain boundaries enclosing a helicitic F_2 fold.

6.3. MICROCLINE

In the Saas area, microcline occurs as porphyroclasts and as small xenomorphic grains in the orthogneisses. The porphyroclasts may be almost idiomorphic but are more frequently elongate augen showing partial replacement by albite. Microcline predates F_1, for the porphyroclasts have been deformed and rotated by F_1, and have a preferred orientation with their longest dimension in or close to the S_1 plane (see also Reinhardt, 1966, fig. 9). No evidence was found for microcline growth syn- or post-F_1.

The microcline porphyroclasts contain replacement perthite at their margins, and it happens that the margins have been almost completely replaced by roundish, newly grown albite grains. These albites occur also in the composition planes of Carlsbad twins and as apparently isolated grains within the porphyroclast. Some augen consist completely of albite with small, irregularly shaped microcline relics between them; the replacing albites have almost the same optical orientation, making it likely that some kind of 'host control' mechanism has been operating. These augen have occasionally been rotated by F_2, and the albites show undulose extinction, making it likely that replacement of microcline took place before F_2. It is actually believed to have occurred before F_1; see Callegari et al. (1969), and 7.5. Wetzel (1972) considers the microcline augen in augengneisses contained by the Furgg zone in the Val Loranco as pre-Alpine and of possible magmatic origin.

In the Camughera–Moncucco Complex, microcline occurs as porphyroclasts consisting of a single deformed crystal or an aggregate of small microcline grains. The latter is perhaps a former single crystal deformed by F_1, because of the dimensional orientation in S_1 and/or L_1 of the aggregate, and sometimes also of the individual grains (see also Bearth, 1958). The microclines are of pre-F_1 age.

A detailed account of the replacement of microcline by plagioclase is given by Reinhardt (1966) on examples from the Monte Rosa zone.

6.4. COLOURLESS MICA

In the Saas area, phengite with 2V close to 0° (a good indicator for phengite, Graeser & Niggli, 1967), and colourless mica with 2V around 30°, that could be phengite or muscovite (Tröger, 1969), occur in the basement rocks and in the Permotriassic quartzites. Bearth (1957, map description), and Hunziker & Bearth (1969), found muscovite and phengite, and Hunziker (1969) states that muscovite, phengite and paragonite coexist in this and the surrounding areas. Laduron & Martin (1969) found coexisting muscovite, phengite and paragonite, on a submicroscopic scale, in strongly Alpine-deformed paragneisses in the Monte Rosa zone where metamorphic conditions were the same as in the Saas area, and a similar coexistence of different colourless micas can therefore be expected in the present area. Harder (1956) found paragonite together with potassium mica in paragneisses of the Bernhard Nappe in the Val de Bagnes.

These observations indicate that the colourless micas in the Saas area could be muscovite, phengite and paragonite in unknown relative amounts and structural positions, and therefore only colourless mica in general is discussed here. An exception is made for the orthogneisses where in the absence of chlorite, the light-green colour of the micas in the field (Exner, 1965; Bryant, 1967; Graeser & Niggli, 1967) demonstrates that phengite is predominant. Moreover, in the majority of the orthogneisses, biotite is absent too, and probably because phengite replaces to a certain degree the biotite–muscovite pair (Graeser & Niggli, 1967).

Colourless mica has a good preferred orientation of (001) in the axial planes of F_1 folds and is therefore of syn-F_1 age; in the orthogneisses, phengite must have grown syn-F_1 for the same reason. It is, however, not impossible that the micas are older and were rotated into parallelism with S_1 (see 2.1). The preferred orientation of colourless micas in the axial plane of F_2 folds has previously been explained by rotation of pre-existing colourless micas (see 3.4). It is possible that growth of colourless micas continued on a small scale after F_1; it may be related to the growth of the albite porphyroblasts (Bearth, 1952a).

The colourless micas were deformed by F_2 folding, but in many cases subsequent recrystallization led to the development of polygonal arcs (see 3.4).

Intercalations of micaschists in orthogneiss, which consist almost exclusively of colourless mica with subordinate quartz and some kyanite (Bearth, 1957, map description), contain corroded remains of microcline. These micaschists are supposed to have been formed out of initially alkali feldspar rich rock (Bearth, 1952a; Reinhardt, 1966, who describes the reaction biotite + plagioclase + alkali feldspar → quartz + kyanite + colourless mica). As all micas are oriented in S_1, this reaction should have taken place syn-F_1, or maybe before F_1 in the case of micas being rotated into their present positions.

In the eastern part of the Furgg zone, the period of growth of colourless mica is believed to be the same. Wetzel (1972, 1973) observed muscovite, phengite, paragonite and colourless phlogopite (see there for details).

Within metasediments of the Antrona Syncline, colourless mica is oriented in S_1 and was probably formed synchronously with F_1.

In the Antrona amphibolites, sporadic colourless micas are oriented in S_1. Also, porphyroblastic colourless micas up to 5 mm across, occur surrounded by epidote, clinozoisite and amphibole (see also Wetzel, 1972, fig. 27) or omphacite. The latter mineral may be oriented with its c-axis at right angles to (001) of the mica (Alpe Pasquale). These porphyroblasts are randomly oriented in a massive, glaucophane and garnet-bearing, eclogitic amphibolite, and are believed to predate F_1.

In the Camughera–Moncucco Complex, colourless mica has grown syn-F_1 or, maybe, before F_1 for the same reasons as in the Saas area.

In the Moncucco zone, colourless micas are well-oriented in S_2, and it could not be demonstrated that these are rotated older micas, so the possibility of syn-F_2 growth of colourless mica exists. Recrystallization of colourless mica into polygonal arcs is ubiquitous, and late biotites have grown between the recrystallized colourless micas. Ferrara et al. (1962) determine the colourless mica from the central orthogneiss in the Moncucco zone as muscovite.

6.5. BIOTITE

In the Saas area, the oldest biotites occur as occasional inclusions in the microcline porphyroclasts, where they have green or brown colours and contain pleochroic halos around zircon. They are of magmatic origin.

Alpine biotite grew relatively late (Bearth, 1967), and it is most frequent in mafic rocks, where it fills the spaces between albite porphyroblasts and has an ill-defined preferred orientation in S_1. When it occurs in the paragneisses together with colourless mica, the latter always has the better preferred orientation. Some biotites are enclosed by albite, biotite growth having started at about the same time as the albite porphyroblastesis, but the major biotite growth postdates major albite growth, and occurred during the temperature culmination in the Saas area, just before F_2, at the time when oligoclase rims were forming around some of the albites.

Occasional biotites are oriented in S_2 without evidence of their having been rotated, and these are believed to be of syn-F_2 age. Others occur between recrystallized micas in polygonal arcs (Fig. 6-5) and these are of post-F_2 age. Both are, however, quantitatively insignificant, and major biotite growth ceased before or simultaneously with the onset of F_2 folding.

Biotite is common in the eastern part of the Furgg zone, for the temperatures reached there were a little higher than in the Saas area. Apparently randomly oriented biotites are frequent and a minority are oriented in S_2; little evidence was found for post-F_2 growth of biotite. Its period of growth is believed to be identical to the Saas area. Wetzel (1972) gives details on the composition of the biotites.

In the Antrona Syncline, biotite occurs sporadically in the metasediments and more frequently in the amphibolites. It grew during the same period as in the areas discussed previously.

In the Camughera–Moncucco Complex, biotite has a perfect preferred orientation in S_1, and it is therefore believed to have grown syn-F_1. It is frequent in S_2 (discussion see 3.4.2 and 3.4.4), and also between colourless micas forming F_2 polygonal arcs, so its growth must have ceased at about the same time as in the aforecited areas.

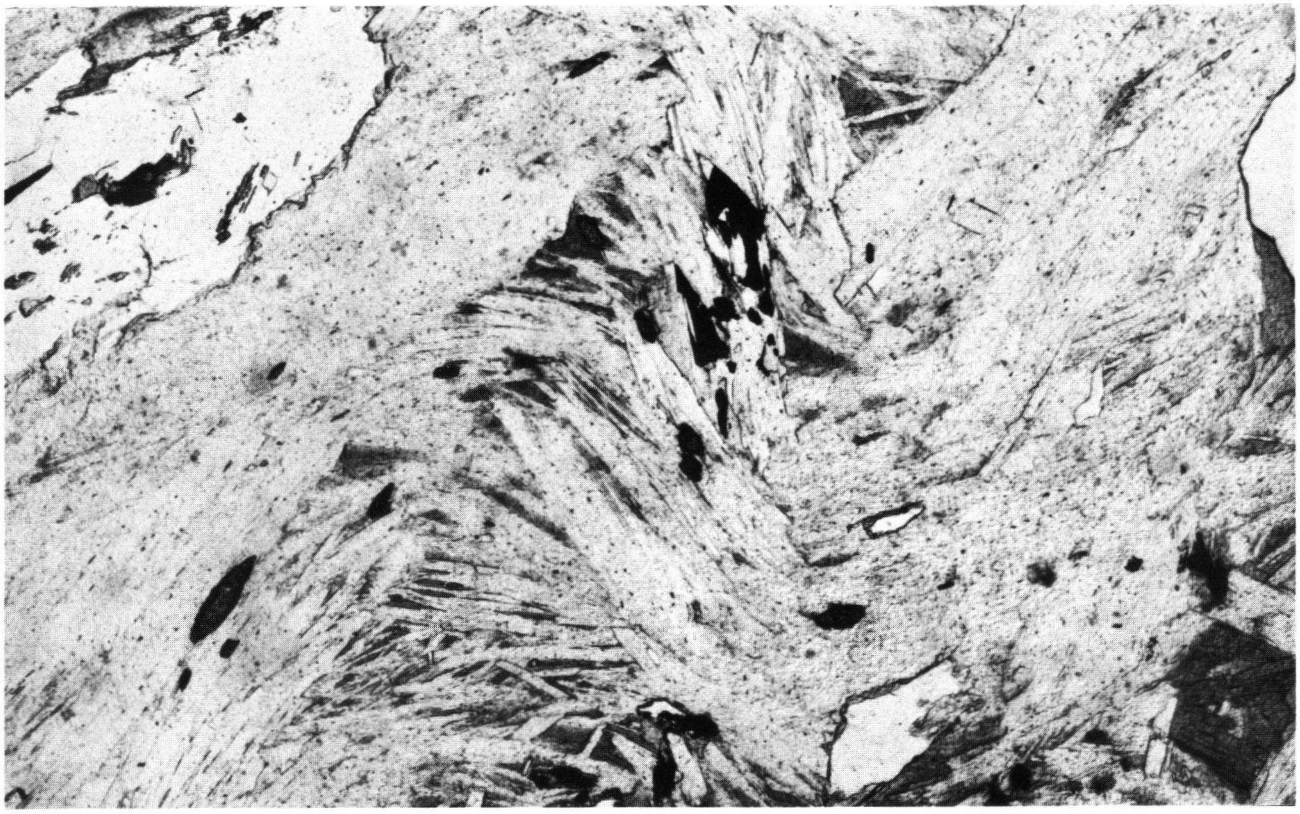
6.6. CHLORITE

In the Saas area, chlorite occurs as isolated grains or aggregates in S_1; as irregularly shaped domains of variable size without a preferred orientation of individual chlorite grains; as a decomposition product of garnet and chloritoid, or in veins and tension fissures. It has a similar period of growth as biotite, extending however for a longer time after F_2. Chlorite is found as inclusions in albite porphyroblasts but is more frequently confined between them, with a weak preferred orientation in S_1. It occasionally occurs in S_2 and as triangular grains between colourless micas in F_2 polygonal arcs (Figs. 3-8, 3-9, 6-6). The irregularly shaped domains of chlorite without a preferred orientation do not have a relationship to structures, and are believed to have formed relatively late, certainly post-F_2. The decomposition of garnet is likely to be of the same age.

In the eastern part of the Furgg zone, the period of growth of chlorite is believed to be the same as in the Saas area. Wetzel (1973) gives a detailed description of chlorites from the Furgg zone.

In the metasediments and amphibolites of the Antrona Syncline, chlorite frequently has a preferred orientation in S_1, and it also occurs together with plagioclase in elongate domains oriented in S_1. The latter occurrence suggests contemporaneous growth of chlorite and plagioclase, post-F_1 to pre-F_2 according to the period of major plagioclase growth established previously.

In the Camughera–Moncucco Complex, chlorite occurs in S_1 of quartz-free gneisses and amphibolites. At Alpe Lusentino it was found in S_1 of the ultramafic body of Montescheno, and chlorite is thought to have grown syn-F_1. At a later stage it was retrogressively formed from biotite, indicated by the frequent chloritization of biotite along the (001) planes.



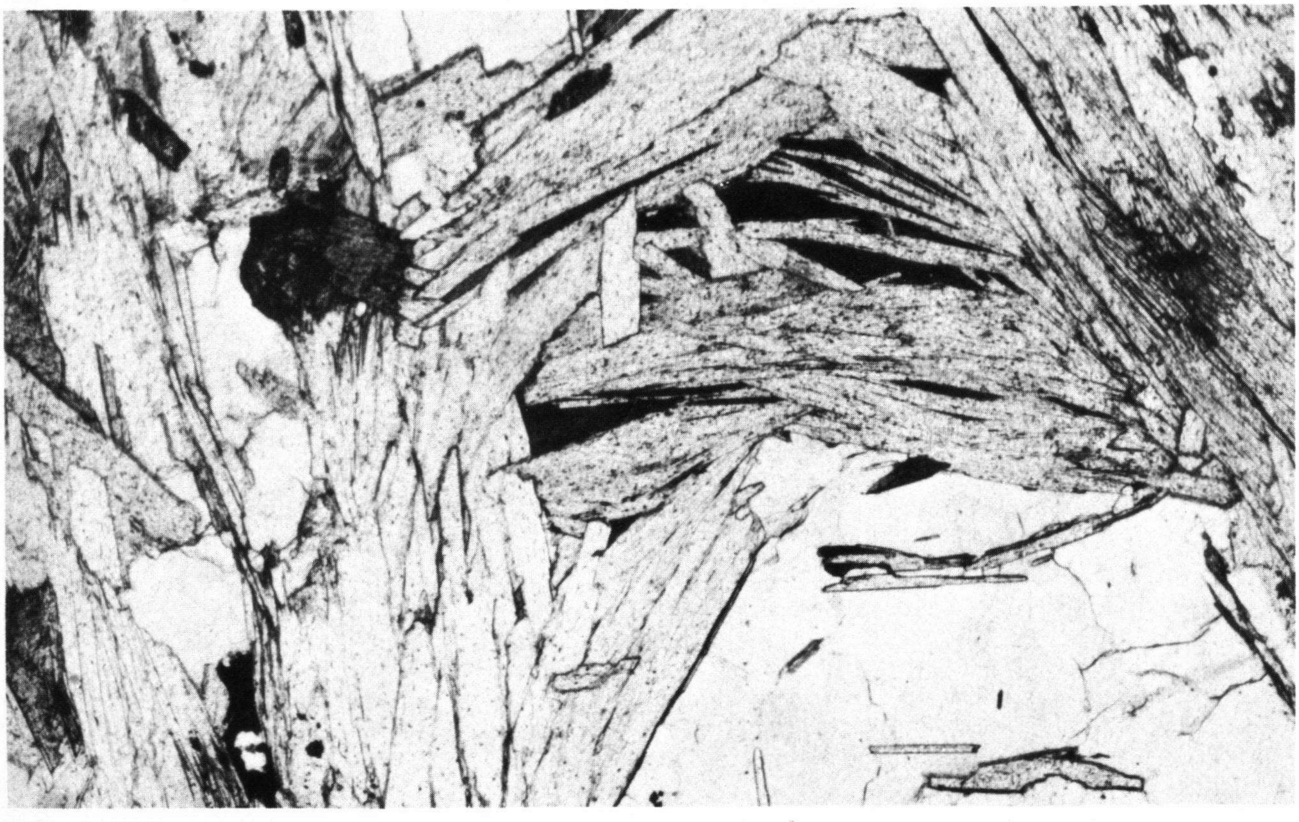


Fig. 6-5. Post-F_2 growth of biotite in the hinges of F_2 folds. The biotites tend towards triangular outlines as they grow between the colourless micas in the polygonal arcs. Paragneiss of the Portjengrat zone, Almagellertal. Parallel polarizers. Upper photomicrograph: 30×, lower: 10×.

6.7. GARNET

Structural evidence for the age of garnet is infrequent in the present area, and its period of growth could not always be accurately determined.

In the paragneisses of the Saas area, many garnets have been rotated by F_2, and they deflect the S_2 planes, demonstrating that they predate F_2. A planar internal fabric of quartz or rutile inclusions occurs sporadically and it shows that some garnets grew after the development of a schistosity. Garnet growth between F_1 and F_2 seems, therefore, not impossible, and it should have begun relatively early since garnet inclusions occur in albite porphyroblasts. It could not, however, be demonstrated that the garnets are of Alpine age, and they might be Hercynian relics as are some of the garnets in the Monte Rosa Nappe s.s. (Bearth, 1952a, 1963). Alpine age can be proved only for those garnets that have inclusions of chloritoid (Bearth, 1963, 1967), but in the Saas area they are small in number and do not have an internal fabric.

The composition of the garnets has not been investigated but they are likely to be almandine-rich; this composition was determined by Wetzel (1972) in garnets of the Furgg zone, and by de Béthune et al. (1968) and Laduron & Martin (1969) in the Monte Rosa zone, in rocks with a composition and metamorphic grade comparable to the paragneisses of the Saas area.

A planar S_i is not infrequent in garnets from paragneisses in the eastern part of the Furgg zone, and it is defined by elongate quartz inclusions of relatively large size, making the S_i planes observable in the field (near the lake-shore at Alpe del Gabbio). The total surface of the quartz inclusions may be more than half the surface of the garnets in thin section. In some cases S_i is continuous with external S_1, and in others the garnets were rotated by F_2, demonstrating that they grew between F_1 and F_2 and are of Alpine age.

Wetzel (1972) reports different phases of crystallization of garnet, depending on rock composition and metamorphic conditions.

In garnetiferous quartz–mica schists from the Antrona Syncline, south of the Pizzo Montalto, the garnets have a planar S_i that becomes curved at their margins (Fig. 6-7), and which is formed by more or less elongate quartz inclusions. S_i is discontinuous with the surrounding S_1 planes because the garnets have been rotated by F_2. These garnets, that can only be of Alpine age because they occur in Mesozoic rocks, have a period

Fig. 6-6. Chlorite displaying a preferred orientation in S_2 (parallel to the short side of the photomicrograph) together with rotated albite porphyroblasts with a planar S_i. Paragneiss, Hannig. 10×, parallel polarizers.

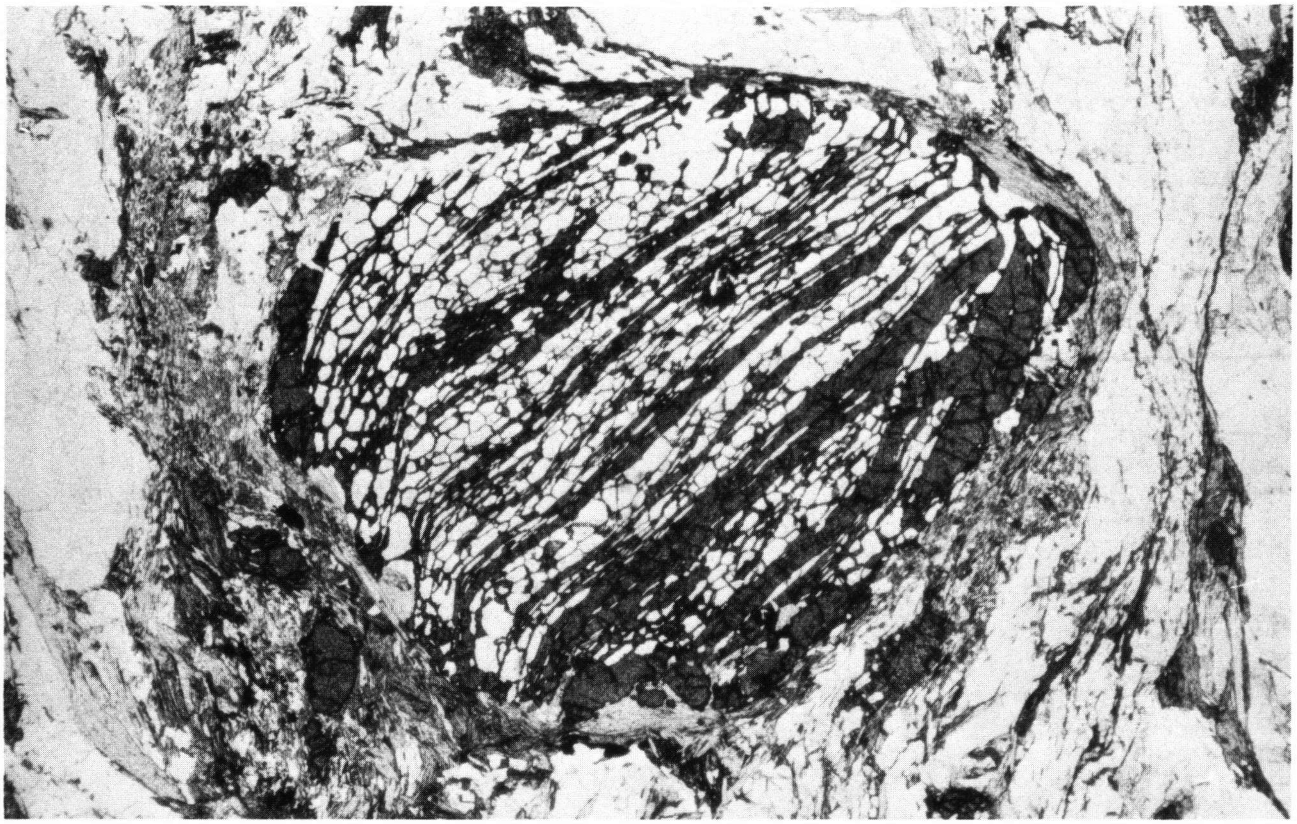


Fig. 6-7. Garnet in quartz-rich Mesozoic metasediments from the Antrona Syncline (Pizzo Montalto). S_i indicates major growth post-F_1 and pre-F_2, and minor growth syn-F_2. 20×, parallel polarizers.

of growth that began post-F_1 and lasted until early syn-F_2. They are strongly reminiscent of the garnets described by Goossens (1970) from Mesozoic rocks associated with the zone of Zermatt-Saas Fee.

The age of the garnets contained by the Antrona amphibolites is uncertain.

In the Camughera–Moncucco Complex, like the paragneisses of the Saas area, few relations between structures and garnet growth were found. Some garnets have an S_i identical to the garnets from the Furgg zone. Helicitic F_2 folds were found only once, in a garnet from staurolite gneiss, and they were surrounded by a rim of massive garnet. These observations indicate that in the Camughera–Moncucco Complex, garnet growth could take place from some moment post-F_1 to post-F_2.

6.8. CHLORITOID

In the micaschists of the Saas area, occasional corroded grains of chloritoid have a preferred orientation in S_1, making it probable that they grew syn-F_1, or pre-F_1 if they were rotated into their present position. They were rotated and deformed by F_2 folding. In a garnet–chloritoid–quartz schist the chloritoid decomposed before F_2, as is shown by the reaction products, colourless mica (paragonite?, Wetzel, 1972) and chlorite, that have been deformed by F_2.

The same relations were found in the eastern part of the Furgg zone. Wetzel (1972) cites examples of replacement of chloritoid by almandine, and it would therefore predate garnet, which complies to a syn-F_1 or pre-F_1 growth of chloritoid.

Chloritoid has not been observed east of the Furgg zone.

6.9. KYANITE AND STAUROLITE

Bearth (1957, map description) reported kyanite from the Saas area, where it occurs not infrequently in micaschist intercalations in the orthogneisses (see 6.4), and sporadically in the paragneisses, where Zwart (pers. comm.) found it at Hannig. Its age is uncertain but it could be pre-syn F_1 (see also 6.4).

It was also found in the Camughera–Moncucco Complex, where it occurs with a random orientation in a staurolite–kyanite gneiss with post-F_2 recrystallized plagioclase, and it seems that kyanite grew simultaneously with the recrystallization of plagioclase. The staurolite contains inclusions of garnet, and it is believed to have grown at the same time as kyanite. Bearth (1956a) reported staurolite from the northern limb of the Antrona Syncline, along the road from Bognanco Terme to San Lorenzo.

6.10. AMPHIBOLE

In the Saas area, amphibole occurs in the zone of Zermatt–Saas Fee and in the mafic intercalations in the basement rocks, and sporadically in the paragneisses. With the exception of the lens in the Wysstal, the mafic intercalations are of very small size and were not distinguished on the map. They are believed to be of Palaeozoic age in the Bernhard Nappe (Bearth, 1962), but they might be Mesozoic in the Portjengrat zone.

The dominant rock-type is prasinite containing an amphibole with the optical properties of actinolite. Transformed eclogite with amphibole of Barroisitic affinity occurs in the lens in the Wysstal (see alsow Wetzel, 1972). According to Bearth (1967) the amphiboles have grown contemporaneously with albite in the mafic rocks, during what is called here the second Alpine metamorphic event. This is confirmed by intergrown skeletal amphibole and albite in the prasinite lens in the Wysstal, and by the relations to be discussed below.

The amphiboles do not possess an internal fabric, and their period of growth is shown only by their mode of occurrence. They never have a preferred orientation parallel to L_1, but a poor preferred orientation in S_1 is frequent. The amphiboles in the prasinites are situated between albite porphyroblasts, and are of larger size than the amphiboles enclosed by the porphyroblasts. Therefore, the period of amphibole growth began at some moment post-F_1, and before or simultaneously with the onset of albite porphyroblastesis. Some amphiboles have been deformed by F_2, whereas others are parallel to the F_2 fold axes, and could have grown in that position, making it possible that amphibole growth terminated after F_2 at about the same time that albite growth ceased. The evidence for post-F_2 growth, however, is limited to one case of undeformed, fine amphibole needles in an intensively F_2-folded rock.

For an extensive description of amphiboles from the Furgg zone and details on their period of growth, see Wetzel (1972, 1974). Post-F_1 growth of actinolitic amphibole took place as in the Saas area, but it is likely that some amphiboles grew syn-F_1 because of their strong preferred orientation in S_1. Good evidence for syn-F_1 growth are grammatites from a dolomite at the Alpe Pasquale with a linear preferred orientation parallel to L_1.

Apart from sporadic glaucophane, the first generation of amphiboles (supposed to be pseudomorphs after pyroxene, see 2.4) in the Antrona Syncline are actinolitic, and have a very light green or almost colourless core. They are dusted by inclusions of titanite, and have been replaced (see 2.4) by blue-green amphiboles that grew contemporaneously with F_1. Most of the amphiboles are of the latter kind, and many have a weak preferred orientation and could have grown between F_1 and F_2. They were deformed during F_2.

A well-defined preferred orientation parallel to the F_2 fold axes is, however, frequent, and it may be so strong that in thin sections perpendicular to an F_2 fold axis, all amphiboles are cut perpendicular to their c-axes. It is probable that these amphiboles have grown syn-F_2, but no evidence for post-F_2 growth was found. Wetzel (1974) gives the succession of amphiboles in the eastern part of the Furgg zone, that is likely to be applicable to the Antrona amphibolites.

In the Camughera–Moncucco Complex, the amphiboles have a strong preferred orientation in S_1 and they are frequently parallel to L_1. They have therefore grown simultaneously with F_1. Evidence for later growth of amphibole is their parallelism to L_2, and it might be that some recrystallization took place after F_2.

6.11. TIME-TEMPERATURE CURVES OF THE SECOND ALPINE METAMORPHIC EVENT FOR THE PRESENT AREA; AND THEIR RELATIONS TO THE THREE FOLDING PHASES

Hunziker (1969, 1970) presented a time–temperature curve for the area south and west of the Simplon–Centovalli Fault, that has been adapted here to suit the Saas area and the Camughera–Moncucco Complex (Fig. 6-8). Hunziker (1970) was consulted for the Rb/Sr critical temperature of younging of Hercynian muscovite (500°), and the closure of the Rb/Sr system in biotite (300°). He assigns a minimal temperature of 450° to growth of phengite in coexistence with quartz and biotite after Velde (1965, 1966) and Velde & Kornprobst (1970), and concludes that phengite growth should have taken place at the time of the temperature culmination. The 450° lower limit has not been retained here, for Hunziker (1972, in Dal Piaz et al.) states that phengite

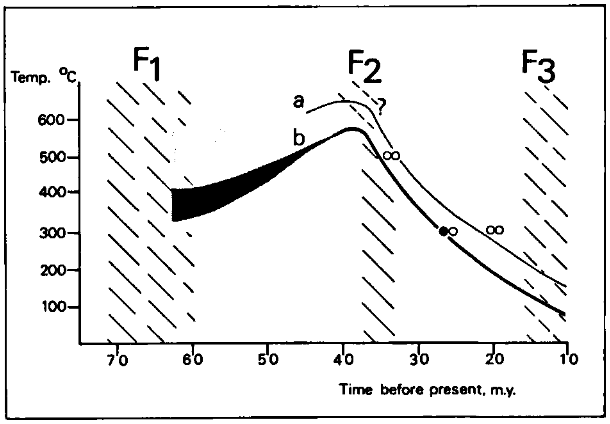


Fig. 6-8. Time–temperature curves for the present area. Dot: biotite cooling age for the Saas area, circles: biotite and muscovite cooling ages for the Camughera–Moncucco Complex (see text for references). Curve a applies to the Camughera–Moncucco Complex, curve b to the Saas area. The ages of F_1 and F_3 folding phases are not accurately known. It is possible that in the Camughera–Moncucco Complex the F_2 folding phase is slightly older than in the Saas area (see 6.11).

growth could have taken place earlier (and consequently at a lower temperature), and in the present area it could be shown that biotite growth postdates phengite growth (see 6.4–5) except perhaps in the Camughera–Moncucco Complex.

The heating portions of the curves, and dating of the F_1 folding phase. – Time–temperature curves for the Saas area and for the Camughera–Moncucco Complex, particularly the Moncucco zone, are illustrated in Fig. 6-8. The existence of a temperature difference between them is concluded from the dissimilar periods of growth of biotite, amphibole and plagioclase in the eastern and western parts of the area, and the different compositions of amphibole and plagioclase. The relationship of these minerals to S_1 in the Saas area and in the Camughera–Moncucco Complex shows that this temperature difference already existed at the time F_1.

Its interpretation is possible in terms of (1) an increase in temperature from west to east, (2) an increase from a higher to a lower structural level (Monte Rosa Nappe s.l. to Camughera–Moncucco Complex) or (3) as a combination of both. The time necessary for temperature equilibration between quickly superimposed, relatively cool thrust plates makes it unlikely that a temperature gradient from a higher to a lower structural level existed (Oxburgh & Turcotte, 1974), and the interpretation preferred here is a temperature increase from west to east, which is characteristic for the second Alpine metamorphic event in this area (see 6.1 and 6.2.1–2; Niggli, 1970).

In the following, an attempt will be made to attribute an approximate absolute age to the F_1 folding phase. From the different deformations of the frontal and the central part of the Monte Rosa Nappe (see 7.5 for the complete argument) it has been inferred that the F_1 folding phase took place during the final stage of emplacement of the nappes. This event is therefore approximately dated by the growth of colourless micas in S_1. It has been previously argued that the temperature gradient of the second Alpine metamorphic event already existed at the time of F_1, and it is believed (see 7.5) that the penetrative F_1 deformation was assisted by a slight temperature increase in the frontal part of the Monte Rosa Nappe. Therefore, no cooling is thought to have taken place between F_1 and the temperature culmination (Fig. 6-8). In the present area it is not possible to date S_1 by mica cooling ages.

In the Sesia–Lanzo zone, however, where the second Alpine metamorphic event was less intensive than in the present area, Dal Piaz et al. (1972) found colourless micas with radiometric ages of approx. 57 m.y. in greenschist facies and high-pressure facies mineral associations that were both formed during the first Alpine metamorphic event. They write (translation by this author): "One has in general the impression that the Eoalpine metamorphism began with a static or pseudostatic phase and afterwards proceeded under prevailing synkinematic conditions". The latter have been associated here with the nappe movements, and because in the Sesia–Lanzo zone the first Alpine (Eoalpine) and second Alpine metamorphic events are separated by a period of relatively low temperature (Dal Piaz et al., 1972; for comparison with other areas see also Ayrton & Ramsay, 1974), it may well be that the age of 57 m.y. marks a cooling event between the cessation of the nappe movements, and the onset of the second Alpine metamorphic event in that area. If this is true, S_1 may have been formed at about 65–60 m.y. before present, and this would also mark the time when the second Alpine metamorphic event began in the present area.

In this interpretation, S_1 has been formed during high-pressure conditions that were beginning to change to the somewhat lower pressure, higher temperature conditions of the second Alpine metamorphic event.
Jäger (1973) states that the highest pressure in the area west of the Lepontine Nappes occurred 65 m.y. before present, which seems to coincide with the formation of S_1 in the nappes.

The time of the temperature culmination. Dating of the F_2 folding phase. – Hunziker (1969, 1970) states that the temperature culmination takes place at about 38 m.y. before present in the western part of the area. There, the F_2 folds postdate the temperature culmination since they deformed and rotated the oligoclase-rimmed albite porphyroblasts that date the culmination (see 6.2.1–2). Moreover, all plagioclases that postdate F_2 in the western part of the area are albites. The F_2 folding phase is therefore believed to have begun a little after the culmination, at about 36 m.y. before present. The albite–oligoclase isograde should have been folded by F_2, but this was only realized after the field work had been completed, and it could not be demonstrated with present sampling.

F_2, the main folding phase, took place during the Early–Middle (?) Oligocene, and it is not believed to have lasted longer than a few million years. This can be argued by the occasional presence of undeformed albites and biotites in F_2 fold hinges, which indicates that after folding ceased, the rocks had not cooled further down than to a temperature corresponding to the quartz–albite–epidote–biotite subfacies of the Barrovian (growth of kyanite after F_2 in the Moncucco zone) greenschist facies. The reactions leading to the formation of biotite and albite are not known, and accurate temperatures cannot, therefore, be given, but a glance at Fig. 6-8 shows that even when the rocks had cooled down to a temperature close to 400 °C the folding phase cannot have lasted more than 4 m.y.

It is likely that F_2 folding in the Moncucco zone took place a little earlier (see below).

The cooling portion of the curves. – The curve for the Saas area is after Hunziker & Bearth (1969); their samples KAW 374–377 give Rb/Sr radiometric biotite ages of 27.6 ± 1.1, 27.6 ± 4.7, 26.1 ± 1.5 and 26.1 ± 2.7 m.y. respectively. The cooling part of the curve for the Camughera–Moncucco Complex has been adapted after the Rb/Sr biotite and muscovite ages by Ferrara et al. (1962). The muscovite ages are from orthogneiss north of the Cava di Mica 'i Mondei' (33.9 ± 8.2 m.y.), and from the Monte Rosa zone where in muscovites of different sizes ages of 34.5 ± 3.5 and 33.6 ± 3.3 m.y. were found. One biotite age from the Monte Rosa zone was used (19.7 ± 1.7. m.y.), and the biotite ages are from the pegmatite of the Cava di Mica 'i Mondei' (25.9 ± 5.4 m.y.) and from orthogneisses east of Cresta (19.7 ± 3.9 m.y.). Data from the Camughera zone in the present area are not available.

The very young Rb/Sr muscovite ages (16–21 m.y.) and biotite ages (12–15 m.y.) northeast of the Simplon–Centovalli Fault (Jäger et al., 1967; Hunziker & Bearth, 1969; Hunziker, 1969, 1970) do not appear in the present area, and the cooling part of the curve for the Camughera–Moncucco Complex therefore has more affinity to the Saas area than to the Lepontine Nappes.

The shape of this curve, however, leads to discrepancy with the occasional growth of kyanite and staurolite in the Moncucco zone (see 6.9) for which the time is not available in Fig. 6-8. A first explanation could be that the curve does not apply to the Moncucco zone, but although the age determinations in the Moncucco zone are few, they do not substantially differ from those in the Monte Rosa zone, and it is felt that over the relatively short distance Saas area – Camughera–Moncucco Complex the shape of the cooling curves should not be too different. A second explanation, and the one preferred here, is that the F_2 folding phase in the Moncucco zone is a little older than in the remainder of the area. This possibility is substantiated by the different properties of F_2 folding in the Moncucco zone (see 5.4.3 and 5.4.5).

The F_3 folding phase. – This folding phase (including both F_3 and F'_3 folds) took place during the cooling stage of metamorphism, and it cannot be dated by its relationship to the growth of metamorphic minerals. It is tempting to postulate some connexion between the Brevettola Antiform and the known differential movements along the Simplon–Centovalli Fault (Hunziker, 1970), in analogy to the Vanzone Antiform (Laduron, 1974), but apart from the fact that the fault has possibly been folded by the Brevettola Antiform (see 5.4.1), no field relations were established between the two structures. The presence of F_3 folds in the Saas area and in the Mischabelrückfalte (Zwart, in prep.) indicates that F_3 folding was a more regional feature, caused by some late event or sequence of events in which movements along the Simplon–Centovalli Fault might also find a place, but which is not restricted to such movements; in fact, the postulated deformation of the fault by the Brevettola Antiform would allow it to play only a passive role after F_3 folding began.

The presence of F_3 folds from the Mischabelrückfalte to the Brevettola Antiform shows that during F_3 folding, stress conditions existed over the whole area from the Mischabelrückfalte southwards. Apart from the Brevettola–Vanzone Antiform (for the present purpose regarded as continuous, see 7.3) they do not significantly affect the regional structural geology. This suggests that F_3 deformation was concentrated in the Brevettola–Vanzone area. All this might be explained by regional tilting of the Pennine Nappes, as a result of uplift in the north, so that stresses built up between the Mischabelrückfalte and the 'root-zone' led to the formation of the F_3 fold generation. The deformation was concentrated in the 'root-zone' (Brevettola–Vanzone) antiform while the rocks north of it transmitted the stress almost without internal deformation. This model implies, however, significant southerly to southeasterly relative movement of the Monte Rosa Nappe, the Bernhard Nappe and the Antrona Syncline. Such a movement could have been accommodated along the Simplon–Centovalli Fault, acting as a passive structure.

Following the suggestion of Laduron (1974) that the location of the Vanzone Antiform was predetermined by the structure of the Monte Rosa Nappe s.l., it may be that this fold propagated itself into successively deeper rocks until it finally folded the Simplon–Centovalli Fault, after which movements along that fault which were related to the folding must have come to a standstill. It does not seem impossible that these movements were registered by Hunziker (1970) as uplifting of the block NE of the fault, between 16 and 18 m.y. before present; these ages would then indicate approximately the period of F_3 folding.

CHAPTER 7

CONCLUSIONS AND DISCUSSION

7.1. SUMMARY OF THE POST-NAPPE STRUCTURAL GEOLOGY

The essence of the post-nappe structural geology has been formulated previously, when the limbs and one closure of the post-nappe major folds with F_2 and F_3 relative age were used to define structural zones of inverted and steeply dipping S_1 or non-inverted and shallowly dipping S_1 (see 5.1 and Fig. 5-1). These major folds are the following:

1. The Mittaghorn Synform – F_2. It is the synform that brings the southern inverted limb of the Mischabelrückfalte back into non-inverted position. Fig. 3-3 shows a small-scale coupled fold of the same geometry from the Rotblattgletscher area.
2. The Trifthorn Antiform – F_2. This antiform,

south of the Mittaghorn Synform, is the cause of the antiformal structure of the Portjengrat zone. It is the continuation (Nieuwland, 1975) into the Saas area of the antiformal Balmahornrückfalte (Te Kan Huang, 1935). The latter is exposed in the Zwischbergental northeast of the present area. It is proposed (see 5.2.6) to replace the name Trifthorn Antiform by the name Balmahornrückfalte in subsequent papers.

3. The Antrona Syncline – F_2. Its northern limb corresponds to the overturned southern limb of the Trifthorn Antiform. The axial plane of the Antrona Syncline continues from the Antrona Syncline itself through the easternmost part of the Furgg zone into the frontal part of the Monte Rosa Nappe s.s., through which it is believed to continue until it meets the axial plane of the Trifthorn Antiform in the Mattmark area. Most of the Portjengrat zone lies in the inverted northern limb of the Antrona Syncline, and this area of inverted S_1 terminates where the folds die out against each other at Mattmark.

4. The Brevettola Antiform – F_3. This fold forms the transition to the steeply dipping 'root-zone' in the Camughera–Moncucco Complex, and it is believed to be continuous with the Vanzone Antiform (see 7.3). For priority reasons, the Brevettola Antiform ought to be renamed the Vanzone Antiform when continuity between the folds is demonstrated in the field.

All major folds are featured on insets 1 and 2 in Enclosure 1, where they are presented in schematic profiles, and a schematic map is to be found in Enclosure 2.

7.2. PROPERTIES AND AGE OF THE THREE FOLDING PHASES

F_1 folds occur everywhere but they are not as abundant as the F_2 folds, and compared with them of relatively small size. They do not show a preference for basement or cover rocks.

The F_1 folds are similar-style, isoclinal or rootless folds, frequently intrafolial, and the concomitant lineation is the intersection of S_1 and the compositional layering. S_1 is the predominant schistosity plane, with the exception of some marginal parts of the Portjengrat zone and the Moncucco zone. The F_1 fold axes have variable orientations as a result of refolding by F_2 and F_3, but it is probable that the initial directions of the F_1 folds were not everywhere the same (see 7.5). The folds are correlated in time with the last stage of nappe emplacement, when differential movements along isolated thrust planes or movement zones were replaced by penetrative deformation (see 7.5). It is suggested that the F_1 folding phase might have terminated at about 65–60 m.y. before present (see 6.11).

The F_2 and F_3 folds postdate the emplacement of the nappes, for they have folded the nappes together with the intercalated Mesozoic rocks. The similar-style F_2 folds may possess an axial-plane schistosity S_2 that locally forms the principal S-plane. Their geometry is strongly dependent on the lithology in which they were formed, and they are nearly always disharmonic and sometimes extremely so. They have pervasively deformed both nappe and cover rocks. Unless the F_2 folds were refolded by F_3, their axial planes dip shallowly to moderately W to N, and their fold axes plunge shallowly to moderately W to NW. The F_2 folds are associated with a crenulation lineation and mineral lineations of several types.

F_2 folding in the Moncucco zone, where the F_2 folds have different properties, may have taken place a little earlier than elsewhere in the present area where it is antecedent to the culmination of metamorphism at 38 m.y. before present. Folding is thought to have begun at about 36 m.y. before present (Early (Middle?) Oligocene), and to have had a maximum duration of 3–4 m.y.

The fold axes of the parallel-style F_3 folds are nearly always subparallel to the F_2 fold axes, and the axial planes dip steeply to vertically, in a N to NW direction. They have intensely deformed the nappe rocks in the Camughera zone and in a portion of the Moncucco zone; elsewehere they occur only occasionally. They are associated with the simultaneously formed F'_3 folds, that have the opposite asymmetry and an opposed direction of axial plane dip. F_3 and F'_3 are accompanied by a crenulation lineation.

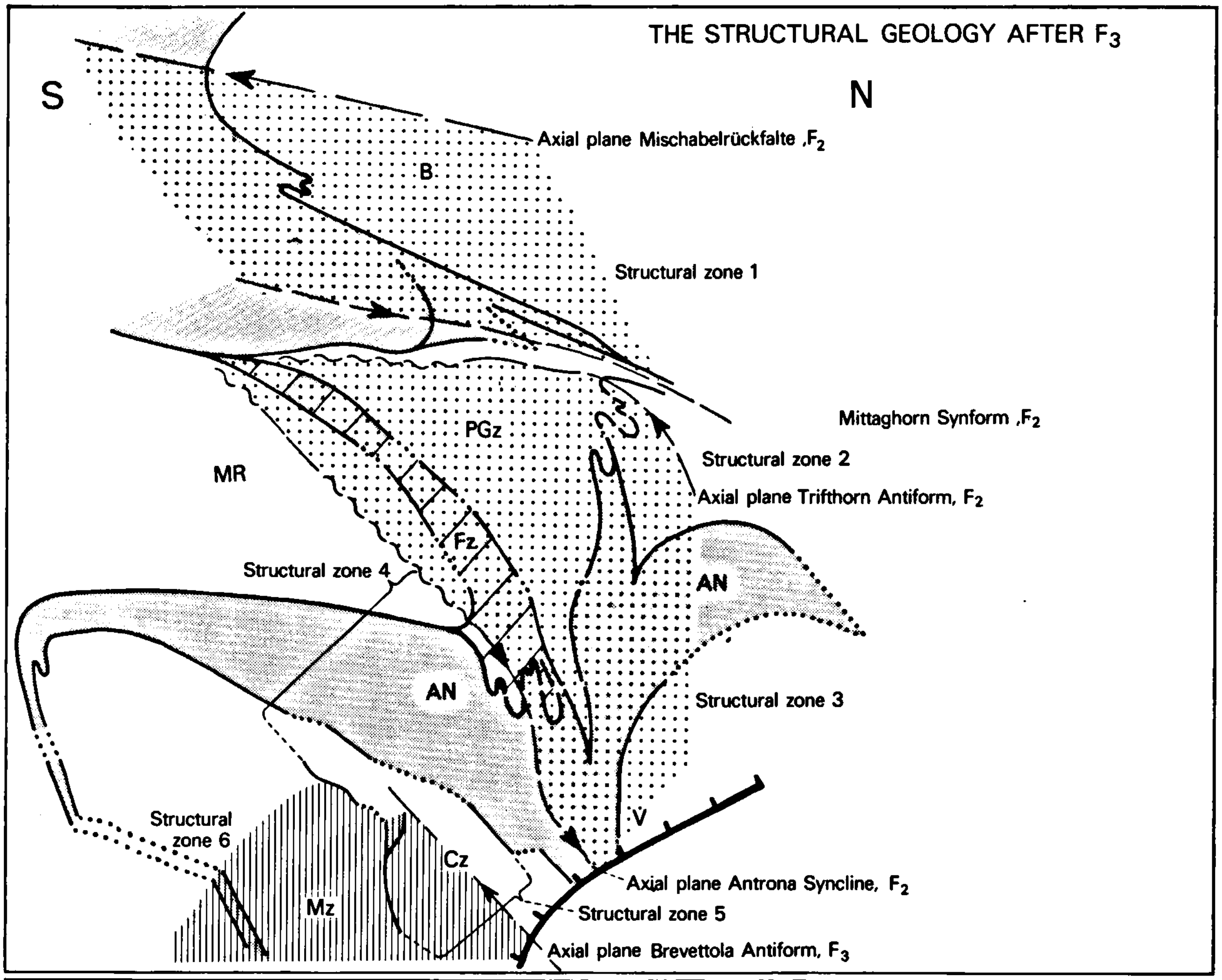
It has not been possible to attribute a definite absolute age to the F_3 folding phase.

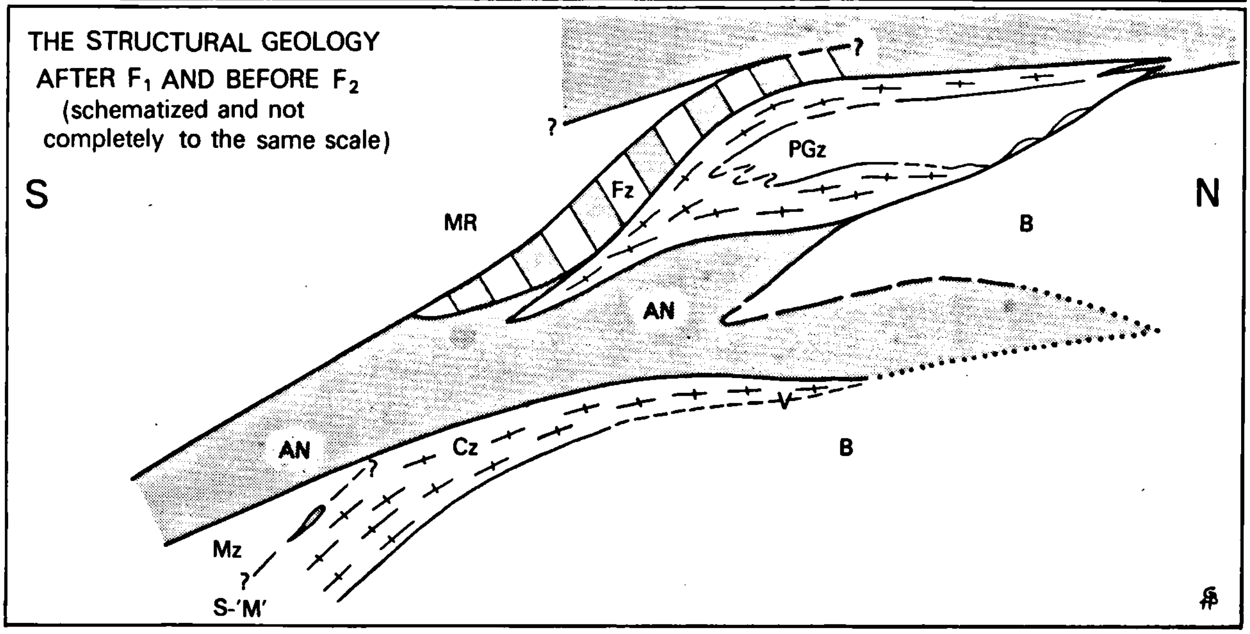
7.3. THE 'ROOT-ZONE' – AN F_3 STRUCTURE

In 5.4.1 it has been demonstrated that the overturning of the 'root-zone' in the present area is the result of F_3 folding by the Brevettola Antiform. This fold is believed to be continuous with the Vanzone Antiform (see 1.3.3), and the arguments for this are:

1. Laduron (1974) considers the Vanzone Antiform as an F_3 fold in a locally established sequence F_1, F_2 and F_3, though Laduron & Merlyn (1974) state that the Vanzone Antiform has been initiated during F_2. The three phases recognized by Laduron (1974) have the same properties as the three folding phases of the present area, and they are likely to be the same phases. Both antiforms are postcrystalline (Laduron, 1974;

Fig. 7-1. A profile through the present area and a tentative reconstruction of the structural geology post-F_1 and pre-F_2. See text (7.4) for explanation. Symbols: B: Bernhard Nappe; PGz: Portjengrat zone; Fz: Furgg zone; MR: Monte Rosa Nappe s.s. and Monte Rosa zone; AN: Antrona amphibolites; Cz: Camughera zone; V: Verosso gneisses; S-'M': Salarioli-'Mulde'; Mz: Moncucco zone. The profile and the reconstruction are derived from Enclosure 1 and are not in one plane. The dots indicate the structural zones where S_1 is steeply dipping or has been inverted as a result of F_2 folding. The hatched area is the 'root-zone' region of steeply dipping and inverted S_1, brought into that position by F_3 folding. See Fig. 5-1 for a schematic map showing the structural zones.


THE STRUCTURAL GEOLOGY AFTER F3
S
N
Axial plane Mischabelrückfalte ,F2
B
Structural zone 1
Mittaghorn Synform ,F2
PGz
Structural zone 2
MR
Axial plane Trifthorn Antiform, F2
Fz
Structural zone 4
AN
AN
Structural zone 3
V
Structural zone 6
Cz
Mz
Axial plane Antrona Syncline, F2
Structural zone 5
Axial plane Brevettola Antiform, F3


THE STRUCTURAL GEOLOGY
AFTER F1 AND BEFORE F2
(schematized and not
completely to the same scale)
S
N
?
?
PGz
Fz
MR
B
AN
V
AN
Cz
B
Mz
?
S-'M'

Laduron & Merlyn, 1974; this paper) and the orientation of the F_3 fold axes and axial planes is not very different.
2. Both folds form the transition between the nappes with their interlayered Mesozoic rocks and the 'root-zones'. The inferred connexion between their axial planes passes southeast of Antrona–Schieranco, where in the Camughera zone the present author found F_2–F_3 fold interference and asymmetric F_3 folds indicating an antiform in the southeast. It is likely that the axial plane runs close to the boundary between the Camughera and Moncucco zones. No evidence was found for the continuation of the Vanzone Antiform as suggested by Bluementhal (1952; see 1.3.3).

7.4. RECONSTRUCTION OF THE STRUCTURAL GEOLOGY BEFORE F_2

The reconstruction in Fig. 7-1 is very approximative because the strain state of the rocks during F_2 is unknown. It serves only to show the approximate relations between major structural units before F_2 and after F_1.

It has been assumed that before F_2 the nappe boundaries were shallow dipping or recumbent, corresponding to low-angle thrusts or recumbent overthrusts of slices of basement rock (see also Bortolami & Dal Piaz, 1970) with S_1 parallel to the thrust planes. The nappe boundaries, that are supposed to have been nearly planar before F_2, consist either of a single major thrust plane (Saas Fee-Zwischbergen boundary), of a movement zone where the deformation was not limited to a single plane and has probably been penetrative (Furgg zone), or of rock units such as the Antrona amphibolites where the zones of nappe movement could not be localized.

Unfolding of the F_2 and F_3 folds gives the following results. The Saas Fee–Zwischbergen boundary shows that the Portjengrat zone overlies the Bernhard Nappe, but for a very short distance only. The Portjengrat zone also overlies the northern branch of the Antrona amphibolites. Where the Furgg zone separates the Monte Rosa Nappe s.s. from the Portjengrat zone, S_1 is in inverted position and has a northerly to northwesterly dip (structural zone 3). Before F_2, S_1 was in non-inverted position, and the Furgg zone consequently must have been horizontal or have had a small southerly dip, and have covered the Portjengrat zone. The same reasoning makes it clear that the frontal part of the Monte Rosa Nappe s.s. has covered the Furgg zone. Only the augengneiss of Saas Fee and the Mesozoic cover of the Portjengrat zone, in the Mittaghorn and south of it, are in non-inverted position (structural zone 2), and they retain their orientation in the reconstruction.

The maps of Bearth (Geol. Atlas der Schweiz, Blatt Saas, 1954) and Blumenthal (1952) show that the eastern part of the Portjengrat zone is bounded by Mesozoic rocks and the Palaeozoic cover rocks of the Furgg zone. In the reconstruction, the Portjengrat zone becomes a large lens that is bordered on all sides by cover rocks or syn-nappe thrust planes. Evidence for internal movement zones has also been found (fold 9, 5.2.2) but they are of subordinate importance. The augengneiss of Saas Fee and the large orthogneiss body of the Mittelrück and the Furggtal lie at almost the same structural level, and they may have been continuous before F_2. They cover the central mass of the paragneisses, that disappears towards the east, and these in turn lie above the large body of orthogneisses in the northern slopes of the Almagellertal.

In the reconstruction, the shape of the Portjengrat zone resembles a large rootless F_1 fold, but this assumption is not supported by the facts. The asymmetry of most F_1 folds in the Portjengrat zone indicates a closure to the east (that is, in the south in Fig. 7-1). The other asymmetry occasionally occurs close to the Furgg zone. Symmetrical folds are exposed below Alpe Campo in the Val Bognanco, close to where the Portjengrat zone terminates. These observations do not contradict the possibility that the Portjengrat zone should be an F_1 fold, but the fact that one asymmetry predominates, also where the section of Fig. 7-1 suggests that the other limb of the supposed F_1 fold should be found, is a strong argument against it. The cause for the predominance of one of the two possible asymmetries of F_1 folds is discussed in 7.5.

Unfolding of the Antrona Syncline does not significantly change the orientation of its southern limb that separates the Monte Rosa Nappe s.s. from the Camughera–Moncucco Complex, but the steep dip and inverted position of S_1 in its northern limb (structural zone 3) is suppressed. The reconstruction shows that the Camughera zone and the Verosso gneisses have the same structural position below the Antrona amphibolites and are likely to be connected under their amphibolite cover.

This connexion, that has been made before (see 1.4.3) involves a change in lithology (Bearth, 1956a). The Verosso gneisses consists of infrequent orthogneisses and granite and dominant paragneisses of the Mischabel–Siviez type, but orthogneisses prevail in the Camughera zone, and Bearth (1956a), although admitting the geometrical possibility of this connexion, rejects it on these grounds.

The orthogneisses occur, however, in the same structural position; in the Camughera zone they adjoin the Antrona amphibolites, and in the Verosso gneisses they are only locally separated from the Antrona amphibolites by some tens of metres of paragneiss (see also Geol. Atlas der Schweiz, Blatt Saas, Bearth, 1957). The geometry of the Camughera parasitic fold can be explained by thinning of the orthogneisses towards the north (see 5.4.1), and it is recalled that the thickness of the orthogneisses in the Camughera zone before F_2 is much less than their apparent thickness now, because of F_2 folding. Therefore, it is believed here that the necessary change in lithology actually does occur, and that the Antrona amphibolites mask a connexion between the Bernhard Nappe (Verosso gneisses) and the Camughera zone, which east of the Antrona Syncline has been removed by the Simplon–Centovalli Fault.

The present structural sequence in the Val Brevettola section is Moncucco zone–Salarioli-'Mulde'–Camughera zone, from the highest unit downwards. This sequence is not altered by the unfolding of the Brevettola Antiform, since north of the ultramafic body where it is exposed, the rocks are not yet in the overturned position of the 'root-zone'. Neither does reversal take place when the Camughera parasitic fold (F_2) is unfolded, for the sequence is contained in the latter's southern, non-inverted limb. It is concluded that after F_1, the Moncucco zone must have covered the Camughera zone and the interjacent Salarioli-'Mulde'.

These relationships contradict the generally held opinion (Argand, 1911, 1916; Blumenthal, 1952; Bearth, 1956a, 1958, 1964) that the Camughera zone is the upper unit. It may be that F_1 folding has inverted the sequence but it is also possible that it is primary and that the Moncucco zone covers the Camughera zone, in which case the lithology of the Moncucco zone and its presence as the cover of basement rocks suggest some analogy with the Furgg zone. It has not been possible to choose between these possibilities, and this problem must await further investigation.

7.5. NAPPE BOUNDARIES, NAPPE FORMATION AND F_1 FOLDING

Monte Rosa Nappe s.l. – In recent literature and on recent geological maps (see 1.2.1 and 1.4.2) the Portjengrat zone is considered as the frontal element of the Monte Rosa Nappe s.l. This is confirmed in the present paper, where the Saas Fee–Zwischbergen boundary has been presented as an almost continuously mappable feature (see 5.2.7). Moreover, the prominent orthogneisses in the Portjengrat zone, and its cover of Furgg zone rocks which characteristically occur in the Monte Rosa Nappe s.s. (Dal Piaz, 1966; Wetzel, 1972) demonstrate strong lithological affinities between both units.

The Monte Rosa Nappe s.s. contains abundant pre-Alpine mineral associations in almost undisturbed pre-Alpine rocks that are separated by zones of Alpine deformation and mineral growth, where an Alpine schistosity developed (Bearth, 1952a; Dal Piaz, 1964, 1966, 1971; Laduron, 1974). Relic pre-Alpine rocks and mineral associations are absent in the Portjengrat zone, where penetrative S_1 is omnipresent, suggesting that the Monte Rosa Nappe s.s. and the Portjengrat zone became disconnected at an early date, before the development of S_1, and that the structural history of both units has been different since then. This is in agreement with Roesli (1946) who describes the Portjengrat zone (with a slightly different boundary, see 1.4.2) as a more or less individual element between the Monte Rosa and Bernhard Nappes.

It is therefore concluded that the Saas Fee–Zwischbergen boundary separates the Monte Rosa Nappe s.l. from the Bernhard Nappe, and that the Portjengrat zone is the most strongly deformed portion of the Monte Rosa Nappe s.l. which before the formation of S_1 became detached from the Monte Rosa Nappe s.s.

The prevailing Palaeozoic rocks in the Furgg zone contain intercalations of Mesozoic rocks that Wetzel (1972), after Bearth (1957a) explains as 'Einspiessungen' (piercements) of underlying 'Mulden' (synclines) of Mesozoic rocks, but which are characteristically found as lenses and intermittent layers (Wetzel, 1972). The Furgg zone has been subjected to strong Alpine deformation associated with intensive boudinage on every scale (Bearth, 1957a; Wetzel, 1972; this paper). F_1 together with later folding has been important, but in the studied part of the Furgg zone, Wetzel's statement holds only so far that in some cases the lenses are rootless F_1 folds. Here, it is suggested that these Mesozoic rocks form a portion of a Mesozoic sequence covering the Furgg zone, which subsequently was deformed simultaneously with the Palaeozoic rocks.

The reconstruction of Fig. 7-1 shows that the Monte Rosa Nappe s.s. overlies the Furgg zone and a large portion of the Portjengrat zone. In an analogy with Dal Piaz (1966) who explains the structure of the Monte Rosa Nappe s.l. as a number of overthrust slabs of basement rock, it is thought here that the Monte Rosa Nappe s.s. was overthrust on the Portjengrat zone, with the interjacent Furgg zone as the movement zone. This is confirmed by Wetzel (1972) who argues that the Furgg zone is an 'intrakristalliner Scherhorizont' (intercyrstalline shear horizon), and it explains the particularly strong Alpine deformation in the Furgg zone. The intense F_1 deformation, resulting in a pervasive S_1, numerous F_1 folds and wide-spread boudinage, in the Stelli, Furgg and Portjengrat zones is partly explained by continued overthrusting, but a slight increase in temperature during the final stage of nappe emplacement is believed to have been an associated factor in the formation of penetrative F_1 structures in the frontal part of the Monte Rosa Nappe s.l. (see below).

Relations between nappe emplacement and the development of S_1. – In the Gran Paradiso Massif, Callegari et al. (1969) observe how the effects of Alpine deformation are concentrated in the planes and zones of Alpine movement between large slabs of almost unaffected Hercynian basement rock, and they describe how in these movement zones, Alpine minerals grow in porphyritic, more or less cataclastic rocks derived from previously unfoliated Hercynian granites. The igneous plagioclase is replaced by aggregates of albite, and a quartz-rich vs. plagioclase-rich layering may come into existence. There is replacement of alkali feldspar porphyroclasts by albite, and a replacement of igneous by metamorphic biotite, and the subsequent formation of a schistosity ('tessitura scistosa') leads to the development of orthogneiss with augen formed by the remains of alkali feldspar.

The structural and petrographical observations of Bearth (1952a) and Dal Piaz (1964, 1966) in the Monte Rosa Nappe s.s. are in good agreement with the observations of Callegari et al. (1969), and the orthogneisses of the Portjengrat zone could have been formed in the same way as the orthogneisses of the Gran Paradiso. Apart from the early biotite growth, which did not take place in the Monte Rosa Nappe s.l. (Bearth, 1952a), the growth of albite, the replacement of microcline by albite, and the formation of a compositional layering would have a place in the sequence of events which fits the observations in the Portjengrat zone. Callegari et al. (1969) do not, however, describe penetrative early Alpine deformation from the interior of the slabs, and the same applies to the largest part of the Monte Rosa Nappe s.s. (Bearth, 1952a; Dal Piaz, 1966). In contrast, the Portjengrat zone and the remainder of the frontal part of the Monte Rosa Nappe s.l. have been penetratively deformed to the extent of eliminating most traces of individual slabs and movement zones, and when Callegari's mechanism is applicable, all orthogneisses there must have once been cataclastic and have recrystallized during the formation of S_1. Moreover, it would be unrealistic to exempt the rocks interlayered with the orthogneisses – the paragneisses and the Furgg zone rocks – from having been subjected to this deformational history, and in this way the frontal part of the Monte Rosa Nappe s.l. would in fact become a large-scale movement zone itself. Though intensive F_1 deformation cannot be denied, the latter conclusion seems improbable, and it is believed here that a slightly higher temperature in the frontal part of the Monte Rosa Nappe s.l. made the pervasive deformation possible. An argument for this is the similarity between the paragneisses in the Portjengrat zone and in the zone of Mischabel–Siviez, where pre-Alpine mineral associations fail almost completely (Bearth, 1964), while they still persist in the Monte Rosa Nappe s.s. where the metamorphic grade of the second Alpine metamorphic event was at least as high. This suggests a somewhat higher temperature during the formation of S_1 north of the Monte Rosa Nappe s.s.

The change from a cataclastic to a recrystallized microstructure, as described by Callegari et al. (1969) could by itself be evidence for a temperature increase, unless it has been caused by a change from initially very high to lower strain rates at a more or less constant temperature during overthrusting. Because of the location of the Monte Rosa Nappe s.l. between two mafic rock units representing former oceanic crust (Bearth, 1974; they are the zone of Zermatt–Saas and the Antrona amphibolites), it might be possible to explain the different structure in the frontal and rear parts of the Monte Rosa Nappe s.l. in terms of plate tectonics, with its frontal part representing the forward portion of a subducted remnant of continental crust. This would be the more strongly deformed part and also have been subjected to the highest temperature, linking the present author's ideas with the formation of the orthogneisses as demonstrated by Callegari et al. This possibility, however, cannot be proved with the information available to the author at this stage.

From the preceding text it is evident that in the Monte Rosa Nappe s.l. two stages of nappe emplacement can be recognized. The first is a stage of non-penetrative deformation on the scale of the nappe, when all movement is concentrated in zones and planes between almost unaffected slabs of basement rock (see Callegari et al., 1969; Bearth, 1952a; Dal Piaz, 1964, 1966; Wetzel, 1972). The second stage is characterized by penetrative deformation during which a pervasive S_1 and numerous F_1 folds were formed. In places, this stage affects only the movement zones (rear part of the Monte Rosa Nappe s.l., see references above); elsewhere, it obliterates the former structure of basement slabs and movement zones (frontal part of the Monte Rosa Nappe s.l.), depending on physical conditions such as temperature, bulk strain, strain rate, etc. The F_1 folds in the Portjengrat zone demonstrate that this second stage is related to nappe emplacement, since they have one asymmetry almost to the exclusion of the other, and so demonstrate unidirectional movement throughout the whole of the folded unit. The F_1 folds do not preferentially occur in the evident syn-nappe movement zones such as the Mesozoic metasediments.

Of the two stages in the formation of the Monte Rosa Nappe (Fig. 7-2) the first is believed to have been the most important regarding the transport of the nappe, the second has caused most of the pre-F_2 deformation of the nappe (see also Ayrton & Ramsay, 1974).

It has not been possible to relate the emplacement of the nappes with the axial directions of the F_1 folds. The

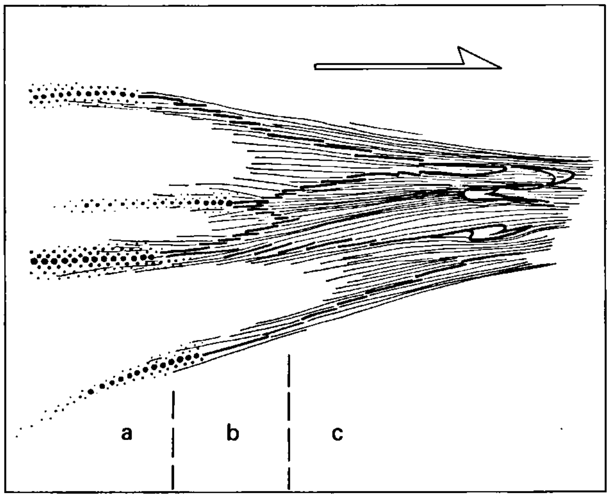


Fig. 7-2. Structural development of the thrust sheets (nappes) in the present area. The shown sequence of events exists in time and place. (a) Relatively cool thrust sheet, no penetrative deformation. This is the early stage of development of the Monte Rosa Nappe, preserved in the Monte Rosa Nappe south of the Stelli zone (see Bearth, 1952a; Dal Piaz, 1964, 1966; Callegari et al., 1969). (b) Transitional stage. (c) Penetrative deformation associated with a temperature increase leads to the development of S_1 and F_1. Later stage of development of the Monte Rosa Nappe, exemplified by the Portjengrat zone and the Stelli zone.

F_1 fold axes in the marginal parts of the Portjengrat zone are perpendicular to the generally inferred northward movement of the nappes, but in the central part of the Portjengrat zone the F_1 fold axes are more or less parallel to it (see 3.1, evidence of interference patterns). The margins have been subjected to the more intense F_2 deformation, which might be related to this change, but it is also possible that differences in strain rate, and strain differences in terms of magnitude, time or place during F_1 folding are responsible for the change in axial direction. At present, the F_1 folds are insufficiently known to answer this problem.

It is not inconceivable that folds were formed in the movement zones during the first stage of nappe emplacement, and if so, they are to be expected in the Mesozoic rocks in particular. It is possible that some folds in the Antrona amphibolites belong to this stage (see 2.2) but apart from this, little evidence was found in the present area for the existence of such a pre-F_1 fold generation.

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