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The Glarus Overthrust: Field Evidence and Mechanical Model

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ABSTRACT

The results of field investigations in the Glarus area (eastern Switzerland) are presented to provide a framework for a discussion of the mechanics of movement along the main Glarus thrust. The thrust contact is everywhere marked by a layer of calc-mylonite (Lochseiten limestone) with a thickness of 1–2 metres. The rocks below it show three phases of deformation: 1. early diverticulation, emplacement of the slip sheets; 2. penetrative ductile deformation, major folding and development of penetrative cleavage; 3. complex heterogeneous deformation, affecting the phase 2 cleavage in a zone about 300 m thick below the thrust. Only phase 3 is related to thrusting, although the structures in no way suggest large rotational bulk strain. Similarly, in the rocks above the thrust (mainly Permian "Verrucano" clastics), an intense, pre-thrusting phase of ductile deformation and a much weaker, thrust-related phase can be recognized. Detailed study of the Lochseiten calc-mylonite revealed that on the whole the idea of an extremely ductile mylonite layer along the interface between two more or less rigid blocks taking up all differential movement along the base of the upper block is correct. Except in the south, post-thrusting features are remarkably absent, and the broad arching shape of the thrust plane is probably an original feature. The main phase of thrusting (Miocene) with a minimum displacement of 35 km occurred in a maximal time span of the order of 10 m.y. This movement was taken up by pseudoviscous flow in the 1 metre thick calc-mylonite giving a minimum shear strain rate around 10^{-10} sec^{-1}. Comparison with the results of triaxial tests on Solnhofen limestone reveals that such rocks are much too strong to flow at such a fast strain rate. It is supposed that this discrepance between nature and experiment is mainly due to different deformation mechanisms operating in the two situations.

ZUSAMMENFASSUNG

Die Resultate von Felduntersuchungen im Kt. Glarus (Ostschweiz) sollen den Rahmen für mechanische Überlegungen an der Glarner Überschiebung liefern. Der Überschiebungskontakt ist überall durch eine Lage von Kalkmylonit (Lochseiten-Kalk) begleitet, die etwa 1–2 Meter mächtig ist. Die Gesteine im Liegenden zeigen drei Deformationsphasen: 1. frühe Divertikulation und Platznahme von Gleitbrettern; 2. duktil-penetrative Deformation, Hauptfaltungsphase mit Entwicklung einer penetrativen Druckschieferung; 3. komplexe und heterogene Verformung, welche die in Phase 2 angelegte Druckschieferung in einer ungefähr 300 Meter mächtigen Zone direkt unter dem Überschiebungskontakt ergreift. Nur Phase 3 kann mit der Überschiebungsphase korreliert werden,

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wobei aber diese spätesten Strukturen keinesfalls einem rotationalen Deformationsplan entsprechen. In ähnlicher Weise kann auch im darüberliegenden Gesteinspaket (vor allem Verrucano) eine duktile prä-Überschiebungsverformung beobachtet werden, gefolgt von einer viel schwächeren syn-Überschiebungstektonik. Detaillierte Beobachtungen am Lochseiten-Kalk bestätigen die Idee, dass es sich bei diesem Kalkmylonit um ein extrem verformbares Gestein handelt, welches Differentialbewegungen zwischen zwei, im wesentlichen rigiden Blöcken aufnimmt. Abgesehen vom südlichsten Untersuchungsgebiet fehlen auffälligerweise Spuren einer späteren Deformation, was zur Vermutung führt, dass die Bogenform dieser Überschiebungsfläche primär ist. Die Hauptphase der Glarner Überschiebung (Miozän) dauerte bei einem minimalen Überschiebungsbetrag von 35 km maximal ungefähr 10 Mio. Jahre an. Dieses Vorgleiten wurde durch pseudoviskose Fliessverformung des etwa 1 Meter mächtigen Mylonites ermöglicht, was einer Scherungsrate von ungefähr 10^{-10} sec^{-1} (Verhältnis von Gleitbetrag zu Mächtigkeit der Fliesszone pro Sekunde) entspricht. Der Vergleich mit Resultaten triaxialer experimenteller Verformung von Solnhofen-Kalk zeigt, dass solche Gesteine eine zu grosse Fliessfestigkeit besitzen bei dieser hohen Verformungsgeschwindigkeit. Es wird deshalb vermutet, dass die Diskrepanz zwischen Naturbeobachtung und Experiment darauf zurückgeführt werden kann, dass die Deformationsmechanismen in den beiden Situationen nicht dieselben waren.

CONTENTS

A. Introduction

Observations along the well exposed Glarus overthrust, where older conglomerates and slates of Permian age have overridden younger sediments of upper Cretaceous and Tertiary age (often of flysch type), gave rise to far-reaching hypotheses as early as the middle of the last century. ESCHER VON DER LINTH mentioned the existence of a large overthrust already in 1846 (see ALB. HEIM 1929, p. 218). Unfortunately he never frankly wrote down his ideas, afraid of being taken as a fool. In 1878 ALB. HEIM published his idea of a "Glarus double fold", mainly influenced by his observations on the ductility of rocks during folding and by the presence of an intermediate layer of highly deformed limestone mylonite (the Lochseiten limestone) below the Permian, which he took for the inverted limbs of two huge overturned folds facing one another (see ALB. HEIM 1878, plate 7). BERTRAND (1884), a French geologist, who himself had never visited Glarus, reinterpreted Heim's profiles by postulating a single overthrust because he was more familiar with discontinuous structures.

During the early 20th century, more and more geologists became familiar with the idea of large overthrusts and, as a reaction to the early ideas, the role of thrusting was often overestimated in the Alps. New objections arose mainly from the mechanical viewpoint, particularly regarding the stability of overthrust blocks, starting with SMOLUKOWSKY's work in 1909. All mechanical models of overthrusting up to quite recently were based on the assumption of a rigid block moving over a rigid substratum. The resistance to overthrusting was calculated by considering the "frictional", and by some authors also the "cohesion", term in the well-known Mohr–Coulomb equation for brittle fracture. HUBBERT & RUBEY (1959) took into consideration the effect of pore pressure in lowering frictional resistance.

It was HSÜ (1969*a*) who first proposed an alternative treatment of the resistance to overthrusting, by introducing a thin layer of ductile material between the two blocks, inspired by the 1–2 metres thick band of mylonitic limestone along the Glarus overthrust. Only during a later phase did he call upon minor thrusting along a cohesionless shear plane, as is seen clearly cutting through the Lochseiten limestone at the type locality near Schwanden, Glarus (see HSÜ 1969*a*, Fig. 1). KEHLE (1970) proposed a similar model, assuming pseudoviscous flow of much thicker intermediate "layers". It is interesting to note that these new approaches are a step towards the early ideas of ALB. HEIM, who never believed in a completely detached overthrust.

It is the aim of this contribution to provide field observations in the particular case of the Glarus overthrust east of the Linth valley and to confront these data with possible mechanical models. Many of the observations are valid only in this very special case. In particular not all thrust planes are associated with a mylonitic layer and even less show clear evidence that mylonite formation is synchronous with and directly related to overthrusting.

The field work was carried out with the following problems in mind:

a) An important assumption behind the mechanical analysis of overthrusting concerns the rigidity of the substratum below the thrust plane. As KEHLE (1970) pointed out, displacement of a thrust block could also be achieved by simple shear flow of a décollement zone several hundreds of metres thick. So the question arises whether the highly deformed and incompetent "Flysch" units below the Glarus thrust represent such a décollement zone or whether all horizontal displacement was taken up solely by movement on the thrust plane and/or flow in the Lochseiten limestone. In the first case, folding below the thrust plane should be synchronous with and directly related to the thrusting. This conclusion has, in fact, been drawn by SIEGENTHALER (in press) and partly also by WERNER (1973), who both carried out structural investigations in the Glarus area. If this view is right, net displacement along the Glarus overthrust could be reduced by a considerable amount, depending on the bulk shear strain in the décollement zone. It will be shown however that most of the internal deformation in the "Flysch" units pre-dates overthrusting.

b) Since the overthrust block also suffered internal deformation the question of how this deformation relates to overthrusting arises here again. In particular it is important to know whether the overthrust block remained stable during overthrusting or not.

c) The thrust contact was followed over the whole area of outcrop in order to obtain a representative picture of the nature of this discontinuity. In particular the question of the relative importance of the two mechanisms mentionned above arose: thrusting by flow of the intermediate limestone layer and later frictional movements along a post mylonitic shear plane. No important traces of such a second frictional phase of overthrusting were found however.

d) Since both TRÜMPY (1969) and HSÜ (1969*a*) believe that upheaval of the Aar massif caused arching of the preexisting overthrust plane inducing later frictional sliding along the northdipping slope of 10° or more it was important to know whether the shape and orientation of the thrust plane were really affected by later differential vertical movements or not. It was found that essentially one single phase of thrusting along an arched surface largely unaffected by later tectonism is more likely.

B. General geology

A modern description of the geometrical pattern of the Helvetic nappes in eastern Switzerland, together with a kinematic interpretation, has been given by TRÜMPY (1969), so only a brief summary of the geological setting is needed here (see Fig. 1).

The main Glarus thrust separates the Permian Verrucano at the base of the Helvetic nappe complex very clearly along a single plane from the underlying heterogeneous tectonic units. Secondary thrust planes subdivide the different Helvetic nappes mainly at a higher stratigraphic level and towards the front of the Helvetides (see Fig. 2).

The units below the main Glarus thrust can be subdivided into the following tectonic and paleogeographic units, also separated by thrust planes:

1. Parautochthonous and autochthonous cover of Aar massif, consisting a) of strongly folded Mesozoic and Eocene formations with associated minor thrusting and b) a Tertiary sandstone Flysch sequence (North Helvetic Flysch), partly stripped off 1a.

2. Blattengrat (South Helvetic) and Sardona (Ultrahelvetic or Penninic) slip sheets superimposed on 1 at an earlier stage by stripping off of the uppermost stratigraphic levels from the area far south of 1, in the case of the Sardona unit even south of the area which later formed the Helvetic nappes. Those two slip sheets, although only partly of flysch type, and unit 1b are grouped together as the "Flysch" units in the following.

3. Allochthonous slices of mostly Mesozoic cover of the Aar massif – the Subhelvetic nappes – also transported northwards before the main Glarus thrust formed.

It is important to stress that none of the secondary thrust planes below and above the main Glarus thrust seem to be connected geometrically with the main thrust itself, where exposed. The frontal part of the main thrust is unfortunately not exposed and the frontal thrust of the Helvetic nappes along the Alpine border is probably not directly connected with it (Fig. 2).

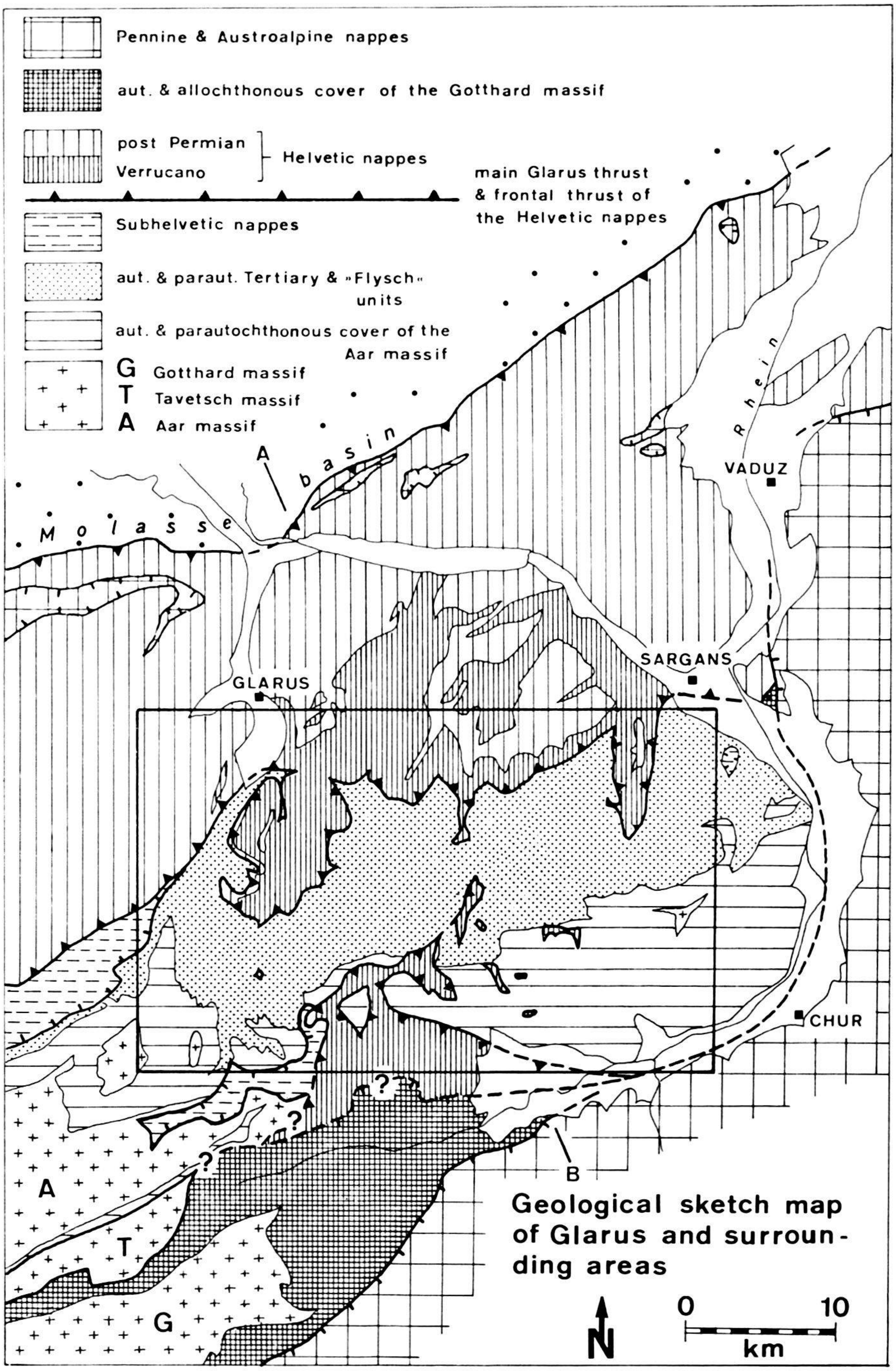


Fig. 1. Geological sketch map of Glarus and surrounding areas.

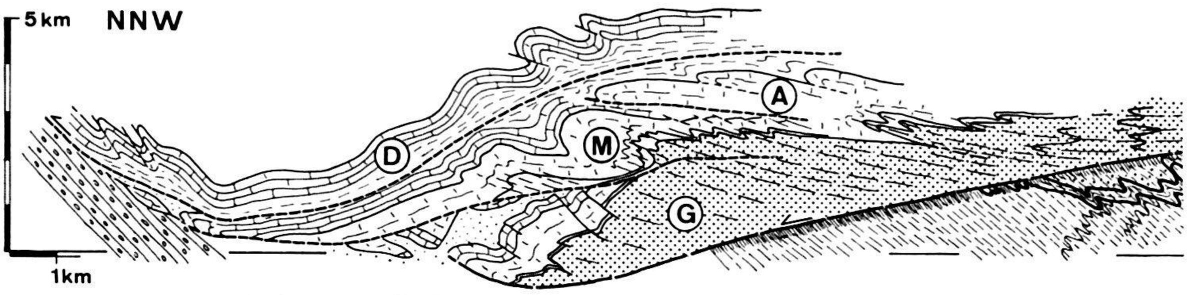


Fig. 2. Schematic cross section of Glarus overthrust (mainly after OBERHOLZER 1933 and TRÜMPY 1969). For the trace of the profile see line labelled A–B in Fig. 1.

C. Field evidence

1. Structure in the "Flysch" units

Detailed studies were carried out in the "Flysch" units mainly in a small area on the western side of Sernftal between Engi and Elm (for localities mentioned in text, see Fig. 10). Here it was possible to distinguish three phases of tectonic activity, only the latest being related to the overthrusting. Judging from more isolated observations in other areas, this analysis is also representative for the rest of the "Flysch" terrain in the Glarus area.

a) Phase 1: Diverticulation of the "Flysch" units

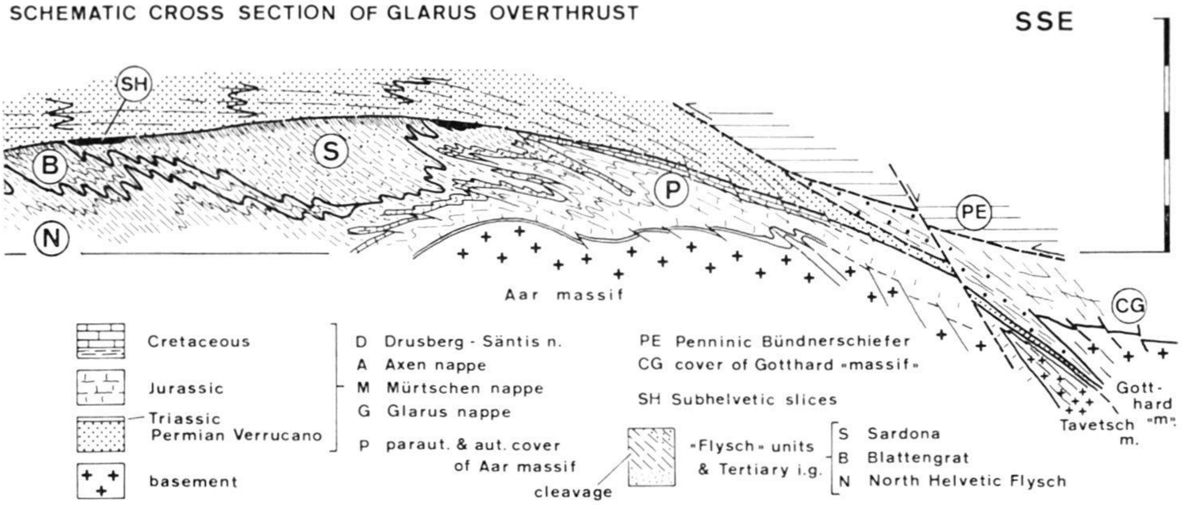
The Blattengrat and Sardona slip sheets were emplaced at an early stage on top and even north of the sedimentary pile which later formed the Helvetic nappes (see TRÜMPY 1969), and were later overridden by them during movement on the main Glarus thrust.

The tectonic contacts related to this early phase of probably gravitational sliding are clearly refolded during a later penetrative deformation (phase 2, see later). Where tectonic contacts of the Blattengrat slip sheet with the parautochthonous sandstone Flysch are exposed, the later penetrative cleavage clearly crosscuts the contact, which itself is often oblique to bedding planes in the sandstone Flysch. No signs of cataclasis or mylonite formation were found along those old thrust planes either because transport was aided by high pore pressures in a still unconsolidated and wet sediment or because of the overprinting by the second phase deformation.

Judging from the high variability of fold axis orientation in phase 2 folds it must be assumed that the bedding planes were at least tilted into different orientations during this first phase. According to BISIG (1957) the large scale piling up of nummulitic limestone layers inside the Blattengrat unit may be at least partly due to large scale isoclinal folding and/or imbrication during this earliest phase.

b) Phase 2: Folding associated with a penetrative cleavage

Small scale structures of this phase are the dominant feature of the whole complex except for a zone immediately underlying the main Glarus thrust which was strongly affected by later deformation (phase 3, see later).



Almost ideal similar folds (half wavelengths in the 1–100 metre size order) are associated with a penetrative axial plane cleavage in shaly and silty layers. Coarser grained sandstone layers show spaced and refracted cleavage planes. Nowhere were displacements of bedding planes along cleavage planes observed. The sandy layers often fall into class 1C, incompetent shaly layers into class 3; this leads to an overall geometry of class 2, similar folds (fold classification after RAMSAY 1967). Second order folds are very scarce. In the more uniform shales of the Blattengrat slip sheet, bedding is completely overprinted by cleavage.

The interlimb angle is often around 35° but the inverted limb forms a much smaller angle (less than 10°) with the axial plane than the normal limb does (Fig. 3). As

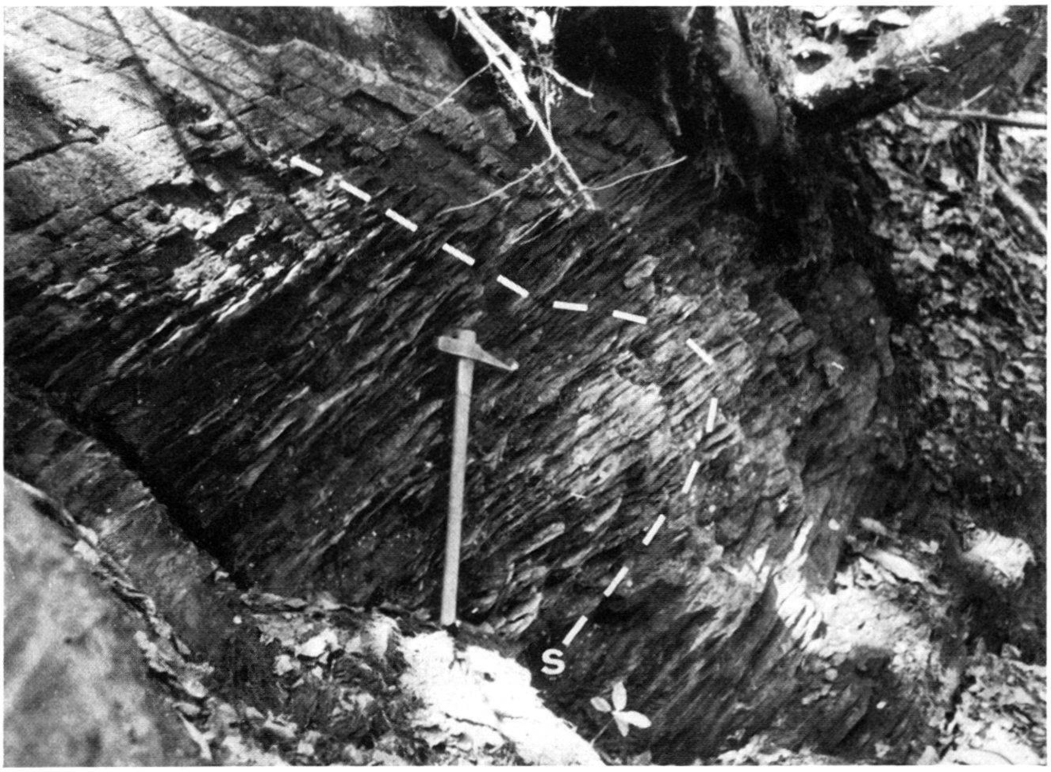


Fig. 3. Similar fold in Engi slates (North Helvetic Flysch). Note increased intensity of the axial plane cleavage near the hinge and in the inverted limb (S = bedding).

a consequence of this, shortening perpendicular to bedding is more intense in the inverted limbs, where bedding and cleavage planes almost coincide. The well known roofing slates ("Dachschiefer") from Engi were extracted from the inverted limbs of these folds. No lineations related to this phase were found except for the well developed cleavage/bedding intersections, but WETTSTEIN (1886) derived stretching subparallel to the dip azimuth of cleavage planes indicated by deformed fishes (X:Y axial ratio between 1.3 and 2).

The phase 2 folds face northwest, and where the "faltenspiegel" can be distinguished (i.e. on the normal limbs), it dips generally northwest (SIEGENTHALER, in press). Cleavage and axial planes have a strikingly constant dip direction (southeast) but a somewhat more variable dip angle, with a mean orientation 147/40 (see Fig. 4A). No systematic change in the dip was observed approaching the main thrust, except for occasional more steeply inclined cleavage planes producing increased scattering of poles (Fig. 4B). Under the likely assumption that these planes mark a plane perpendicular to maximum shortening, a uniform distribution of strain results.

The fold axes define a nearly complete great circle identical with the mean axial plane orientation. This high variability of fold axis orientation indicates a variable

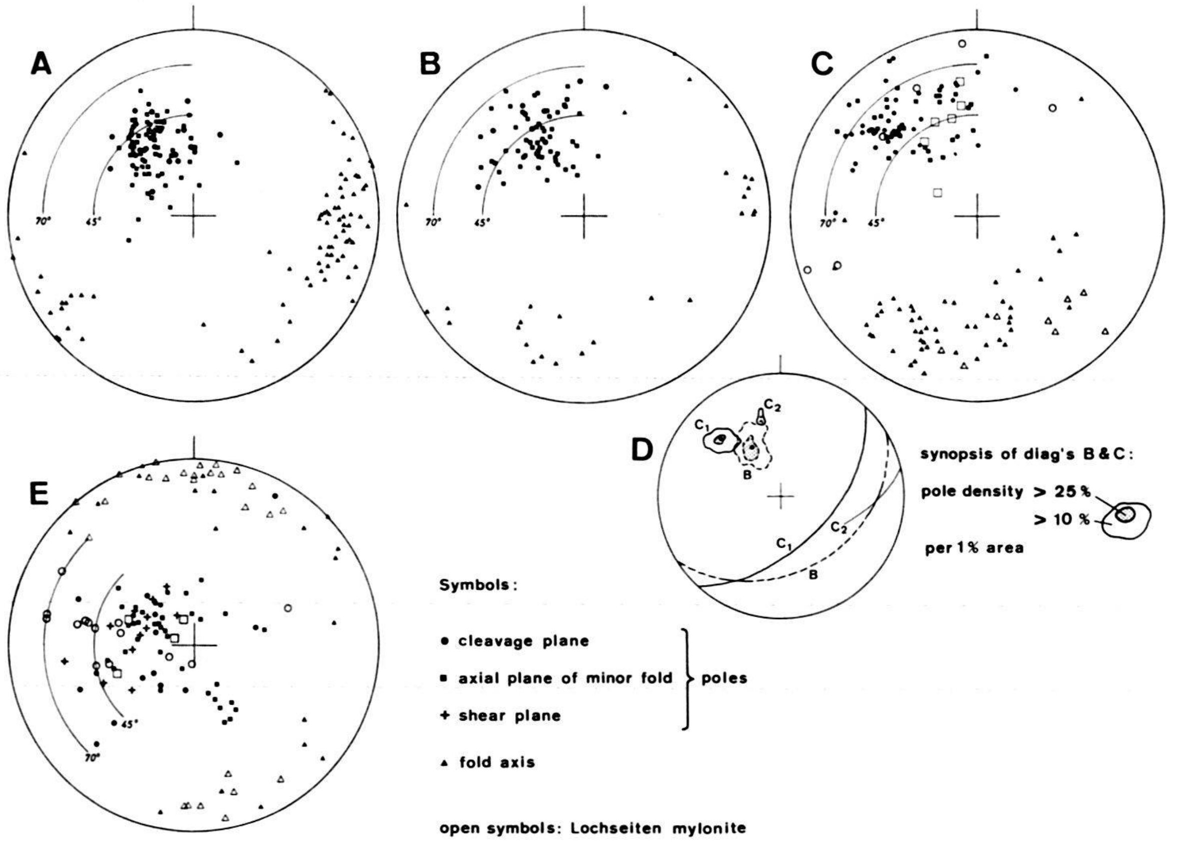


Fig. 4, A/B. Petrofabric diagrams of phase 2 structures.
A: measurements made more than 300 metres below the overthrust, where phase 3 deformation is very weak or absent. B: uppermost 300 metres, in places where phase 2 structures are still recognizable.
Fig. 4, C/E. Petrofabric diagrams of phase 3 structures near the overthrust plane.
Measurements in diagrams A, B and C from the area west of Sernft valley between Engi and Elm, mainly in the North Helvetic Flysch unit. Measurements in diagram E from the Lochseiten type locality, near Schwanden. Equal-area projection, lower hemisphere.

orientation of bedding planes previous to folding also inside the North Helvetic Flysch unit.

c) Phase 3: Deformation related to overthrusting

Structures which clearly postdate the penetrative cleavage start to appear at a level around 300 metres underneath the main Glarus thrust and are absolutely dominant in the uppermost 50 metres. This alone suggests a strong relationship of this phase 3 deformation with overthrusting. In lower tectonic levels only a few and isolated traces of post-cleavage deformation are found (calcite-filled joints and internal boudinage, mainly in the slaty Engi-"Dachschiefer"; conjugate kink bands, in the Blattengrat shales). Phase 3 structures vary strongly from place to place, depending on the lithology and on the orientation of earlier structures. The deformational style strongly contrasts with phase 2 in that deformation was much less ductile and pervasive, and seems to have been accompanied by redistribution of calcite. This contrast in style allowed for a distinction between structures of the two phases in places where the age relationship was not directly deduced.

It is somewhat difficult to name planar structures developed during phase 3, since many transitions are observed. The planar structures plotted in Figure 5, and named

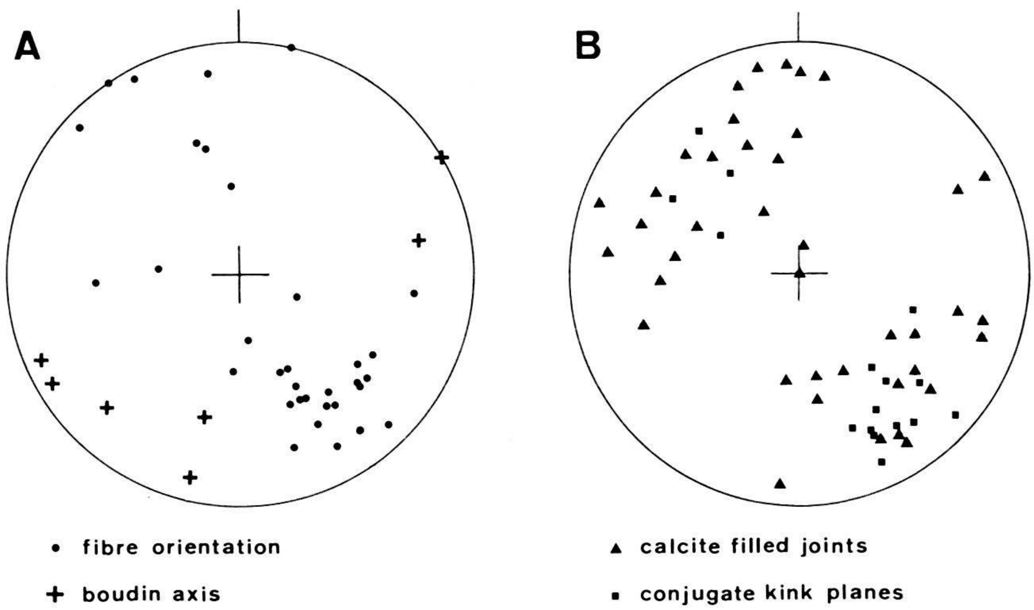


Fig. 5. Petrofabric diagrams of phase 3 structures from the area west of Sernft valley between Engi and Elm, mainly in the North Helvetic Flysch unit. Equal-area projection, lower hemisphere.

"calcite-filled joints", often appear in badly oriented conjugate sets and mostly in slaty horizons (Fig. 6D). The older cleavage planes are often dragged and bent near such joints. Because of this, and since they show the same orientations as conjugate kink bands (the latter are developed mainly in the Blattengrat shales), they are best interpreted as poorly developed conjugate shear planes with later calcite infilling (possibly due to suddently raised pore pressure). True tension joints, filled with calcite fibres and without any bending of older cleavage planes, are also found. The orientation of these fibres, which may be sometimes slightly curved, together with slickenside-like fibres overgrowing old cleavage planes, is plotted in Figure 5.

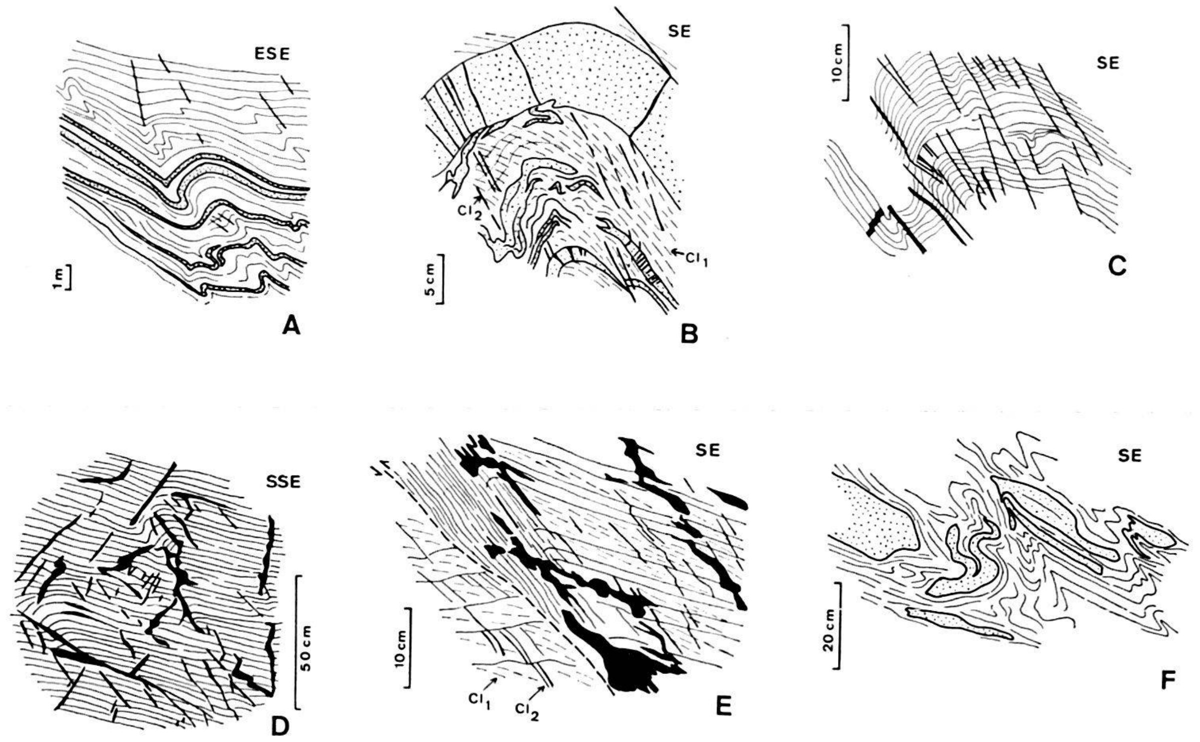


Fig. 6. Sketches of phase 3 structures. Black areas = calcite infillings: for further explanations, see text.

The planar structures plotted in Figure 4C are best described together with phase 3 folds starting further away and approaching the main thrust. Most of the following observations were made on the southeast flank of Kärpf, west of Elm.

About 300 metres below the overthrust spaced, subparallel sets of planes, sometimes calcite-filled, appear. They are named and plotted in Figure 4C as cleavage planes because they grade into crenulation cleavage or axial plane cleavage of sinusoidal folds in higher levels. They always crosscut phase 2 cleavage planes at higher angles of dip and develop preferably in places where the older cleavage planes dip at relatively low angles.

In an intermediate zone, beautiful crenulation cleavage with pressure solution along cleavage planes develops in more slaty horizons, again consistently more steeply inclined than the old schistosity. Where the angle of dip of the phase 2 and 3 cleavages nearly coincides a lensoid, somewhat undulating parting develops. Sinusoidal buckles develop in regions with strong lithological contrasts (sandstone/shale interbedding with thin sandstone layers, Fig. 6A). It is characteristic for these folds (half wavelength varying from a few cm to a few m) that axial planes diverge and converge and can only be traced over very short distances. Tension joints related to the buckling of competent layers are common. The phase 2–3 cleavage relationship is clearly exposed in individual buckles (Fig. 6B). Both cleavages, or the phase 3 cleavages only, often have calcite infillings, which may be buckled again by further deformation (Fig. 6E). This indicates a very complex deformational history during phase 3, although it could not be satisfactorily subdivided into sub-phases. Another type of fold intermediate between true buckles and crenulation occurs in regions with less pronounced competence contrasts but with a strong previous schistosity. This type grades towards a microlithon (Gleitbretter) structure (Fig. 6C). Shearing of folds is observed only exceptionally (Fig. 6E).

In the uppermost 50 metres and especially just below the Lochseiten limestone the tectonic style is rather chaotic. Remnants of phase 2 structures are seldom recognised. The steeply (60–80°) inclined, now penetrative, phase 3 schistosity is often lensoid and curving around boudinaged competent layers or detached small scale buckles (Fig. 6F). Refolded folds are common, indicating that more than one generation developed during this latest phase, since the refolded folds are of phase 3 type. Refolded calcite veins are abundant in this zone. However, here again, it seemed impossible to obtain a coherent subdivision of sub-phases. Probably the zone near the main thrust suffered a continous deformation with redeformation of earlier formed structures. This very chaotic deformational picture seems also to be due to the complete confinement of the "Flysch" below the Lochseiten mylonite layer. The relationship of these phase 3 structures with the mylonite layer will be discussed later.

Calcite fibres strike predominantly SE–NW (Fig. 5) probably best indicating the direction of relative displacements (Durney & Ramsay 1973). Their exact orientation depends on the orientation of the calcite veins filled with fibres. Tension joints, conjugate kink bands and boudinage axes strike SW–NE (Fig. 5). This direction is perpendicular to the orientation of fibres and also to the direction of thrusting along the Glarus thrust inferred from the general geological setting of the Glarus area. Hence the orientation of the intermediate strain axis in the "Flysch" during phase 3 was probably SW–NE and subhorizontal.

Fold axial and cleavage planes have roughly the same dip direction but a higher dip angle than the corresponding phase 2 structures (Fig. 4). This tendency for the formation of a set of steeply inclined cleavage planes during and related to overthrusting is difficult to understand. The only thing it clearly demonstrates is that no significant simple shear deformation parallel to the overthrust plane occurred in the "Flysch" units, since such a flow field would rotate already formed planar structures progressively into an attitude subparallel to the thrust plane. It is also unlikely that the phase 3 cleavage planes represent a set of shear planes because the boudinage of competent layers and fold hinges near the overthrust suggests that the cleavage planes are oriented perpendicular to the direction of maximum shortening. Also occasionally observed displacements along those planes are contradictory. Possibly, in time and space, very variable pore pressures (pressure solution and calcite infillings along the same set of cleavage planes) complicate the strain picture, together with the severe boundary condition that material is completely confined below the overthrust plane, to such an extent that an overall finite strain field cannot be derived at all.

The orientation of fold axes is significantly different from phase 2 axes; they dip steeply southwards (between SE and SW). From the attitudes of the two sets of cleavage planes (Fig. 4D), fold axes would not be expected to be plunging S or SE. Additionally, the SE plunging fold axes have a trend subparallel to the supposed direction of relative displacement, as do also folds in the Lochseiten calc-mylonite (open symbols in Fig. 4). Although an explanation is difficult to find (rotation of fold hinges inside the fold axial plane into a direction parallel to maximum extension is just a possibility), this shows once again that fold axes are misleading in an evaluation of transport direction on the basis of the assumption that their orientation is perpendicular to a "movement direction".

2. *Structure of the parautochthonous cover of Aar massif*

In recent years a series of detailed structural studies have been carried out in the southern part of the Glarus area (BÜRGISSER 1973, FELDER 1973, LAMBERT 1971, STRASSER 1972, PFIFFNER 1972*a*, *b*, and BÜRGISSER & FELDER 1974). The tectonic style of this region contrasts strongly with that of the Helvetic nappes above the main Glarus thrust in that a strong axial plane cleavage is developed, related to large scale, nearly isoclinal folding, often of similar type even in the Mesozoic carbonates, including Triassic dolomites. This highly ductile and penetrative deformational style is comparable to the phase 2 deformation in the "Flysch" units, although the strike (E–W) and dip (less than 40°, usually 20°–30°) are somewhat different (see Fig. 10). It is probable that this deformation is contemporaneous with or immediately followed by a metamorphism (M. FREY et al. 1973), characterized by the paragenesis stilpnomelan–biotite–alkali feldspar in the south.

That this phase of ductile deformation pre-dates the Glarus thrust at least in its final stage is proved by the discordance between the axial plane cleavage, major folds and associated minor thrusts, and the main thrust itself. However, in this area, a well defined zone of postcleavage deformation has not been reported. Only in the Tertiary a second crenulation cleavage, more steeply inclined than the first cleavage, is observed at least in isolated places the author visited. In many places the older cleavage steepens upwards towards the thrust plane (see FELDER 1973) and only in the zone about 1 metre below the Verrucano is it suddenly turned into parallelism with the contact, where it also grades into a mylonitic lamination (see page 263). Hence we reach the same conclusion as in the "Flysch" units, that significant simple shear deformation during overthrusting is confined to the mylonitic layer, about 1 metre thick, below the thrust plane.

3. *The Subhelvetic thrust sheets*

The term "subhelvetische Decken" was introduced by TRÜMPY (1945, 1969) to denote detached thrust sheets or nappes, lying below the main Glarus thrust and composed mainly of calcareous Mesozoic formations of Helvetic facies type. Together with the parautochthonous units, they represent a possible source rock from which the Lochseiten calc-mylonite was derived.

They outcrop in the form of major thrust sheets only west and southwest of the present area (Griesstock nappe, F. FREY 1965; Cavistrau nappe, KÄCH 1969, 1972; thrust sheets in the area of Piz d'Artgas, TRÜMPY 1945). Only a few remnants in the form of isolated lenses are found between the "Flysch" units and the main Glarus thrust east of Linth valley (Foostock, Segnespass, see Fig. 2), where Verrucano forms the base of the Helvetic thrust block. These units seem to have been detached before the main Glarus thrust moved ("frühhelvetische Überfaltungsphase", see TRÜMPY 1969).

Below the main Glarus thrust on the southern and eastern flank of Foostock, a sheet of homogeneous limestone (probably lower Cretaceous "Schrattenkalk") up to 70 metres thick is preserved, quickly thinning out to the west and north (see OBERHOLZER 1933, Atlas table 27, Figs. 4, 6). One has to distinguish in fact two overthrust contacts here, one below the limestone sheet, cutting phase 2 folds in the "Flysch", and one above this limestone sheet, identical with the main Glarus thrust. Only the uppermost 1–2 metres of this limestone sheet are mylonitized (i.e. true Lochseiten calc-mylonite) and this mylonitization post-dates highly ductile straining recorded by deformed ooids. This Subhelvetic limestone sheet was therefore internally strained before thrusting occurred along the top contact and probably also along the base. All this suggests thrusting of this Subhelvetic unit in an intermediate phase between phase 2 deformation in the "Flysch" and final movements along the main Glarus thrust, in front of the advancing thrust block. Comparing this ocurrence with the Subhelvetic Griesstock nappe further west, one has to note that a Lochseiten-type mylonite is absent along its base, whereas it is nicely developed at the base of the Griesstock nappe (F. FREY 1967, p. 665–667).

A similar Subhelvetic lens occurs southwest of Segnespass (Tschingelhörner), there mainly consisting of upper Jurassic limestone. Here again it is completely detached, and shows no geometric relationship with adjacent parautochthonous fold structures (see BÜRGISSER & FELDER 1974). The possibility that these Subhelvetic sheets are the remains of the inverted limb of a huge recumbent fold with a Verrucano core, otherwise drawn out to form the Lochseiten calc-mylonite, is very slight considering the absence of Triassic and lower Jurassic remnants. However another locality (Saasberg), west of Kärpf, seems to support such an idea (see SCHIELLY 1964).

4. The Helvetic thrust block

Where the main Glarus thrust is exposed east of Linth valley, Verrucano directly overlies the thrust contact with the exception of the Kärpf area already mentioned. A subdivision into different nappe units is no more possible near the base of the thrust block since the internal thrusts die out southwards and downwards (Fig. 2). Thus it is appropriate to use the term Helvetic thrust block. The thickness of the detached Verrucano reaches 1500 metres and in most places this sedimentary sequence is in a normal position.

WERNER (1973) carried out structural investigations in the Glarner Freiberge between the Sernft and Linth valleys, mainly in the Verrucano sediments. He distinguished four deformational phases, which I believe can be summarized into two main phases:

Phase 1 and 2: The combined effect of those two phases is to produce a strong axial plane cleavage, now in a subhorizontal position, associated in places with nearly isoclinal folding (see Fig. 2). Additionally stretched pebbles and reduction spots indicate a N–S oriented intermediate strain axis. During WERNER's phase 2 local thrusting occurred.

Phase 3 and 4: Deformation during those phases is essentially brittle (kink-type folding in phase 3 and conjugate minor faulting suggesting an ESE–WNW oriented intermediate strain axis, combined with northward shearing during phase 4).

WERNER comes to the conclusion that his phases 1–3 pre-dated the main Glarus thrust and that his phase 4 was synchronous with overthrusting. Also my own observations, confined to a zone very close to the thrust contact, lead me to the conclusion that cleavage formation in the Verrucano pre-dates overthrusting, since this fabric is affected by later small scale folding and cataclasis near the thrust. However, in most places, the earlier cleavage planes lie subparallel to the thrust plane. Nearly isoclinal large scale folds, with associated subhorizontal axial plane cleavage, are beautifully developed at the eastern flank of Piz Segnes.

This highly ductile deformational style of pre-overthrust age is also found in higher tectonic levels of the Helvetic block, at least up to the Quartenschiefer formation (Triassic) of the Glarus, Axen and Mürtschen nappes. These slates, lithologically analoguous to slates in the Verrucano, show the same penetrative axial plane cleavage with N–S elongated reduction spots (see RYF 1965, MARKUS 1967). Somewhere in the Jurassic level, the traces of this earlier deformation die out and the Cretaceous sediments are bare of any signs of penetrative cleavage formation.

Although detailed structural work in the Helvetic thrust block is still outstanding, apart from WERNER's investigations, and my own observations are very limited, it seems likely that one has to distinguish essentially two deformational styles and phases. Evidence for separate phases mainly comes from investigations on low grade metamorphism in the Helvetic Alps of Eastern Switzerland (FREY et al. 1973 and pers. comm.), indicating that this metamorphism is postkinematic in respect to some internal nappe boundaries within the Helvetic thrust block, but predates clearly the main Glarus thrust.

Phase 1: Pre-overthrust deformation (see phase 1 and 2 above) accompanied by thrusting, prior to and during low grade metamorphism.

Phase 2: Contemporaneous with the main movement on the Glarus thrust, thrusting and major folding at higher levels and in the frontal parts of the thrust block, leading to the final individualization of the different nappes. These internal nappe boundaries, however, have no geometrical connection with the basal thrust plane.

It is premature to clearly assign internal thrust planes in the Helvetic thrust block to one of these two phases but the thrust contacts shown in Figure 2 between the individual Helvetic nappes are probably of syn-overthrust age (or at least reactivated during overthrusting).

An estimate of the thrust block thickness must take into consideration that not only Helvetic Verrucano and Mesozoic formed the thrust block (with a probable thickness of around 3 km, after removal of the overlaps caused by thrusting during the second phase mentioned above) but also sediments of more southerly, Pennine and Austroalpine, origin, now mostly eroded (TRÜMPY 1969). Higher tectonic units are still preserved south of the overthrust culmination, in the Rhine valley north of Ilanz (Ilanzer Verrucano, cover of Gotthard "massif") and north of the culmination in the Säntis area (higher Flysch units). More widespread northward thrusting of Pennine and Austroalpine nappes occurs east of Rhine valley (Fig. 1) and west of our area in Central Switzerland. So the estimate of 5–6 km (TRÜMPY 1969) must probably be regarded as a minimal thickness.

Towards the Rhine valley in the south, where the root zone of the Helvetic nappes is located, the Helvetic part of the thrust block gets very thin and probably was replaced by a now eroded pile of Pennine (and Austroalpine ?) units. The root zone became extremely thinned out after the Helvetic nappes were emplaced (for a discussion of the root problem see TRÜMPY 1963, 1973).

5. The main Glarus overthrust and the Lochseiten calc-mylonite

a) Description of thrust contacts

Roughly three types of thrust contact can be distinguished:

Type 1: A thin layer of Lochseiten mylonite (1–2 m thick) forms the only intermediate layer between "Flysch" units and the Verrucano of the thrust block (Fig. 7). This is the normal situation, as found at the famous Lochseiten type locality near Schwanden.

Type 2: A thicker layer of Mesozoic limestone (up to a few metres) separates "Flysch" and Verrucano. The uppermost one or two metres are mylonitized and show the characteristic lamination (see later). The unmylonitized limestone below is in a similar position to the Subhelvetic thrust sheets already described.

Type 3: Mesozoic limestones of the parautochthonous zone are separated from the Verrucano by 1–2 m mylonite obviously derived from them. Here mylonitization occurred more or less in situ. This type is, of course, restricted to the southernmost exposures.

It is common to all thrust contacts that the calc-mylonite is always present, with a few exceptions only, and never exceeds a thickness of 2 metres, even if more limestone is available in situ (as it is the case in contacts of type 2 and 3). Whereas the predominating type 1 contacts are compatible with the idea that limestone is dragged and smeared out along the base of the thrust block, the other contacts indicate that the thrust block partly overrode limestone occurrences already emplaced before mylonitization started.

Two deviations from the otherwise surprisingly uniform appearance of the thrust contact as a straight and simple subhorizontal cut accompanied by a thin layer of Lochseiten mylonite have to be mentioned. Firstly, Lochseiten mylonite is absent over a longer distance in Chrauchtal (between Gulderstock and Foostock). There, extreme cataclasis, brecciation and complete disintegration of Verrucano and "Flysch" occur in a zone a few tens of metres thick along the thrust contact. This beautifully demonstrates what happens along the thrust contact if differential movement is no longer taken up by flow of the mylonite. In places where the mylonite is present, no such loss of cohesion and brecciation in the surrounding rocks is observed. Secondly, small scale imbrications of already formed Lochseiten mylonite with "Flysch" and/or Verrucano can be observed in a few places. Such an imbrication zone, only a few metres wide, is beautifully exposed north of Foostock.

Inside and towards the top of the mylonitic layer, where the mylonite lamination lies subparallel to the overthrust plane, a sharp planar septum is observed in many places, sometimes filled with a few millimetres of fault gouge (Fig. 7 and HSÜ 1969*a*,

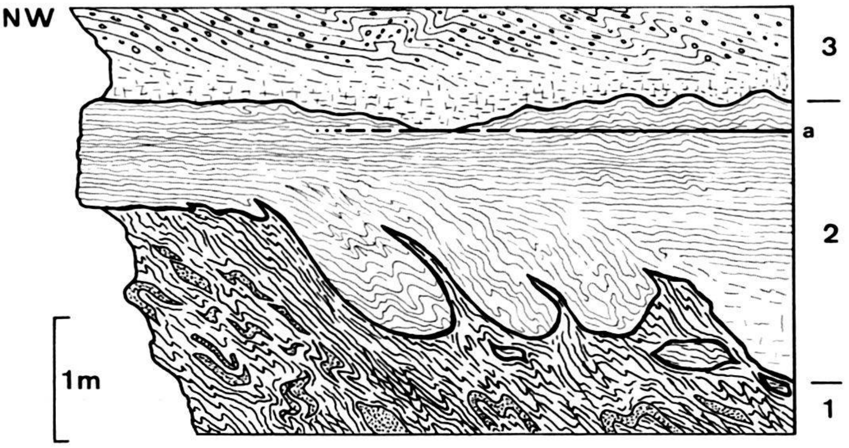


Fig. 7. Composite diagram of a "type 1" overthrust contact showing synoptically the main features.
1: Chaotic phase 3 deformation in the "Flysch", boudins of Lochseiten mylonite.
2: Lochseiten mylonite layer: thin layer without a lobate interface with the "Flysch" (left), development of a lobate interface and folding of the already formed mylonitic lamination at the base (centre) and remnants of unmylonitized limestone at the base (right). a: planar septum often present in zone 2 (see text and Fig. 8).
3: Verrucano clastics, showing folding, cataclasis and destruction of pre-overthrust cleavage at the base.

Fig. 1). However, brecciation of the mylonite leading to cohesionless kakirites, as seen frequently along latest tectonic discontinuities developed under brittle conditions in limestone formations (for instance, near Obstalden along Walensee, see TRÜMPY 1969, pp. 133, 134), was nowhere observed.

Because the traces of post-mylonitic movements under brittle conditions are so weak, they cannot represent a major subsequent phase of frictional sliding over any significant distance (more than a few metres). Additionally, if major post-mylonite movement had occured, there is no reason why this should have taken place always inside the still coherent and, under brittle conditions, very strong Lochseiten mylonite (see p. 277) and not inside the "Flysch" units or inside shaly formations in the Helvetic thrust block which are weaker and by far more suitable for building up high pore pressures. Under low temperatures, the Lochseiten mylonite layer was certainly not a zone of preexisting weakness.

The lower boundary of the Lochseiten mylonite in type 1 contacts often shows a lobate interface with the "Flysch" sediments, with southeast dipping and upwards closed, pinched in Flysch wedges, especially where the mylonite layer exceeds 1 metre in thickness. The wedges never reach the Verrucano/mylonite interface. Geometrically this lobate interface is strikingly similar to the folded interface between two layers with a high viscosity contrast, whereby the Lochseiten mylonite would have the higher viscosity. Often boudins of Lochseiten mylonite are found inside the "Flysch" sediments near the base of the Lochseiten layer also suggesting a higher relative viscosity for the mylonite. Such a deduction however is in complete contradiction to the evidence that Lochseiten mylonite took up major displacement along the overthrust contact separating two rigid blocks, and therefore represents a layer of low relative viscosity.

One could argue that this apparent contradiction shows that one has to assume a two phase model, with relatively ductile and low viscosity behaviour of the mylonite during a first phase, followed by buckling and boudinage of the already formed and now high viscous mylonite at a later stage. This succession of events, however, is still difficult to maintain, since no buckling or homogeneous shortening in a direction parallel to the mylonite/Verrucano boundary is observed.

Structures inside the mylonite layer itself lead to a similar conclusion. The mylonite lamination is folded especially at the base of the Lochseiten layer and inside the lobes (Fig. 7). Those folds show a similar orientation to the phase 3 folds inside the "Flysch" (Fig. 4C), with steeply inclined fold axes and SE dipping axial planes. Also at the Lochseiten locality (Fig. 4E), where the overall orientation of the structures is very atypical (mainly E dipping planars, here suggesting east–west relative displacements, see WERNER 1973), there is a good conformity between structures in the "Flysch" and the mylonite. This, together with rarely observed interfolding of Lochseiten mylonite with "Flysch" gives an additional argument for the phase 3 structures in the "Flysch" being contemporaneous with overthrusting.

The axial planes of these, strictly speaking, post-mylonite folds turn in an upwards direction asymptotically subparallel to the Verrucano/Lochseiten mylonite contact and the only slightly undulating lamination in the top part of the mylonite layer (Fig. 7). This rotation and transition of axial planes into the lamination at the top indicates that folding at the base is not consistently younger than mylonite formation, and that mylonite formation and folding of already formed mylonite must have occurred simultaneously. Geometrically a similar observation is made in type 3 contacts, where steeply inclined (60°–80°) S dipping axial planes of probably syn-overthrust folds suddently turn parallel to the overthrust plane together with a progressive development of the mylonite lamination (Fig. 8, bottom). All this indicates an extremely inhomogeneous finite strain field with rotation of the plane perpendicular to maximum shortening inclined at a high angle at the base into an orientation subparallel to the overthrust zone at the top, where simple shear without folding in a direction parallel to the overthrust zone occurred.

Thus the more competent behaviour of Lochseiten mylonite at the interface with the "Flysch" sediments still remains problematic. Apparently, low viscosity in a simple shear configuration at the top does not automatically imply low viscosity in a buckling configuration at the mylonite/"Flysch" interface.

It must be noted that the "Flysch"/Lochseiten mylonite interface is often absolutely planar, especially in places where the total thickness of the mylonite does not exceed 1 metre. Beautifully developed and measurable internal fold structures in the Lochseiten mylonite are rather exceptional and chaotic, unmeasurable drag folds with axial planes which cannot be traced over more than a few mm and slight undulations of the lamination are by far more widespread. Folding is also absent in type 2 contacts, where a lamination parallel to the Verrucano base gradually develops in the uppermost 1–2 metres below it.

In spite of the partly confusing structural evidence, the following conclusions emerge:

1. Simple shear flow with rotation of already formed planar structures is confined to a zone approximately 1 metre thick below the Verrucano and neither in the

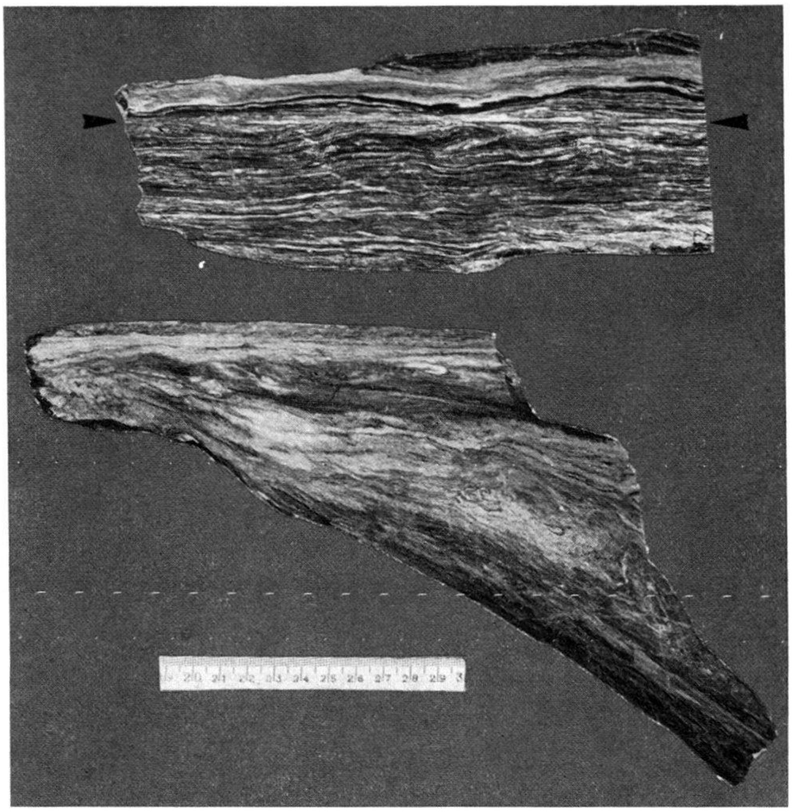


Fig. 8. Polished surface of two Lochseiten mylonite specimens (NNW to the left, cm-scale). Top: representative specimen of Lochseiten mylonite found in "type 1" and "2" contacts (from Foostock). The top surface to the right represents the interface with the base of Verrucano. Note the thin planar septum (arrows) cutting through the undulating lamination and filled with redeposited cacite.

Bottom: Lochseiten mylonite from a "type 3" contact (Ringelspitz). Note folding of the lamination at the base and asymptotic turning of the axial planes into an attitude subparallel to the Verrucano–Lochseiten mylonite contact, which is located a few centimetres above the upper boundary of the specimen.

Flysch structures nor often even in the basal zone of the Lochseiten mylonite are indications of such a simple shear flow field to be found. It must be emphasized that "simple shear flow" here just means flow with a strong simple shear component. Since no elongation parallel to the movement direction is observed, significant flattening across the lamination could also occur (see p. 267).

2. The planar septum inside the Lochseiten mylonite cannot account for the minimal horizontal displacement of at least 35 km nor for significant final displacements. Probably this cut represents the very last increments of displacement when conditions became unfavourable for flow and overthrusting stopped. Therefore the assumption that the total overthrusting of the upper over the lower block took place by flow in an intervening calc-mylonite layer approximately 1 metre thick (Hsü 1969*a*, his major phase of thrusting) is correct.

3. Not only the parautochthonous limestones which would have to be smeared out for several kilometres underneath the advancing thrust block can be taken as a possible source for mylonite formation, but also Subhelvetic limestone occurrences

emplaced either by earlier thrusting or at the front of the advancing nappe. But still, the presence of limestone mylonite over such a wide area, in most places with a thickness of 1–2 m, strongly suggests high mobility of this limestone under the prevailing conditions.

b) Lithology of Lochseiten calc-mylonite

The term "Lochseitenkalk" is normally used as a general designation for the band of limestones separating the Verrucano and "Flysch" units along the main Glarus overthrust and includes unmylonitized limestones in type 2 contacts. The term "Lochseiten calc-mylonite" used here is restricted to coherent tectonites with a characteristic lamination usually a few millimetres thick produced by dark finegrained and light coarser grained carbonate material (Fig. 8, 9a). Such mylonites however are not restricted to the main Glarus thrust; they are also found along other thrust planes in the eastern Helvetic nappes, although only locally (for instance, at the base of Griesstock nappe and along the basal thrust of Mürtschen and Axen nappes). Incoherent tectonic breccias and kakirites were not found along the main Glarus thrust in the area investigated, except for some few outcrops along the eastern side of Linth valley between Schwanden and Linthal (for the western side of Linth valley see SCHINDLER 1959).

Individual, slightly undulating or folded laminae cannot be traced over longer distances, they often are cut at low angles by other sets of laminae or they are offset by microfractures or stylolites. The appearance of this lamination is probably best described by the term "Knetstruktur" (kneading structure) used by HEIM (1922, p. 388). In southern areas, along type 3 contacts, the mylonite lamination is generally less pronounced, and more homogeneous, very fine grained white or yellowish limestones similar to very fine grained marbles occur.

By some authors (notably HEIM 1878, 1921) an extremely reduced inverted stratigraphic succession was postulated inside the band of limestone following the main Glarus thrust and the mylonite was believed to represent the smeared out inverted limb of a large fold nappe. Evidence for this was given by the presence of a thin yellow band of carbonates, interpreted as being derived from Triassic Röti dolomite, between Verrucano and the normally grey coloured mylonite, interpreted as derived from Jurassic limestone. This yellow colour, however, is not due to a dolomitic composition but to small contents of impure quartz. Around 50 X-ray diffraction analyses were made on mylonite samples: dolomite was found in only a few samples (especially from the Pizol area) and then always less than 25 percent, and many yellowish samples contain no dolomite at all, but traces of albite and quartz.

In the few places where trasitions between well preserved and determinable source material into mylonite can be studied, the unmylonitized limestone was found to be either a fine grained upper Jurassic limestone or a lower Cretaceous arenitic limestone with a coarse grained sparitic matrix (Öhrlikalk and Schrattenkalk formations). Dolomitic, shaly or quartz rich formations were never found mylonitized to give a Lochseiten-type rock.

Unfortunately, systematic petrofabric analysis of this mylonite with X-ray and electron microscope methods is still outstanding so that only microscopic observations are available (see Fig. 9):

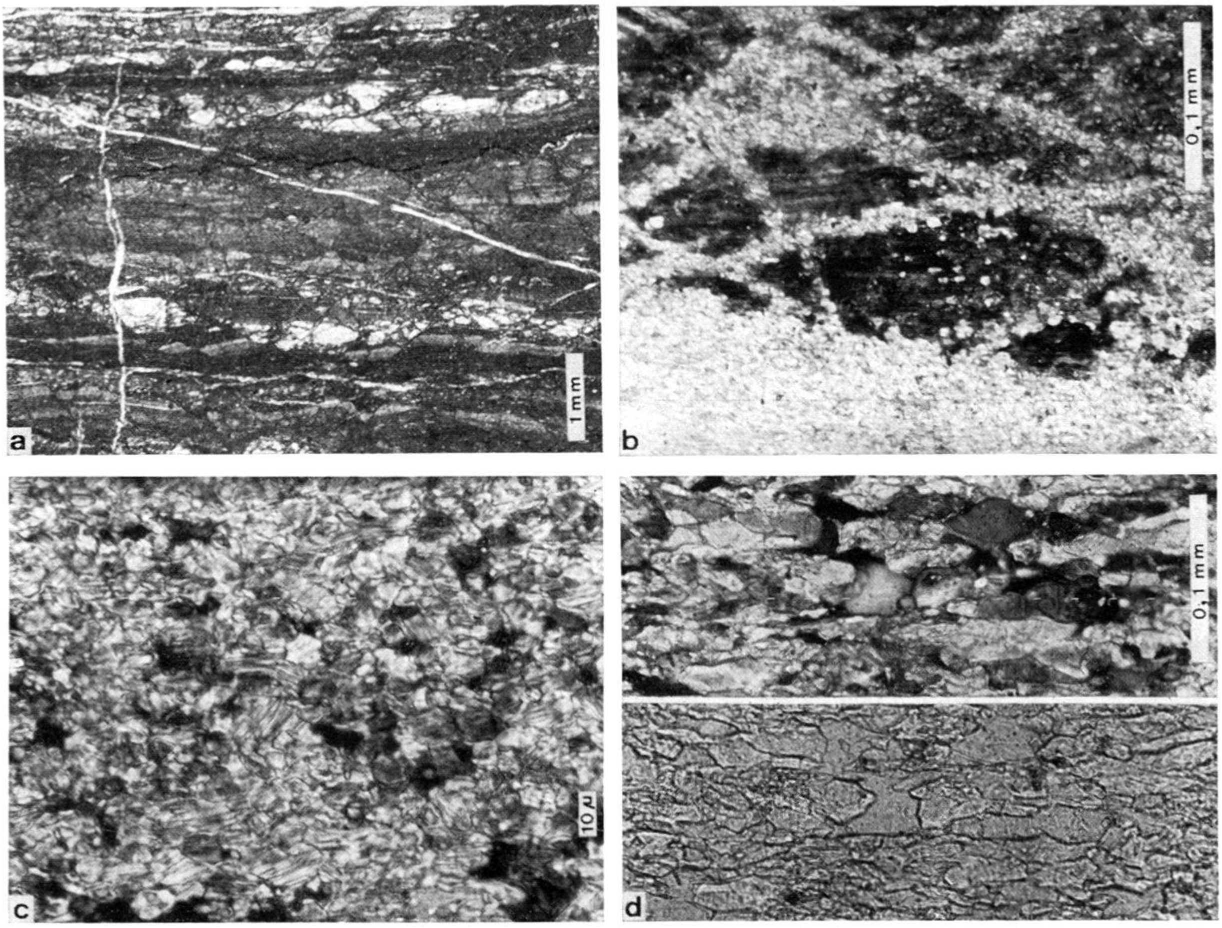


Fig. 9. Thin sections, Lochseiten mylonite.
a: Thin section of Lochseiten mylonite (same specimen as Fig. 8, top). Light laminae with coarser grained redeposited calcite aggregates, affected by subsequent shearing, and granulation. Dark laminae with very fine grained calcite aggregates (see Fig. 9c). Normal light.
b: Intraclast of a calcite single crystal as a relict of the sparitic matrix of a Cretaceous limestone, with a mortar texture developing around it and along distinct planes (transition into Lochseiten mylonite). Similar pictures are seen in the light laminae of already formed mylonite (Fig. 9a). Crossed nicols.
c: Fine grained calcite fabric as found in the dark laminae of Lochseiten mylonite. Note heavy twinning and deformation of twin lamellae. Crossed nicols.
d: Granoblastic texture found in marble-like Lochseiten mylonite in southern areas. Isolated quartz grains with undulatory extinction are recognized. Top: crossed nicols; base: normal light.

The sparitic matrix of the Cretaceous limestone first becomes strained, with heavily bent twin lamellae indicating translation along other intracrystalline glide planes. Along grain boundaries or inside the grains a mortar texture develops (Fig. 9b). In an intermediate stage, lensoid domains of single crystals, elongated parallel to a developing lamination, are surrounded by this very fine grained material (one or a few μm). In the mylonite as such no relicts of the old fabric (such as ooids or intraclasts) are seen any more but similar transitions from coarser grained calcite crystals redeposited along veins 0.05–1 mm wide are seen. Those veins are oriented preferentially parallel to the foliation. Sometimes very fine grained calcite aggregates with a strong preferred crystallographic orientation look like single crystals under low magnification. Stylolites oriented always parallel to the foliation in dark laminae indicate pressure solution. The age relationships between stylolites and calcite veins are so complex (sometimes pressure solution and redeposition occurs along the same plane) that the two processes must have occurred together.

Looking at a single thin section, one sees all those processes – granulation, pressure solution and redeposition in coarse grained veins which subsequently are granulated again – together in different domains. Microfractures usually indicating stretching in all directions parallel to the foliation are also frequent, indicating flattening perpendicular to the foliation, as also the stylolites. As in many quartzitic mylonites, no stretching lineation is developed. This additional flattening and/or continuous transformation of the fabric could be an explanation for this.

The marble-like tectonites in the southern areas are equally fine grained (around 0.01) mm, with a granoblastic tecture similar to many marbles (Fig. 9d). The grains are slightly flattened parallel to the lamination and show no first-sight indications of a preferred orientation. Stylolites are rarely developed and calcite veins are absent. Possibly some recrystallization under somewhat higher temperatures occurred here.

Isolated post-tectonic porphyroblasts of fresh albite (grain size up to 0.05 mm) occur frequently in the uppermost few cm of the mylonite. The only other minerals found were quartz and dolomite, apart from clay minerals and other impurities concentrated along stylolites. Quartz is often strained and finegrained, polycrystalline aggregates occur as components of a microbreccia or as intraclasts in a calcitic matrix near the base of Verrucano. In other cases definitely post-tectonic silification is observed. Dolomite generally occurs as coarser grained single crystals or aggregates, often clearly of post-tectonic origin.

In order to estimate the temperature of formation of the mylonite, the magnesium content of calcite in the presence of dolomite was determined using X-ray diffraction analysis. The method used is described in THEODORE (1970) and based on experimental work by GOLDSMITH and others (for references see THEODORE 1970). A maximum content of 2.4 mole percent $MgCO_3$ was found in 15 different specimens, giving a temperature of around 390 °C. Since the magnesium content is lowered by post-metamorphic exsolution, no geographic variation can be deduced. It is nevertheless interesting that the highest magnesium contents were found in southern and eastern areas. The accuracy of this temperature determination is probably not too good, but the figure obtained compares well with the conclusion of FREY et al. (1973), that temperatures of 300 °–400 °C were required for the formation of biotite prior to overthrusting in the parautochthonous units.

c) Topography of the overthrust plane and the question of post-overthrust movements

The topography of the main Glarus thrust is quite simple and smooth, with a WSW–ENE trending culmination and an depression around Segnespass (see Fig. 10). The southern flank has an average slope around 15 °, but further south and near the Rhine the slope must steepen to at least 20 ° in order to acommodate the overthrust plane and the parautochthonous cover of Aar massif underneath the Pennine Bündnerschiefer. The northern flank dips 10 °–15 °. Except for the areas north of Engi, south of Segnespass and north of Rotstock, no significant second-order topographical features were detected by contouring. In the first two cases, the contours have been smoothly bent although in the area north of Engi dextral offset along a post-overthrust fault, whose geometry is not well enough known for an exact construction, is more likely, joining the well-known Murgsee fracture further north (see TRÜMPY 1969, Fig. 2). South of Segnespass such an offset along a simple strike slip fault can be excluded because no signs of such a discontinuity are found in the units below the overthrust (FELDER 1973). North of Rotstock a secondary depression is developed whose significance is unknown.

It is surprising how few traces of post-overthrust deformation are found even on a small scale. The only signs of clearly post-overthrust movements, apart from

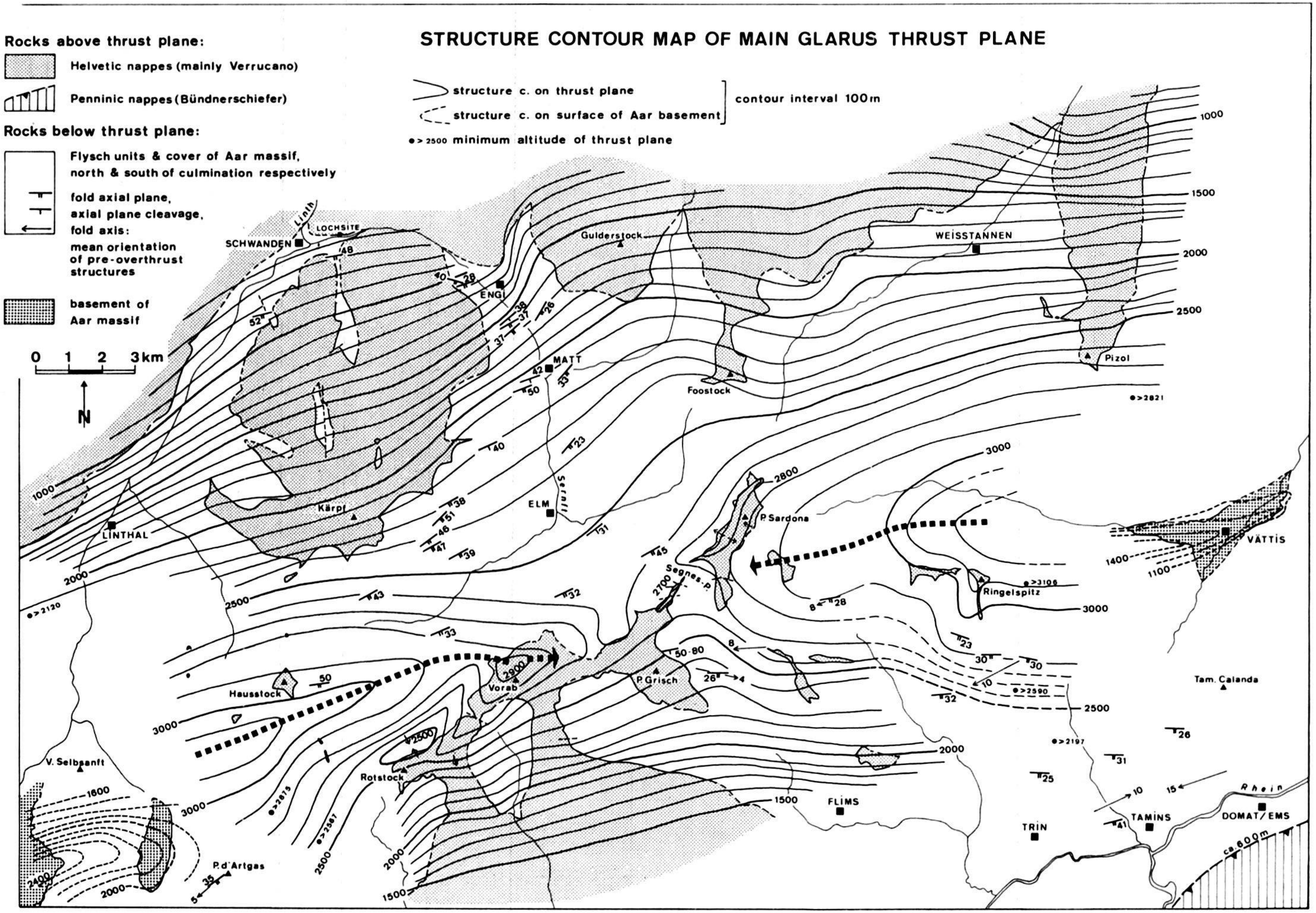


Fig. 10. Structure contour map of main Glarus thrust plane (largely based on FISCH 1961, HELBLING et al. 1948, OBERHOLZER 1920, 1942 and SCHIELLY 1964).

the strike-slip fault north of Engi, are conjugate sets of normal faults beautifully developed around Segnespass (faults bisecting at a 60° angle and striking E–W) and some faulting in the area between the Sernft and Linth valleys (SCHIELLY 1964). Only the S-dipping and better developed set, however, shows detectable displacements of a few metres at a maximum. Further south from Segnespass, this is the only set developed, with increasing downthrow of the southern block (see WYSSLING 1950, plate 3). This faulting may be partly responsible for an overall steepening of the overthrust plane towards the Rhine valley. PFIFFNER (1972*b*), however, observed also steepening of cleavage and fold axial planes in the parautochthonous cover of Aar massif west of Chur, not related to late faulting. Along the Rhine valley, a major late Alpine fracture zone (Rhone–Rhine fault zone, TRÜMPY 1973), with a component of vertical downthrow of the Pennine block to the south runs parallel to the "root zone" of the Helvetic nappes along Tavetsch massif, and probably causes a steepening and fracturing of the southern flank of the overthrust plane. It is interesting to note that the same late Alpine discontinuity, turning north around Chur (Fig. 1), leaves no traces in the easternmost areas of our contour map in that no eastward plunge of the culmination or the overthrust plane as a whole is seen.

Although post-overthrust movements concentrated along this fault zone are important, the present shape of the main Glarus overthrust plane is interpreted to be of primary origin. Bending of the overthrust plane by differential uplift of the Aar massif after the block was emplaced (ARBENZ 1934, OBERHOLZER 1933, TRÜMPY 1969, 1973) is unlikely for the following reasons:

a) Bending by differential uplift implies stretching of the overthrust plane by about 2.5 percent, assuming an originally horizontal attitude. This strain would have to be recorded by conjugate normal faulting, since ductile extension is extremely unlikely and vertical tension fractures are not developed. Such widespread systematic faulting is not observed: only the southern flank is additionally steepened by such offsets, as already mentioned, and along the northern slope normal faulting is not observed at all.

b) In Figure 2 it is seen that the "Flysch" units, which are of an enormous thickness north of Aar massif, completely thin out southwards between the autochthonous cover of Aar massif and the curved overthrust plane. This suggests a pre-overthrust culmination of Aar massif (basement surface also has been contoured in Figure 10).

c) If the culmination is of post-overthrust age and the thrust plane is brought into a subhorizontal position by upwards rotation of the two flanks by 10°–20°, all the other planar structures would have to be rotated as well. This would result in a primary dip of phase 2 fold axial planes in the "Flysch" around 50°–60° and of 0°–20° in the parautochthonous folds (see representative measurements in Fig. 10). Such a discrepance in the angle of dip with a sudden changeover near the future culmination is unlikely.

d) Last but not least, a simple argument for a primary S-dipping flank of the overthrust plane lies in the fact that Permian Verrucano has to be brought along a S-dipping low angle fault on top of the Tertiary somewhere.

It has to be emphasized that only differential uplift of the Aar massif in the present area is rejected. Evidence for post-overthrust uplift of the Helvetic block as a whole is quite strong and tilting cannot be excluded. Also it is quite possible that differential uplift after emplacement of the Helvetic nappes is responsible for the axial culmination of the massifs in Central Switzerland. Such a differential uplift could go together with folding of the basal overthrust of the Helvetic nappes as observed only west of Reuss valley (Spörli 1966, Trümpy 1973).

6. Discussion of field evidence and timing of tectonic events

Tectonic phases were numbered only in the "Flysch" units where geometrical and age relationships clearly suggest three separate tectonic events of which only the latest is related to movement along the main Glarus thrust. The phase 1 event is younger than lower Oligocene, the time of sedimentation for the youngest Flysch sediments involved (Engi slates). Stratigraphical evidence dating the end of overthrusting is less precise. Trümpy (1969, 1973) places the main phase of thrusting (our phase 3) somewhere in the Miocene ($\pm$ Burdigalian), based on the fact that the first pebbles of Helvetic origin appear in Molasse of Helvetian age (Leupold, Tanner & Speck 1942).

Absolute dating of low grade metamorphism in the eastern Helvetic nappes is still a difficult task. Frey et al. (1973) came to the conclusion that the peak of this event has to be placed between 31 and 36 m.y. (lower to middle Oligocene), post-dating what they called "Helvetische Hauptphase". Since then further work was done, and M. Frey (pers. comm.) now suggests a younger age around 20–25 m.y. for the peak of metamorphism, definitely pre-dating final movements along the main Glarus thrust but still post-dating some internal nappe boundaries inside the Helvetic thrust block. Post-tectonic growth of stilpnomelane in the parautochthonous cover of Aar massif has been described by Bürgisser & Felder (1974).

All this suggests a final phase of overthrusting post-dating a period of ductile, penetrative deformation and of low grade metamorphism restricted to the south. So the following interpretation is given (table 1):

Phase 1 does not seem to have a direct relation to later tectonic activity and probably represents gravitational movements induced by thrusting of more internal (Penninic and Austroalpine) units (Trümpy 1969), probably along a pre-existing slope inclined towards the foreland.

Phase 2 in the "Flysch" units is assumed to be contemporaneous with ductile and penetrative deformation in the other tectonic units, mainly because of the similarity in the deformational style. Because of the high ductility it is likely that low grade metamorphism was partly contemporaneous (and partly younger).

Phase 3 events can probably only be separated clearly from phase 2 events in the "Flysch" units. Ductile deformation could represent the initial phase of overthrusting for the following reasons:

a) The zonation of index minerals (Frey et al. 1973, Fig. 3) strongly suggests that the sediments of the Helvetic thrust block had already escaped their original position south of Aar massif when metamorphism occurred. The highest metamorphic grade is found in the latter and a metamorphic grade increasing towards the foreland seems unlikely.

Table 1. Timing of tectonic events in the Eastern Helvetic nappes (absolute timing very hypothetical).

my	Epoch	Phase	«Flysch» units	parautochthonous cover of Aar massif	Helvetic thrust block	Metamorphism
0–10	Pliocene / Miocene				post-overthrust movements, of minor importance; mainly along Rhone-Rhine fault	
10–20	Miocene	phase 3	syn-overthrust deformation only along thrust contact	locally, syn o. thrust deformation along thrust contact	displacement of the Helvetic thrust block along the main Glarus thrust; secondary thrusting in frontal and higher parts of the thrust block associated with folding	
20–	Miocene			intermediate thrusting?	thrusting of Subhelvetic units	low grade metamorphism in par. c. of Aar m. & party in Helv. th. block
–30	Miocene / Oligocene	phase 2	ductile, penetrative phase of folding	ductile, penetrative phase of folding & subsequent thrusting	folding with cleavage formation and N-S stretching in Verrucano; minor (?) thrusting	
30–	Oligocene	phase 1	diverticulation, gravity sliding?			
–40	Oligocene		continuing sedimentation in North Helvetic Flysch basin			

b) Ductile folding with SE to S dipping axial and cleavage planes affected all the three main tectonic units listed in table 1. Geometrically the result of this folding together with associated thrusting could well be a piling up of sedimentary units with the Helvetic sediments, of southermost origin, on top. So this deformation would overcome, at least partly, the relative vertical uplift required to place the Verrucano sediments on top of Mesozoic limestones. Only after this, could movement by flow of Lochseiten mylonite occur. In this sense, ductile deformation together with minor thrusting really represents the preparatory phase of overthrusting.

Final overthrusting along a curved thrust plane probably initiated subparallel to pre-existing cleavage planes in the south, as suggested by the subparallel position of the thrust plane with cleavage planes inside Verrucano and parautochthonous units. These planar anisotropies must have played an important role in the initiation of the thrust because no suitable décollement horizons between Permian Verrucano and pre-Permian basement were present.

During downwards movement of the frontal part of the thrust block gravity-induced secondary thrusting probably occurred, leading to the final individualization of the different Helvetic nappe units inside the thrust block. Although this secondary thrusting does not represent an imbrication zone related with the main Glarus thrust directly it shows that the stability of the thrust block was exceeded at least locally.

Because the horizontal displacement of 35 km (measured from the Verrucano front backwards to the Rhine valley) represents just a minimal amount of total displacement, this figure is still realistic even after subtracting displacements during the initial phase, before final thrusting started. Although the duration of overthrusting is unknown, a figure around 10 m.y. is probably a realistic upper limit. This results in a minimum shear strain rate around $\dot{\gamma} = 1 \times 10^{-10}\,\text{sec}^{-1}$ (assuming 35 km displacement and a flowing mylonite layer 1 m thick) and a displacement rate of 3.5 mm/y. The temperature estimate of 400 °C inside the mylonite represents, probably, an upper limit, the thickness of the block is estimated to be 6 km at least.

Post-overthrust tectonism did not reactivate the main Glarus thrust and did not change its shape appreciably: it occurred along fractures such as the Rhone–Rhine fault zone and the Murgsee strike-slip fault.

D. Discussion of a mechanical model

1. Limits of the Mohr–Coulomb criterion of rock failure as a basis for a mechanical model for overthrusting

There are two mechanical problems associated with overthrusting: a) the initiation of overthrusts and b) the problems of stability in a moving and already detached overthrust block along an already formed plane of weakness.

As a criterion for rock failure initiating a thrust fault, generally the Mohr–Coulomb criterion is used (ANDERSON 1951, HAFNER 1951, MÜLLER 1974). In our example the field evidence suggests that overthrusting is essentially the result of crustal shortening but that a period of ductile folding preceded and probably even initiated thrusting. Therefore, in our case, this criterion for brittle fracture is an inadequate basis for the analysis of overthrust initiation. Because there is no adequate basis for treating ductile folding analytically, main emphasis is laid here on the problem how this thrust block could possibly move when already emplaced on top of the footwall, in other words, on top of Mesozoic limestones able to form Lochseiten-type calc-mylonite.

The classical paper by HUBBERT & RUBEY (1959) gives a mechanical analysis especially applicable to already detached overthrusts, based on a modified Mohr–Coulomb criterion (omission of the "cohesion" term τ_0 and account for the effect of pore pressure). Their model has been repeatedly modified, for instance, by HSÜ (1969*b*), who argues that τ_0 cannot be neglected for calculating the shear resistance at the base of an overthrust on the basis of the Mohr–Coulomb criterion in the form

$$\tau_c = \tau_0 + (1 - \lambda)\, \sigma_n \tan \varphi \tag{1}$$

where τ_c is the critical shear stress necessary to initiate rock failure, σ_n the normal stress on the fault surface, $\tan \varphi$ the angle of internal friction, and $(1 - \lambda)$ a term reducing the normal stress by the action of pore pressure, λ being the ratio of fluid pressure to load pressure.

One of the main geological arguments for not neglecting τ_0 was given by the correct observation that the Glarus overthrust did not move along a cohesionless thrust plane but "by flow of a thin layer of pseudoviscous material between the plates" (HSÜ 1969*b*, p. 941), whereby resistance to flow is given by equation (1).

This criterion has in fact been verified experimentally at low temperatures and relatively low confining pressures for describing the strength of rocks also in the ductile field. However, it is known that the value of tan φ decreases and approaches zero with increasing temperatures and/or confining pressures (e.g. HANDIN 1969). This means that the yield strength for plastic flow is no longer governed by the absolute value of the least principal stress but by the stress difference $\Delta\sigma = \sigma_1 - \sigma_3$. This leads to the well known Tresca criterion, which states that flow occurs when the stress difference exceeds a critical value.

In addition flow of rocks under elevated temperatures is also known to be strongly strain rate dependent (for calcite rocks see HEARD 1963, HEARD & RALEIGH 1972, RUTTER 1972, 1974, RUTTER & SCHMID 1975). In this case, the material is said to behave pseudoviscously and its flow strength, still defined by the stress difference, is related to strain rate. Therefore, the Mohr–Coulomb criterion, also when it is used as a criterion for plastic yield, is a fundamentally different concept, even abstracting from the strain rate dependence not being described.

Hence one has to make a choice between a yield criterion for pseudoviscous flow or the Mohr–Coulomb criterion. Here, the first is used since field evidence suggests continuous flow under elevated temperatures without strain hardening. Strain hardening would lead to brittle failure inside the mylonite layer after a relatively small strain increment. The Mohr–Coulomb criterion will be used only to estimate the stability of the overthrust block.

2. Shear resistance governed by flow of a pseudoviscous layer

The constitutive law for thermally activated plastic flow of materials under constant stress can be written as

$$\dot{e} = A \exp(C\Delta\sigma) \tag{2}$$

where $\dot{e}$ is the strain rate at differential stress $\Delta\sigma$ and A is a constant incorporating the temperature effect on the viscous behaviour and various material constants. C is a constant describing the stress dependence of strain rate. This flow law has been successfully applied to calcite rocks in triaxial tests in the temperature range between 300° and 600°C at low strain rates (for details see HEARD & RALEIGH 1972, RUTTER & SCHMID 1975). RUTTER (1974), comparing the experimental results for Solnhofen limestone with those for Carrara marble and Yule marble worked out that the rheological behaviour of such calcite rocks is similar in that a slope of $C/\ln(10) = 3.5\,\mathrm{kb}^{-1}\log_{10}\mathrm{sec}^{-1}$ gives a basis for extrapolation to geological strain rates, plotting the strength of calcite rocks on a stress/log strain rate diagram (Fig. 11) above differential stresses around 800 bars, where an exponential description of stress dependance holds (equation (2)). Additionally the same author found that the presence of pressurized interstitial water does not greatly affect the rheological behaviour at higher temperatures.

If the experimental results from these rocks are applied to calculate shear resistance to flow in Lochseiten mylonite, and hence resistance to overthrusting as a function of strain rate, the following assumptions have to be made:

1. The value of the constant A in equation (2) is not yet known for Lochseiten mylonite. HSÜ (1969*a*) had to use the only data available at that time, which were HEARD'S (1963) data on Yule marble. Here, the data on Solnhofen limestone are used because the structure of this rock is more similar to the structure of Lochseiten

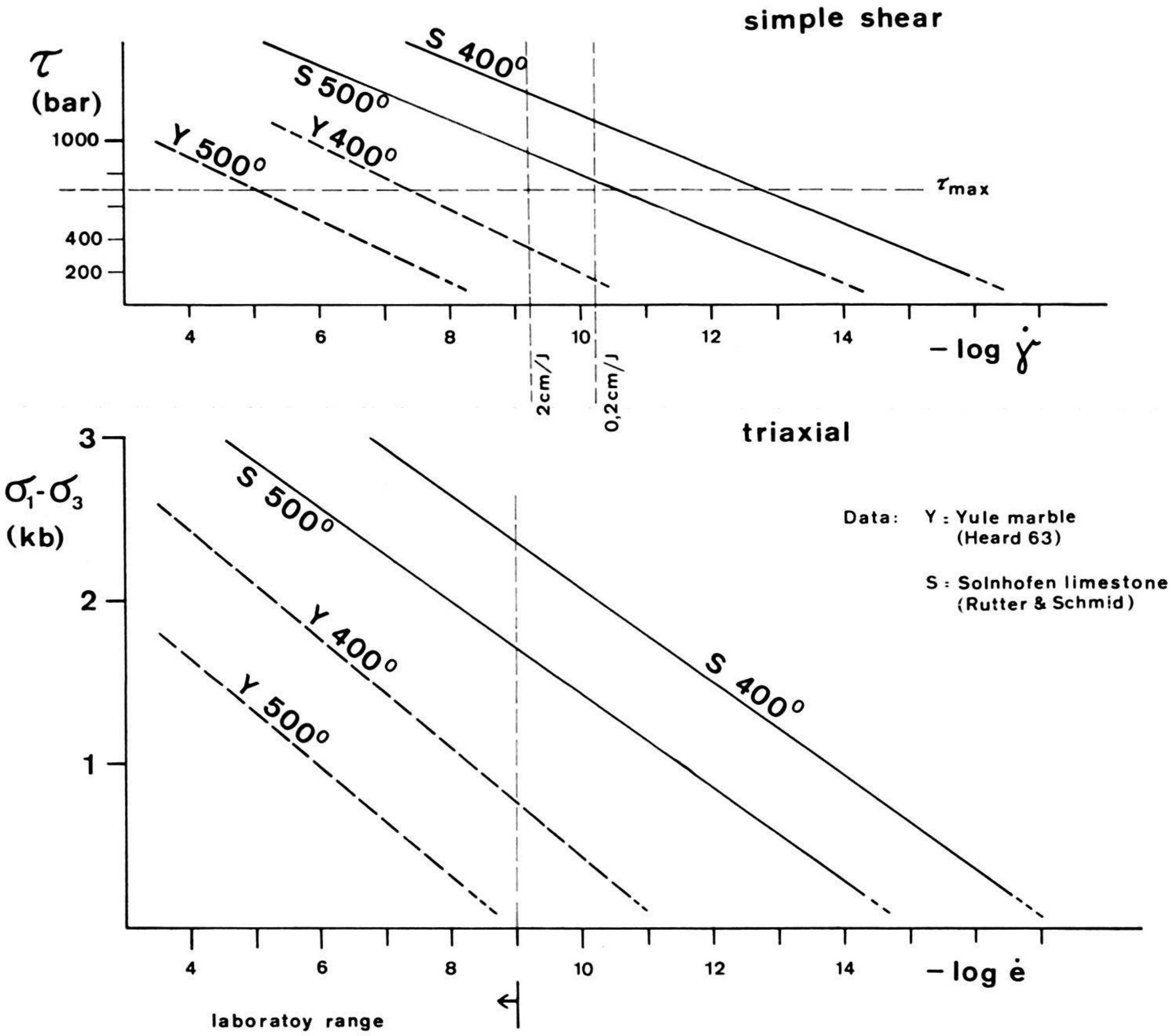


Fig. 11. Stress/strain rate diagrams showing linear isotherms on a semi-log scale fitted to data points obtained by triaxial laboratory tests (for details see HEARD 1963 and RUTTER & SCHMID 1975). For extrapolation into natural strain rate ranges a slope of $C/\ln(10)$ = 3.5 and 3.0 kb^{-1} $\log_{10}$ sec^{-1} was used for Solnhofen limestone and Yule marble respectively (see equation 2). In the upper diagram, the triaxial data have been converted into a simple shear configuration. For horizontal and vertical dashed lines see text.

mylonite, especially with respect to the grain size (10–20 μm in Solnhofen limestone). Additionally the room temperature strength of both rocks is similar (Fig. 12, Table 2) and higher than the strength of marbles which tend to be significantly weaker also at high temperatures.

2. Deformation mechanisms must be assumed to be the same in both laboratory experiments and in nature, since deformation mechanisms really determine the rheological behaviour. Thermally activated flow involves intracrystalline processes, whereas it is possible from the microscopic observations that cataclastic flow was important in the Lochseiten mylonite at least towards the end of overthrusting. Unfortunately, the observable textures probably reflect only the last strain increments in this mylonite and tell us little about the enormous amount of deformation which preceeded them.

In a next step axial shortening or extension performed in the laboratory has to be converted into the simple shear configuration assumed here. Such a conversion was done following NYE (1953) who did the same on the flow law of ice governed by a similar pseudoviscous relationship (power law stress dependence of strain rate). Thereby the flow law for axial shortening or extension (equation 2) is generalized and stress dependence is given in terms of the octahedral shear stress. Then the special case of simple shear flow is rederived. This is an alternative method of calculating the "equivalent viscosity" used by GRIGGS (1939). This conversion then leads to the expression

$$\dot{\gamma}_{xy} = A\sqrt{3}\exp(C\sqrt{3}\,\tau_{xy}) \qquad (3)$$

where $\dot{\gamma}_{xy}$ is the shear strain rate and τ_{xy} the corresponding shear stress. A and C are the experimentally determined constants at a given temperature in equation (2).

This conversion is graphically illustrated in Figure 11 for the 400°C and 500°C isotherms of Solnhofen limestone (based on experimental results in RUTTER & SCHMID 1975) and Yule marble (based on HEARD 1963, Fig. 19).

3. Contradictory approaches towards an estimate of shear resistance at the base of the overthrust block

It can be seen from Figure 11 that the difference in strength of the two rocks at a given temperature and strain rate is enormous. Under the assumption that the 400°C isotherm of Solnhofen limestone is valid for our model, the strain rate estimated from field evidence can be related to shear stress. The vertical dashed lines in Figure 11 indicate strain rates corresponding to overthrust rates of 0.2 cm/year (somewhat below our estimate) and 2 cm/year (HSÜ's, 1969*a*, figure using the 400°C isotherm of Yule marble at a shear stress of 360 bars[2])). This strain rate interval, the strain rates being remarkably high and well above the figure of $\dot{\gamma} = 10^{-14}\,\mathrm{sec}^{-1}$ generally given for natural environments (e.g. HEARD & RALEIGH 1972), gives stresses between 1.1 and 1.3 kb. This stress figure is very high indeed and alternative estimates of the maximal shear stress are carried out in the following:

a) Maximal shear stress allowing for the stability of the overthrust block

The stability of the overthrust block can be assumed to be maintained during overthrusting although locally this assumption is not justified. HSÜ (1969*a*) made the same approach using the equation of LAUBSCHER (1961, p. 244) allowing for the effects of pore pressure:

$$\tau_{max} = \frac{1}{x}\left\{a\,z + [b + (1-b)\,\lambda]\,\frac{\varrho_b\,g\,z^2}{2}\right\} \qquad (4)$$

where x is the length of the overthrust block (35 km), z the thickness (6 km), ϱ_b the density of water-saturated sedimentary rocks (2.3 g cm^{-1}), g the gravitational acceleration (980 cm sec^{-2}), a and b constants in the Mohr–Coulomb equation, whereby $a = 2\tau_0\sqrt{b}$ and $b = (1+\sin\varphi\,/\,1-\sin\varphi)$. However, instead of using the "standard"

[2]) HSÜ (1969*a*) obtained a figure of 360 bars from applying the Mohr–Coulomb criterion as a yield criterion.

constants for a and b suggested by HUBBERT & RUBEY (1959), and also by many others, we can use the contants experimentally derived by room temperature testing of rocks from the Helvetic nappes (unpublished data from BRIEGEL & SCHMID, see table 2).

Table 2. Approximate value of the constants a and b for a straight line approximation of the Mohr–Coulomb criterion in the form $\sigma_1 = a + b\sigma_3$:

	a (bars)	b	brittle–ductile transition (bars)
Lochseiten mylonite	2800	3.5	1000–1300*)
Verrucano	2700	4.0	above 2000
Öhrlikalk	3000	2.4	1000–1200
Solnhofen limestone**)	2900	2.2	800–1000
"standard" values	700	3.0	

*) ductile for less than 10 percent only, with subsequent fracture.
**) data from RUTTER (unpublished).

Using constants a and b for Öhrlikalk (Verrucano has a still higher strength) in equation (4) this leads to the following values, considerably higher than the values obtained using the "standard" constants:

Table 3. Values of τ_{max} derived by equation (4) at various pore pressure to load pressure ratios:

	Öhrlikalk	"standard" values
$\lambda = 0$	780 bar	470 bar
$\lambda = 0.5$	700 bar	360 bar
$\lambda = 1.0$	630 bar	240 bar

The figure around 700 bars is possibly too high for the following reasons: 1. Equation (4) is valid only for horizontally moving overthrusts along a flat thrust surface and 2. the assumption behind this equation, that σ_1 acts parallel to the thrust plane, has been shown to be invalid by FORRISTAL (1972). On the other hand, failure really occurred (syn-overthrust deformation in the Verrucano), indicating that the block did not necessarily remain stable everywhere, and that a figure around 700 bars still might be realistic.

b) The Mohr–Coulomb criterion applied to Lochseiten mylonite giving a figure for the maximal shear stress

This approach results from the argument that high temperature flow had to successfully compete against Mohr–Coulomb brittle fracture in the Lochseiten mylonite. Therefore the strength of this rock derived at room temperature and described by the Mohr–Coulomb criterion gives a maximum figure for the shear stress as well.

Inserting the constants obtained from experimental deformation of Lochseiten mylonite (Fig. 12) in equation (1), whereby σ_n is taken to be equal to ($\varrho_b\, g\, z$), gives the following figures:

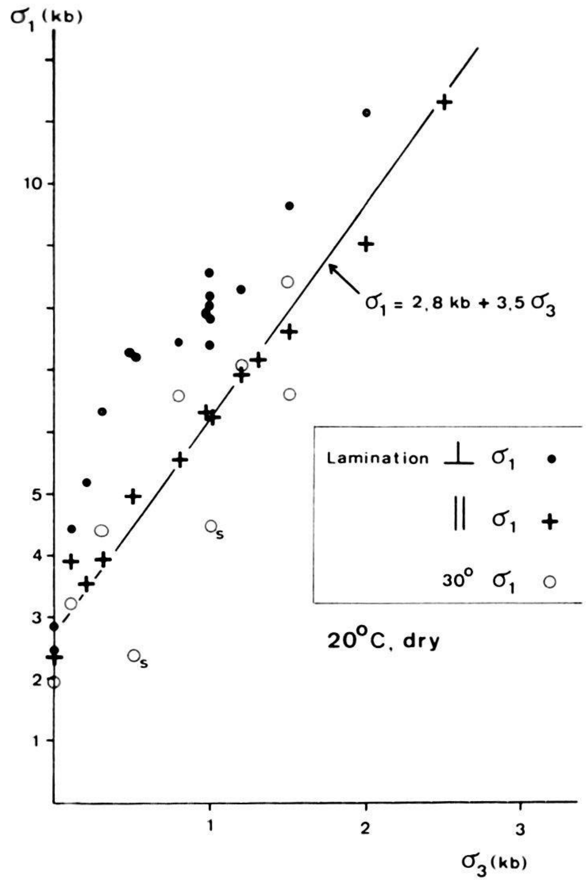


Fig. 12. The room temperature strength of Lochseiten mylonite cylinders on a σ_1/σ_3 diagram (σ_1 = axial plus confining pressure, σ_3 = confining pressure). Points labelled by "s" show the strength of specimens with fracture along a stylolite plane. The straight line represents the visual best fit to points obtained with the lamination parallel to the cylinder axis (see table 2). Unpublished data: U. Briegel and S. Schmid.

$$\tau_{max} = 1640 \text{ bar, for } \lambda = 0$$
$$\tau_{max} = 1195 \text{ bar, for } \lambda = 0.5$$
$$\tau_{max} = 750 \text{ bar, for } \lambda = 1.0$$

From those approaches a maximum shear stress is given by 700 to 750 bars assuming λ to be 0.5 in the overthrust block and 1.0 inside the Lochseiten mylonite layer (horizontal dashed line in Fig. 11). The results plotted in Figure 12 are also important as an argument against a phase of later thrusting under brittle conditions inside the Lochseiten mylonite layer, since this rock is extremely strong at low temperatures.

The maximal possible shear stress at the base of the overthrust together with the minimum strain rate deduced from field evidence clearly indicates that the 400°C isotherm for Solnhofen limestone is far outside the field which is geologically realistic. This however does not mean that Yule marble is a better approximation for Lochseiten mylonite since the two rocks are too different in structure.

4. *Conclusions*

The conclusion from these simple calculations is that the assumptions made for using equation (2) in this natural environment were not fully justified. The resulting shear stresses at the base of the overthrust are too high. The following considerations might be helpful in formulating the problems to be solved by further investigations:

1. Assumption 1 (p. 273) has to be tested by directly deforming Lochseiten mylonite at high temperatures. Unfortunately this could not yet be done. It is possible however

that the viscosity relation established experimentally on Lochseiten mylonite will still lead to a figure too high for the basal shear stress, considering the enormous strength of this rock at room temperature.

2. The critical point is probably that the deformation mechanisms operating in Lochseiten mylonite were not the same as the mechanisms observed in the laboratory (RUTTER & SCHMID 1975). Microscopic observations suggest that intracrystalline deformation alone was not responsible for the bulk of the strain and that other mechanisms were at least partly operating. For determining the deformation mechanisms in Lochseiten mylonite, examination of samples by electron microscopy is needed, together with deformation experiments on that rock giving a geologically realistic viscosity relation now established for the Glarus overthrust.

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