Tertiary kinematics in the Lepontine dome

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SUMMARY: In the Pennine zone of the central (Lepontine) Alps, the mineral-zone boundaries of Tertiary metamorphism and the regional attitudes of foliation and thrust sheets define a dome. We investigate the Tertiary kinematic history of the dome using structural, metamorphic and geochronological data. Thus we distinguish two main ductile deformations, that occurred under differing metamorphic conditions. The first high-temperature deformation (HTD) developed under amphibolite facies conditions. It is responsible for (1) a radiating pattern of stretching lineation trends at the scale of the Lepontine area, and (2) structures indicating overthrusting towards the European foreland in the lineation direction (i.e. N and NW). The HTD event is attributable to NW–SE Tertiary convergence between Europe and the Adriatic sub-plate. The second retrograde deformation (RD) is responsible for differential uplift of the Pennine zone, with formation of an earlier subdome to the E (Ticino subdome, 25–20 Ma) and a later one to the W (Simplon subdome, 15–10 Ma).

In the Ticino subdome, kinematic indicators reveal a horizontal shearing top W. Some major structures such as the Maggia steep zone and the vertical duplication of the Antigorio basement nappe (Bosco–Gurin area) may be related to this westward phase. Dextral transcurrent shearing along the southern steep belt and backthrusting of the central Alps over the southern Alps along the N dipping Insubric mylonite belt are probably coeval with the retrograde deformation recorded in the Ticino subdome. In the Simplon subdome, kinematic indicators reveal a radial motion parallel to the lineation, which plunges towards the SSW in the Verampio window and towards the WSW at the Simplon Line.

Serial uplift of Ticino and Simplon subdomes is attributed to continuing dextral transpression at the boundary between Europe and the Adriatic sub-plate.

Finally, the back-folding event located at the northern limit of the Pennine zone postdates both HTD and RD deformational events described above. It is part of a later pop-up responsible for uplifting the external crystalline massifs.

Introduction

In this paper, we use structural, metamorphic and geochronological data to study the Tertiary kinematics of the Lepontine Alps. Many of the data we have taken from the abundant literature dealing with this area, but some structures (stretch lineation patterns and kinematic indicators) are presented here for the first time.

The Lepontine region appears as a single dome in terms of the mineral zone boundaries (isograds) of Tertiary metamorphism (Niggli 1970) and is usually referred to in this sense as the Lepontine dome (Fig. 1). The region is also a dome, albeit a composite one, on the basis of foliation and thrust sheet attitudes (Fig. 2). The Pennine zone within the dome shows ductile deformations of remarkable intensity and complexity, coeval with Tertiary metamorphism of amphibolite grade and the later cooling history. Thus the Lepontine dome provides an excellent and unusual opportunity for studying deep-seated processes of deformation (probably associated with plate convergence), followed by shallower processes (associated with uplift).

There are several reasons for studying the Lepontine dome. First, it occupies a critical position within the Pennine zone, which is often considered to represent an originally thinned continental margin (or margins) between the European plate proper to the N, and the southern Alps, belonging to an Adriatic sub-plate of African affinity, to the S (Fig. 1). In the Lepontine region, the current boundary between the Pennine zone and the southern Alps is the peri-Adriatic fault system, locally known as the Tonale, Insubric and Canavese Lines (see Schmid *et al.* this volume). Second, the eroded Lepontine dome provides a tectonic window through the Pennine zone. Third, the dome is also at a critical longitude, between the eastern Alps, roofed by Austro–Alpine thrust sheets, and the western Alps, which expose mostly Upper Pennine units (Monte-Rosa and Bernhard sheets, Fig. 1). Finally, in this general area, the peri-Adriatic fault system bends round from an EW attitude (Tonale line) to an almost N–S attitude (southern Canavese line). In such a context, it is perhaps not surprising that the internal structure of the Lepontine dome is complex. Nevertheless, we feel that an understanding of it should be helpful towards a

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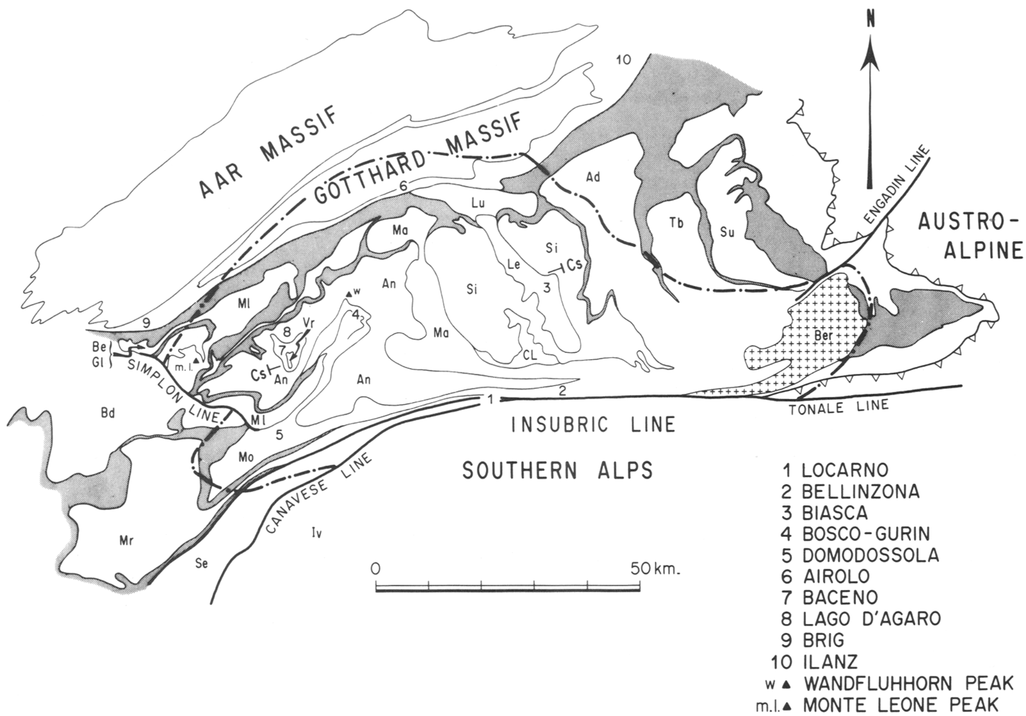


FIG. 1. Simplified geological map of the Lepontine dome. In the Pennine zone, cover rocks (stippled) are interleaved with slices of crystalline basement (Ad: Adula, An: Antigorio, Bd: Bernhard, Cl: Cima-lunga, Le: Leventina, Lu: Lucomagno, Ma: Maggia, Ml: Monte Leone, Mo: Moncucco, Mr: Monte Rosa, Se: Sesia, Si: Simano, Su: Suretta, Tb: Tambo,). Also shown are Verampio window (Vr), Berisal synform (Be), Glishorn antiform (Gl) and Bergell intrusion (Ber). Periadriatic faults (Canavese, Insubric and Tonale Lines) separate the Pennine zone from the southern Alps, which include the Ivrea zone (Iv). Thick dashed line encloses area of amphibolite facies metamorphism of Tertiary age. Cs indicates the line of the cross section shown in Fig. 9. For numbered localities and peaks see key at bottom right.

broader-scale understanding of Alpine structures and kinematics.

Since this contribution aims at regional kinematics, we primarily make use of the large scale patterns of foliations, lineations and shear indicators at amphibolite grade, or later greenschist grade. As a consequence, the reader familiar with the existing literature will find that we do little justice in this paper to the complexity of local deformation phases, especially the geometrical style, orientation data and interference patterns of folding on all scales (such as discussed in Huber *et al.* 1980). This shortcoming simply results from the fact that we are not yet able to fully integrate this kind of structural analysis into a regional kinematic history and that we are not aware of any previous attempts to do so.

We first describe the general structure of the Lepontine dome itself, including the thrust sheet sequence and the foliation pattern. Then we summarize what little is known of its pre-Tertiary history. In more detail, we discuss petrological and geochronological data on the Lepontine metamorphic event. This allows us to refer to metamorphic assemblages when describing linear and planar fabrics or kinematic indicators. Using the patterns of lineations and the associated senses of shear, we finally attempt a Tertiary kinematic interpretation of the Lepontine dome. The inferred motions are so thoroughly three-dimensional, that we are not yet able to use balanced cross-sections, or any other method, to calculate total amounts of displacement.

The structural dome

In the Lepontine region, a geological map of the Pennine zone (Fig. 1) looks much like a sliced onion, for several geometrical reasons.

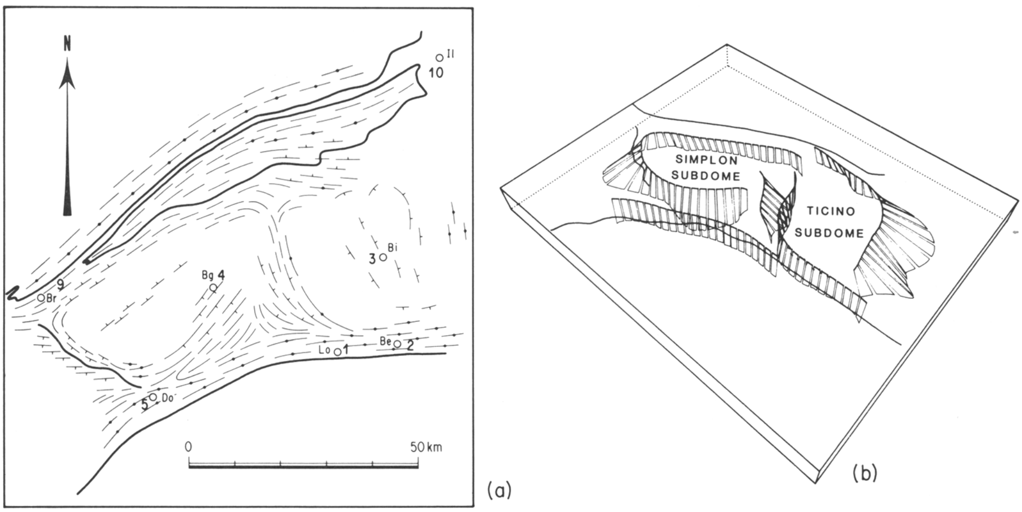


FIG. 2. Current attitude of main foliation in western and central parts of the Lepontine dome. Map at left shows smoothed strike and dip of foliation. Steep dips of 60° to 90° are shown by black dot centred on strikeline; moderate dips of 30° to 60°, by tick on downdip side of strikeline; flattish dips of O° to 30°, by lack of symbol. Numbered localities are as in Fig. 1: Biasca (Bi), Bellinzona (Be), Locarno (Lo), Bosco–Gurin (Bg), Domodossola (Do), Brig (Br) and Ilanz (Il). Data from Schmidt & Preiswek (1908), Knup (1958), Milnes (1974b), Steck (1984), Mancktelow (1985), supplemented by the authors' own measurements. Block diagram at right was drawn by computer using true perspective. Platelets are tangent to foliation as estimated at sea level, after regional averaging in areas 1 km×1 km square and interpolation using multiquadrics (Barbotin and Cobbold, work in progress).

First, in analogy with the shells of an onion, the region is composed of sheets of Hercynian basement, typically a few kilometres thick, interleaved with thinner sheets of Mesozoic sediments (Schmidt & Preiswerk 1908, Argand 1911, Staub 1924, Amstutz 1971, Milnes 1974a). Early authors described these structures as fold nappes, but Milnes (1974b) showed that large scale folding postdates the emplacement of thrust sheets.

Grossly parallel to the sheets themselves (and to the axial surfaces of major post-nappe isoclines) is a composite foliation at amphibolite grade. We have verified and, where necessary, supplemented the foliation attitudes measured by Schmidt & Preiswerk (1908), Knup (1958), Milnes (1974b), Steck (1984) and Mancktelow (1985). We have then smoothed the data (Fig. 2a) and constructed a simple block diagram (Fig. 2b), showing the central and western parts of the region. Thrust sheets and foliation attitudes form a regional dome.

To a first approximation, the dome has mirror symmetry about a central valley (the Val Leventina) running SSE from Airolo to Bellinzona (Fig. 1). At the bottom of this valley are exposed the lowermost granitic gneisses (Leventina, Lucomagno), considered infra-Pennine by some authors (e.g. Milnes 1974a). The overlying Simano sheet dips gently away to the E and W. Further E, the successively higher Adula, Tambo and Suretta sheets dip eastward by up to 25° on average. They are therefore predicted to occur at depths of up to 20 km or so beneath the overlying Austro–Alpine units which form the eastern boundary of the Lepontine dome, and are attributed to the Adriatic sub-plate. In the SW corner of the region, the uppermost Pennine sheets (Monte-Rosa and Bernhard units) are separated from lower ones by the Simplon Line, which is basically a SW dipping normal fault (Mancktelow 1985). This fault also appears to offset the metamorphic mineral zones (Fig. 1). In the N and S of the area are steep belts (Milnes 1974a), one bordering the External Crystalline Massifs (Aar and Gotthard), the other next to the Insubric Line. Both steep belts are generally considered to postdate the main deformational and metamorphic events (Milnes 1974a, Huber *et al.* 1980, Simpson 1982, Steck 1984, Mancel & Merle 1987). The southern steep belt is in fact the southern limb of an antiformal structure, most easily visible at the western and eastern sides of the dome, where the axis plunges gently to the W or E, respectively. In the W, this structure, known as the Vanzone antiform, folds the Monte-Rosa and Moncucco units. In the E, the units between the Bergell intrusion and the Austro–Alpine units are similarly folded. Notice that mineral zone boundaries have arcuate appearances in both these corners of the dome. Finally, to complete the description of the broadly symmetric features of the dome, we notice that the Simplon Line in the W and the Engadine Line in the E are symmetrically oriented and have similar positions at the two sides of the dome.

As well as these more symmetric features, the dome has some pronounced departures from mirror symmetry (Fig. 2b). Thus we distinguish the Ticino subdome in the E, separated from the Simplon subdome in the W by the N–S trending Maggia steep zone, termed the 'Querzone' by Preiswerk (1918) on account of its anomalous attitude. At its SE end, the Maggia steep zone curves into parallelism with the southern steep belt. Immediately SW of the Maggia steep zone, the foliation traces form a triangular transitional domain, between the Maggia steep zone itself, the southern steep belt and the NE striking Bosco–Gurin structure (Fig. 2a). Throughout much of the Simplon subdome, the foliation strikes NE whereas in the Ticino subdome, it strikes NNW.

Pre-Tertiary history

Relict high-pressure mineral assemblages (eclogites) are well known from the Sesia zone, but have also been described in the central Alps (Ernst 1976, Evans & Trommsdorf 1978, Heinrich 1982). Pressures and temperatures estimated from garnet and clinopyroxene and from the stability of jadeite, have a regional NS gradient, the pressure varying from 9–13 kb in the N to 20 kb in the S (Heinrich 1982, 1986). Although this metamorphic event is imperfectly dated, radiometric data suggest a Cretaceous age (Hunziker *et al.* this volume). The high-pressure metamorphism can be interpreted as resulting from crustal subduction before or during collision between two tectonic plates (Trümpy 1980, p. 80). For how long did this process continue? One possibility is that emplacement of basement slices continued without interruption from the late Cretaceous to the late Oligocene; but this is probably too simple a history. Thus, the P–T conditions of eclogite facies indicate a geothermal gradient lower than average, whereas the P–T conditions of Tertiary metamorphism indicate a geothermal gradient higher than average. A thickened crust with a low geothermal gradient is probably difficult to deform and so an alternative hypothesis is to consider a period between the Cretaceous and Tertiary events, during which there was little deformation. Tertiary deformation due to horizontal forces from the Adriatic sub-plate could perhaps have taken place as the crust became thin enough, by erosion after a late Cretaceous uplift (see Hurford & Hunziker 1985).

In any case, the kinematic record of this early stage of crustal collision is now completely overprinted by strong Tertiary deformations and metamorphism.

Tertiary metamorphism and cooling

Numerous petrological and metamorphic studies show that a high-temperature metamorphism took place during Tertiary times in the Pennine zone (Ernst 1973, Frey *et al.* 1974, 1980). Mineral assemblages are those of the amphibolite grade and mineral zones are nested in a concentric pattern on a map (Niggli 1970). Thus there is a systematic decrease of metamorphic grade from southern regions (Locarno and Bellinzona area) towards the N, E and W. Whereas there is no evidence against a possible sub-horizontal attitude of isotherms during the climax of metamorphism (Frey *et al.* 1980), isograd surfaces are now steeply dipping in the Simplon subdome, at the eastern margin of the Pennine zone and in the northern steep zone (Streickeisen & Wenk 1974, Fox 1975, Thompson 1976, Klaper & Bucher-Nurminen 1987). Thus, at first sight the large scale pattern of isograd surfaces looks like the one expected for a true thermal dome (see Turner 1968). However, large scale folding of originally sub-horizontal isotherms may have occurred after the climax of metamorphism (Milnes 1975, Frank 1983). The near coincidence of the structural and metamorphic domes is highly suggestive of such a folding.

The timing of metamorphic events in the Pennine zone is controversial and this in itself suggests that the notion of a simple thermal dome may be misleading. Geochronological data from muscovite–phengite (Rb–Sr) at the western and eastern periphery of the Lepontine dome (Hunziker 1970, Steinitz & Jäger 1981), outside the staurolite isograd, yield ages of 38–35 Ma. These ages have been interpreted as the peak of Lepontine metamorphism (Jäger 1973). Following this interpretation, the younger ages inside the straurolite isograd all have to be considered as cooling ages. This view assumes that there is a single peak in the time-temperature curve (referred to as the 'Lepontine event') and according to this concept the mineral isograds would approximate an isotherm at one and the same instant in time. The ages of 38–35 Ma are identical with those obtained for the greenschist metamorphic event in the western Alps (Bocquet *et al.* 1974, Frey *et al.* 1974, Chopin & Malusky 1980, Chopin & Monié 1984).

In contrast, ages obtained within the Lepontine dome using the U–Pb method on monazites or K–Ar on amphiboles are much more recent (around 25–20 Ma). Hence some workers have reappraised the chronology of the Alpine metamorphism, claiming that conditions sufficient to produce high grade assemblages were sustained until around 25–20 Ma (Hänny *et al.* 1975, Köppel & Grünenfelder 1978, Köppel *et al.* 1980, Deutsch & Steiger 1985, Monié 1985).

Furthermore, structural investigations (e.g. Thakur 1973, Huber *et al.* 1980, Klaper 1982) suggest that more than one episode of mineral growth occurred in the Lepontine area with respect to established deformation phases. Admittedly these deformation phases do not need to be contemporaneous all over the area considered. However, Rb–Sr ages on micas (Steiger & Bucher 1978, Steiger 1983) more directly suggest a second episode of mineral growth, particularly in the northern part of the Lepontine dome, S of the Gotthard Massif. This mineral growth does not necessarily indicate a second peak in temperature, but may indicate that mineral isograds did not form at the same time all over the Lepontine area.

Thus, the age of the so-called Lepontine metamorphic event remains unclear, possibly because the subject has been considered in terms that are too simple. We suggest that the notion of a single metamorphic event, synonymous with a single period of mineral growth and a single peak in the temperature–time curve throughout the entire Lepontine dome, should be abandoned in favour of a more complex and dynamic thermal history, which has yet to be worked out for particular domains within the Lepontine area. The complexity is due in part to very important post-38–35 Ma events, such as the intrusion of the Bergell batholith (30 Ma, Köppel & Grünenfelder 1975), an Oligocene phase possibly related to the intrusions along the Insubric Line (Laubscher 1983) and particularly the subsequent formation of the southern and northern steep belts (Hurford 1986, Schmid *et al.* 1987, Merle 1987).

Cooling of the Lepontine area has traditionally been attributed to 'uplift', without further explanations of its causes. The use of fission track data provided a major breakthrough in specifying the cooling history. For a region N of Locarno, Hurford (1986) demonstrated a period of very rapid cooling between 23 and 19 Ma, followed by slower cooling rates compatible with current uplift rates (Fig. 3, see also Werner *et al.* 1976, Steiner 1984). Since the formation of the southern steep belt appears to be intimately related to backthrusting along the Insubric Line, the period of rapid cooling is best attributed to uplift caused by backthrusting and backfolding during an Insubric phase (Argand 1916), of Oligocene–Miocene age and contemporaneous with northwards thrusting in the Helvetic nappes (Schmid *et al.* 1987). In this interpretation, the emplacement of the Lepontine area into shallow levels of the Earth's crust is the result neither of isostatic uplift nor of underplating and crustal extension (Platt 1986), but of a Neo-Alpine compressional phase.

The geochronological data also indicate that cooling occurred earlier in the E (Ticino subdome) and later in the W (Simplon subdome) (Jäger *et al.* 1967, Wagner *et al.* 1977). This has led to a model of differential uplift within the Pennine zone (Bradbury & Nolen-Hoeksema 1985). Westward propagation of uplift may be due to dextral transpression in front of a rigid indenter as, for example, the Adriatic subplate with the Ivrea zone at its NW corner (Schmid *et al.* this volume).

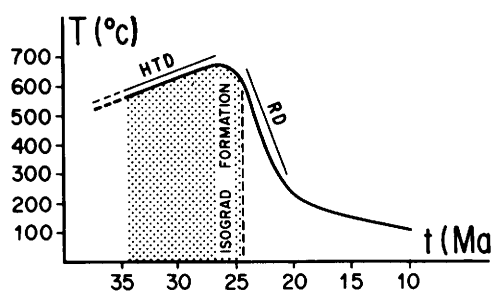


FIG. 3. History of cooling and deformation in the Lepontine dome. Curve gives temperature (T) as a function of time (t) for the last 35 Ma according to model B (II) of Hurford (1986), applicable to the Maggia valley only. There, high-temperature deformation (HTD) ended at about 25 Ma and was followed by retrograde deformation (RD). Within the Simplon subdome, rapid cooling started later than 25 Ma ago (see Schmid *et al.* this volume, Fig. 3).

High-temperature deformation (HTD)

In this section, we refer to deformation that occurred before the onset of rapid cooling. Provisionally, we consider the HTD throughout most of the Pennine zone to have occurred between 40 and 25 Ma. This age bracket is consistent with geochronological data discussed in the previous section and with the culmination of Meso-Alpine shortening in the late Eocene to early Oligocene (38–33 Ma) according to several authors (e.g. Ayrton & Ramsay 1974, Milnes 1978, Hsü 1979, Trümpy 1980, p. 90). Because of the long period of high temperature (40–25 Ma) we cannot further constrain the age of the HTD using metamorphic arguments. We cannot even exclude the possibility that some areas of the Lepontine dome inside the staurolite isograd reached amphibolite grade conditions previous to the event dated at around 38–35 Ma outside this isograd. Furthermore, although the onset of rapid cooling is well dated in the Maggia area at 25 Ma (Köppel & Grünenfelder 1975, Werner *et al.* 1976, Steiner 1984, Hurford 1986), some areas to the N (S of the Gotthard Massif) and W (Simplon domain) were at high temperatures considerably later than 25 Ma ago (Jäger *et al.* 1967, Purdy & Jäger 1976, Wagner *et al.* 1977). Thus we do not consider the HTD to be a well-timed tectonic event, synchronous over the entire area.

Stretch lineation

The main lineation observed in the Pennine zone is a preferred orientation of grain aggregates, individual grains, fold axes, rods and boudins. This structure seems to track the direction of maximum total stretch. Fold axes parallel to the stretch lineation occur at small to intermediate scales (Wenk 1955). Occasional eye structures are probably tranverse sections through sheath folds. We have also observed definite sheath folds in oblique sections (Cobbold 1980).

Optical observations in thin section show that the main foliation and lineation are defined by mineral assemblages of amphibolite grade, such as quartz+plagioclase(An>17)+biotite+staurolite+kyanite+garnet. Thin sections frequently show shear bands associated with the growth of metamorphic minerals. A study of inclusions within garnets from the Simplon area indicates that growth of garnet is contemporaneous with rising temperature and predates the peak of metamorphism in this particular area (Merle *et al.* 1986).

We have measured the attitude of HTD stretch lineations at 882 localities throughout the Lepontine dome (Fig. 4), thus confirming the unusual radiating pattern first described by Wenk (1955). If we compare the attitude of foliation and lineation (see Figs 1, 2 and 4), we see that in the central areas of flat-lying foliation (Ticino and Simplon subdomes), the lineations fan out in directions approximately perpendicular to the peri-Adriatic fault system. Even in the anomalous Maggia steep zone, this radial pattern of flat-lying lineations is not disturbed, the lineations following the strike of the steep zone.

As one approaches the northern steep belt near Airolo, both lineation and foliation acquire steep attitudes, suggesting that they have become folded together about a horizontal E–W axis during later events. Similar geometrical relationships occur as one approaches the southern steep belt, either in the Bellinzona area and further E, where the lineation is nearly vertical, or in the Domodossola area, where it now pitches at about 45°E. This suggests that foliation and lineation were folded together about horizontal axes during a later event in the southern steep belt as well. However the situation is not as simple as this, because for about 50 km between Domodossola and Bellinzona, passing through Locarno, the HTD lineation is flat-lying, often pitching gently E, but sometimes even gently W. If the foliation is restored to the horizontal by rotating about the E–W strike, the lineation remains E–W; whereas immediately to the N, in the flat-lying area, it currently strikes NW. We therefore conclude that the steep belt between Domodossola and Bellinzona cannot have undergone rigid rotation alone during later events. Either the HTD lineation formed in a flat-lying E–W attitude, or it became reorientated during later events as a result of superimposed ductile deformation (strain). This matter is further discussed in the following section.

Kinematic indicators

In our general experience, kinematic indicators are less clear and abundant in rocks of amphibolite grade than they are in rocks of greenschist grade. Nevertheless we have found high-temperature shear indicators in the Lepontine dome, both at outcrop scale (Fig. 5a) and in thin section (Figs 5b, c and d). In deformed granites, we mainly used the relative attitudes of foliation and shear bands with amphibolite

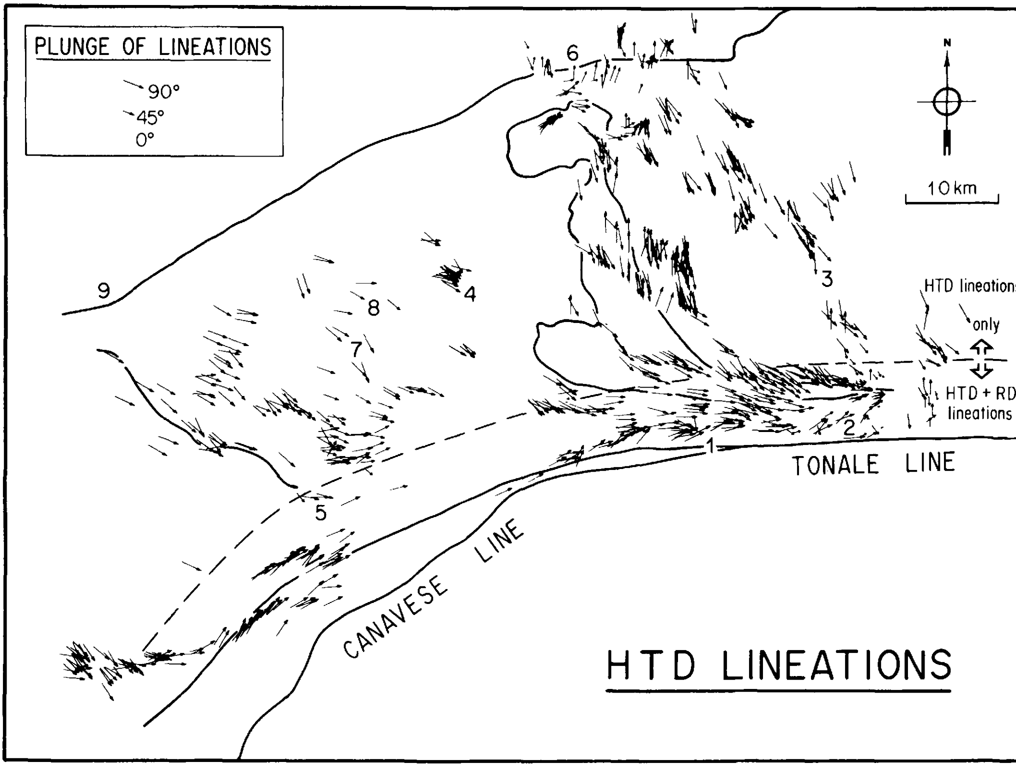


FIG. 4. Current attitude of high-temperature stretch lineations in central and western parts of Lepontine dome (new data only). Each arrow represents a single measurement. Length of arrow is inversely proportional to plunge in degrees (see key). Solid curves are a selection from the geological boundaries of Fig. 1. Numbered localities are as in Fig. 1. The age of the lineations in the southern steep belt (south of the solid broken line) cannot always be inferred with certainty: there are HTD lineations rotated by the backfolding as well as lineations formed during the RD event. For lineations within the Insubric mylonite belt see Schmid *et al.* (1987) (data from the Monte-Rosa area by E. Barbotin).

grade assemblages (C/S structures of Berthé *et al.* 1979), as well as the sigmoidal shapes of recrystallized tails around old grains (Simpson & Schmid 1983, Passchier & Simpson 1986). Such structures, together with asymmetric folds of the foliation, are especially clear in the Leventina orthogneiss (Fig. 5). In paragneisses we had greater difficulty in determining shear senses, but sometimes, like Mawer (1987), we were able to use shear bands, asymmetric folds, rolling structures or rotated boudins (Van Den Driessche & Brun 1987).

In the flat-lying areas of the Lepontine dome, HTD shear senses are consistently top to N to WNW in the lineation direction (Fig. 6). In most areas, there is little or no retrograde overprint, so that these shear directions appear to be original ones, operative during the HTD event. An exception is the Simplon area, where there is a retrograde overprint, discussed in a later section.

In the northern steep belt around Airolo, HTD shear indicators appear to have become folded into an upright position, together with foliation and lineation. Restored to a horizontal attitude, by rigid rotations about E–W axes, the shear indicators have top to N senses, as in the centre of the Ticino subdome. Further evidence for such rotations will be discussed later.

In the southern steep belt, HTD shear indicators have been found so far at two localities (Fig. 6). One is in the Monte-Rosa unit at Villadossola, near Domodossola (Schmid *et al.* 1987); the other is in the Monccucco unit at Ponte Brolla, near Locarno. Both yield dextral strike-slip shear senses in almost perfectly horizontal directions. Were these horizontal shear directions operative during the HTD event, or have the structures been rotated? If foliation is restored to the horizontal, by rigid rotation about strike, the shear senses become top to W, compatible in sense, but not in direction, with

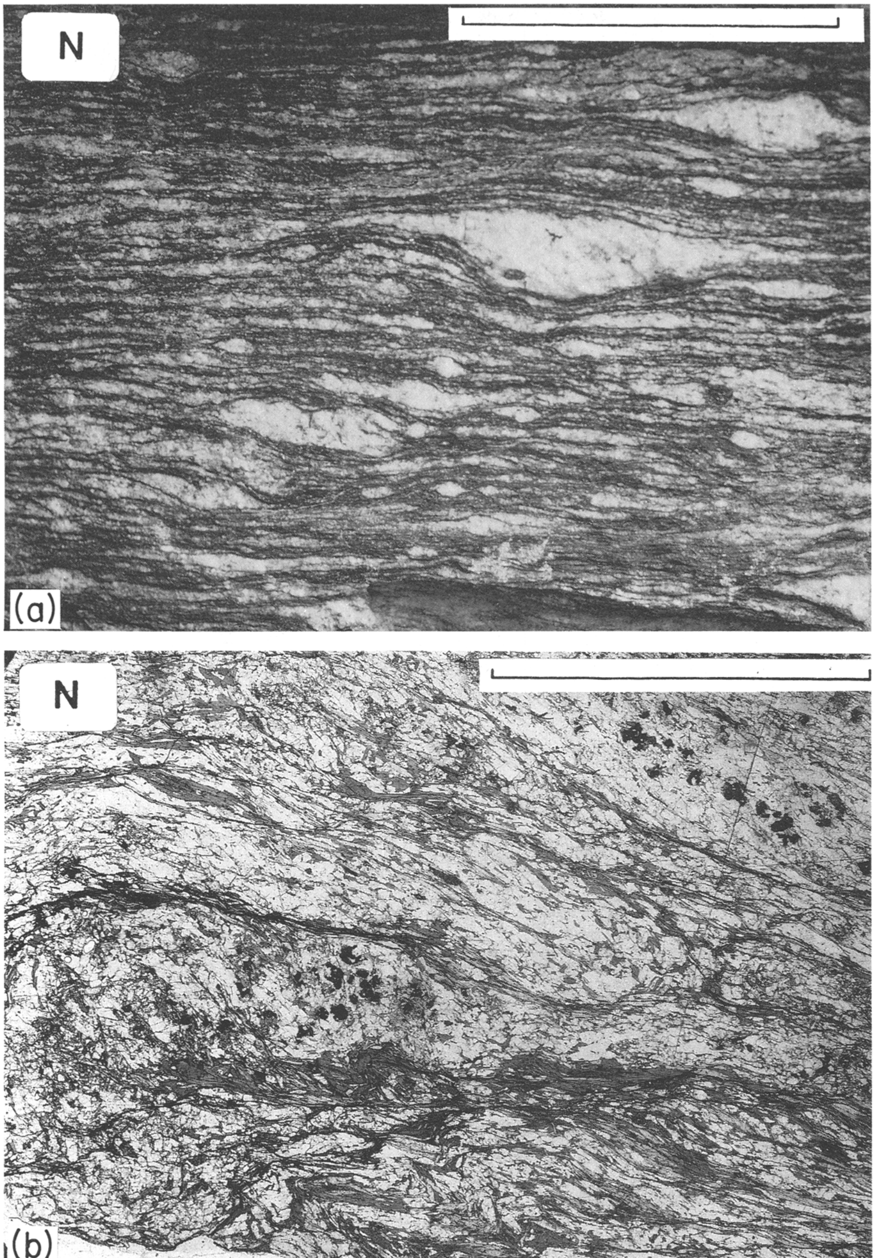
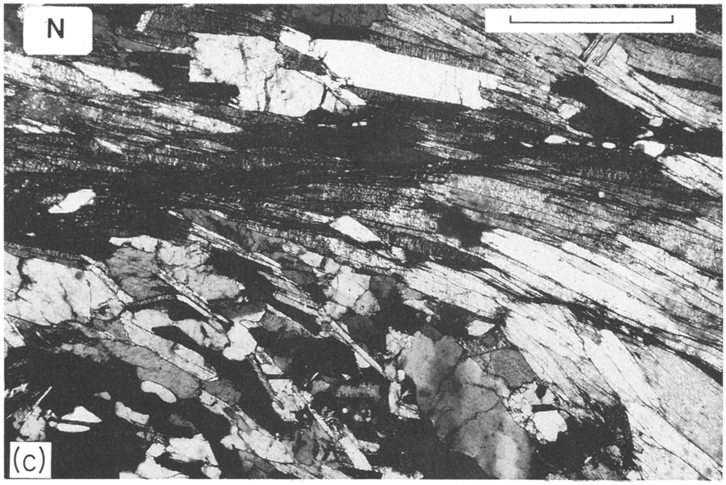


FIG. 5. Kinematic indicators for high-temperature deformation. (a) Shear bands and sigmoidal porphyroclasts in the flat-lying Leventina gneiss. Photograph of vertical outcrop (N at left), scale bar 5 cm long. (b) C–S structures and folds in the flat-lying Leventina gneiss (thin section, vertical, N at left, scale bar 2.5 cm long). (c) and (d) Detail from (b), showing shear bands of top to N sense, containing assemblages of two micas, feldspar and quartz. Dynamic recrystallization of potash feldspar (c) and dynamic recrystallization of quartz by grain boundary migration (d) indicate high temperatures during formation of shear bands. Scale bars 0.83 cm long (c) and 0.20 mm long (d).



the top to NW shears observed in the flat-lying areas immediately to the N. Thus a component of rigid rotation during later events is not impossible; but it does not appear to be sufficient in itself to account for the present attitudes, as discussed in detail in the previous section on HTD lineations. We conclude that there has been additional ductile strain in this part of the southern steep belt. At the Ponte-Brolla locality, shear indicators (shear bands and asymmetric folds and boudins) are coeval with the beginning of anatexis (Gapais, personal communication, 1987). The rocks are migmatitic and there is no evidence of a retrograde ductile overprint. We therefore conclude that dextral strike-slip at Ponte-Brolla occurred at high

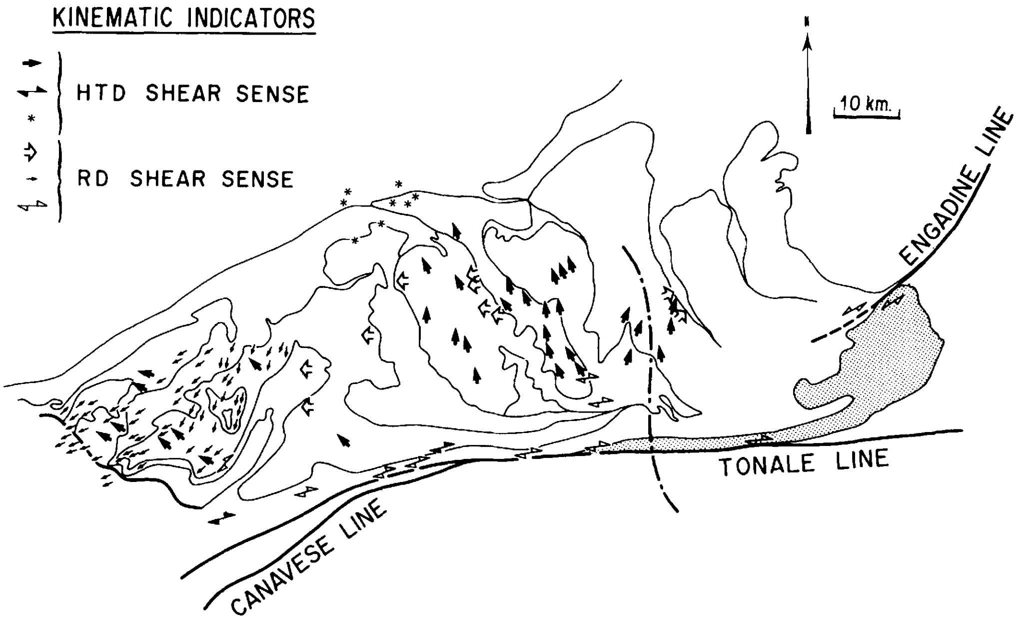


FIG. 6. Shear senses in the Lepontine dome. Each arrow represents a sense of shear, determined from a population of kinematic indicators at outcrop and in thin section, where such a population gave consistent results at a scale of about 100 m. In central areas of flat-lying foliation, horizontal shear components are shown by single arrows representing motion of roof (large black arrows for HTD, wide empty arrows for RD in Ticino subdome, small black arrows for RD in Simplon subdome). Note the reversed sense of shear in the Berisal and Glishorn backfolds. In the southern steep belt and along the Engadine Line, wrench components are shown by zigzag arrows (full black arrows for HTD, empty arrows for RD). In the northern steep belt, asterisks show localities where HTD shear directions are currently vertical (for interpretation, see Fig. 10). East of dashed line, shear-sense indicators have been observed recently at outcrop (by P. R. Cobbold, M. Ballevre, D. Gapais and E. Le Goff), but not yet in thin section, so that metamorphic conditions are not yet reliably known.

temperature. Unfortunately we do not know whether it occurred in the age bracket of 40–25 Ma, or at the beginning of the Insubric phase.

Retrograde deformations (RD)

In this section, we refer to deformations coeval with the cooling of the Lepontine area, i.e. from 23 Ma to 19 Ma in the Ticino subdome (Hurford 1986), and significantly later in the Simplon subdome (Bradbury & Nolen-Hoeksema 1985).

Geological mapping of stretch lineations (Fig. 7), shear indicators (Fig. 8a, b) and other structures reveal late ductile deformations (Steck 1984) superimposed under retrograde metamorphic conditions upon the HTD (Merle 1987, Merle *et al.* 1986). The closing stages of retrograde deformation (RD) demonstrably occurred under greenschist facies conditions in the Simplon area (Mancktelow 1985, Mancel & Merle 1987). Further E, the RD is only occasionally associated with syntectonic growth of chlorite. From studies of the RD, we infer different kinematic histories in the Simplon and Ticino subdomes. The southern and northern steep belts also have rather special histories.

Southern steep belt

The formation of the southern steep belt has been shown to be contemporaneous with mylonitization under greenschist facies conditions along the Insubric Line (Zingg & Schmid 1983, Schmid *et al.* 1987), during the Insubric phase in the sense of Argand (1916). Because the Insubric mylonites formed at the interface between the warm Lepontine domain and the earlier cooled southern Alps (>500°C vs. 150°C at 25 Ma, see Hurford 1986, fig. 7), this retrograde overprint is particularly obvious within the mylonite belt, which simply represents the southern margin of the steep belt. In the SW of the region, the central Alps were backthrust

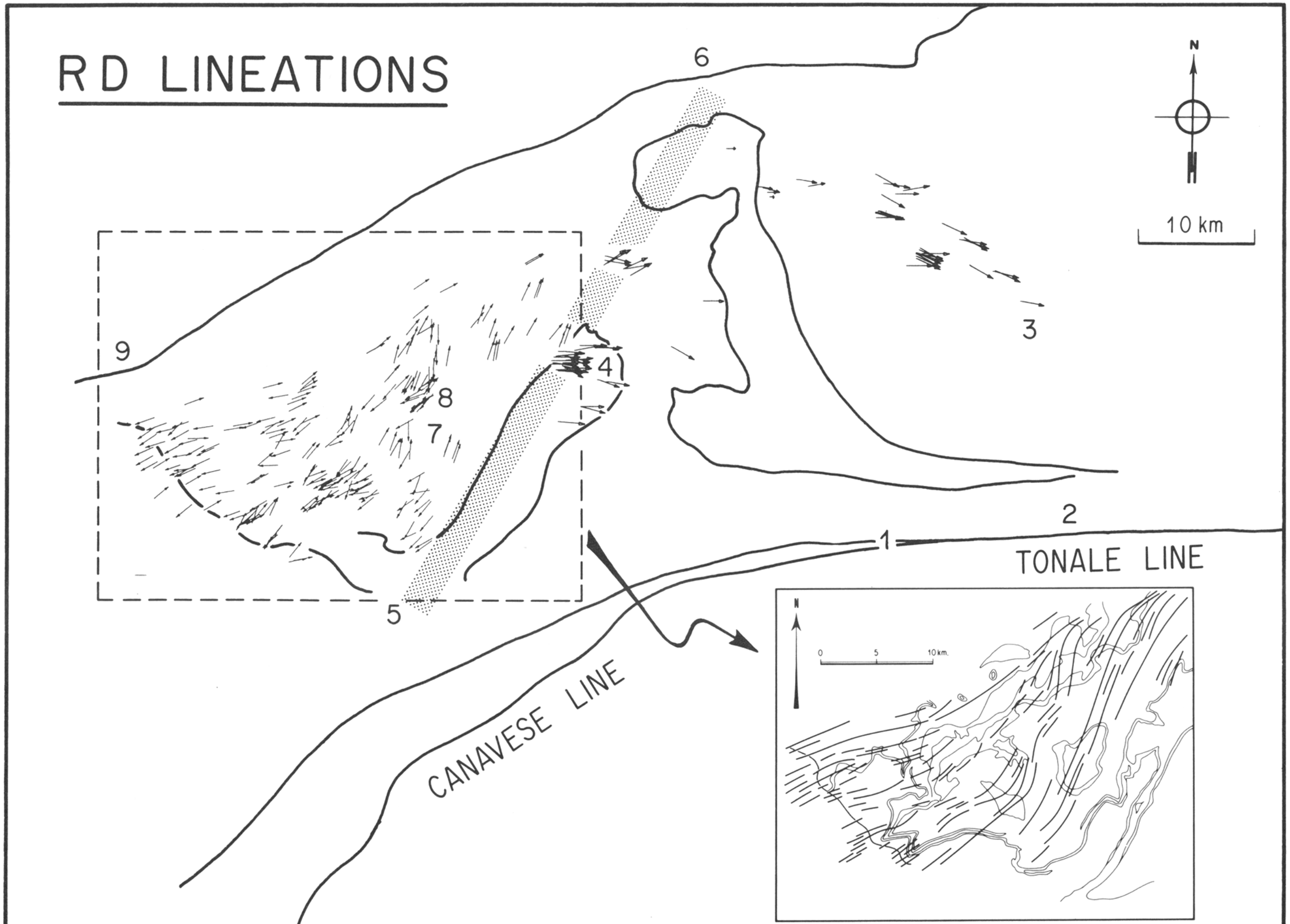


FIG. 7. Current attitude of retrograde stretch lineations in central and western parts of the Lepontine dome (new data, and old data from Merle 1987 and Merle & Le Gal 1988). Stippled zone separates lineations in the E (associated with formation of Ticino subdome, Maggia steep zone and Bosco–Gurin thrust) from later lineations in the W (associated with formation of Simplon subdome and normal faulting along Simplon Line). For legend, see Fig. 4. Inset at bottom right shows lineation trajectories (drawn by hand) in the Simplon domain. For senses of shear, see Fig. 6.

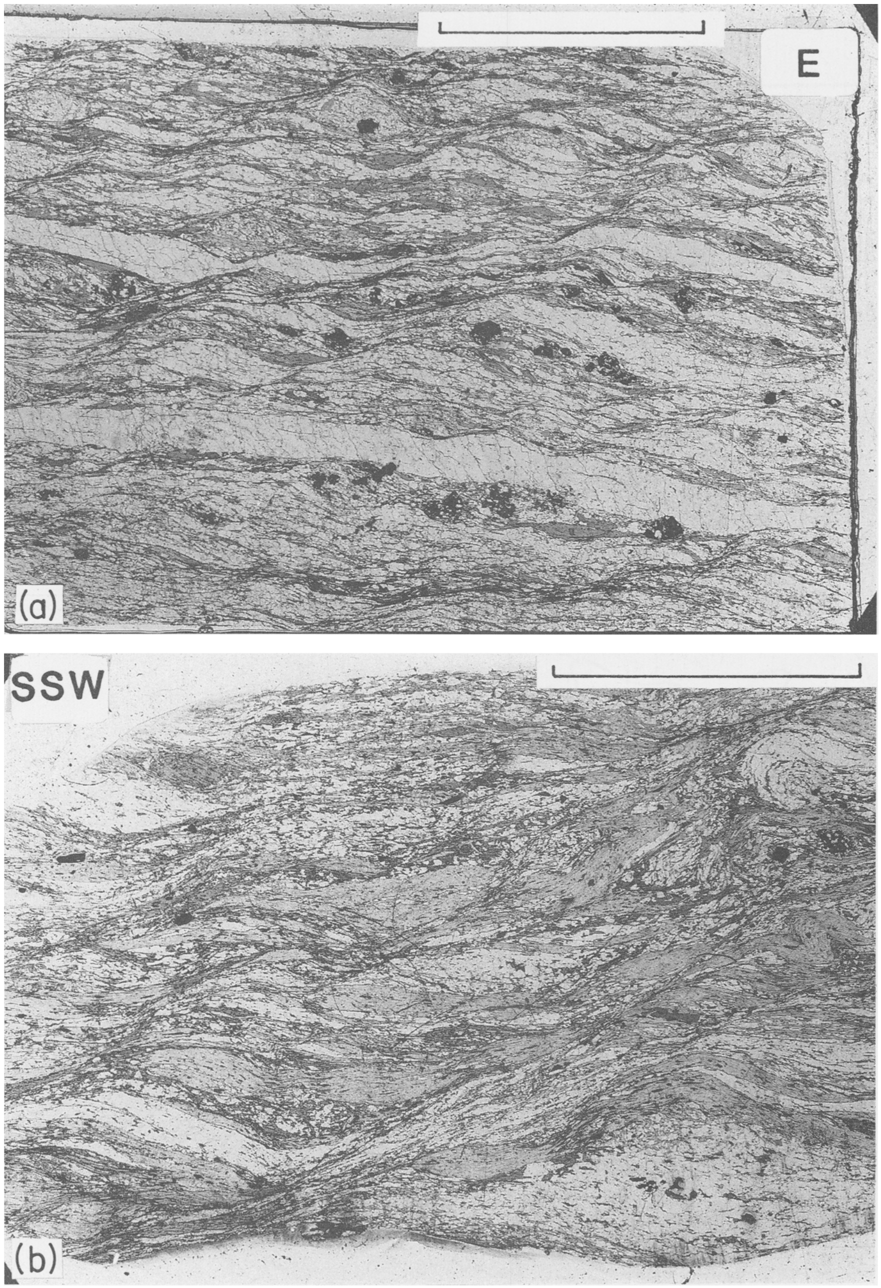
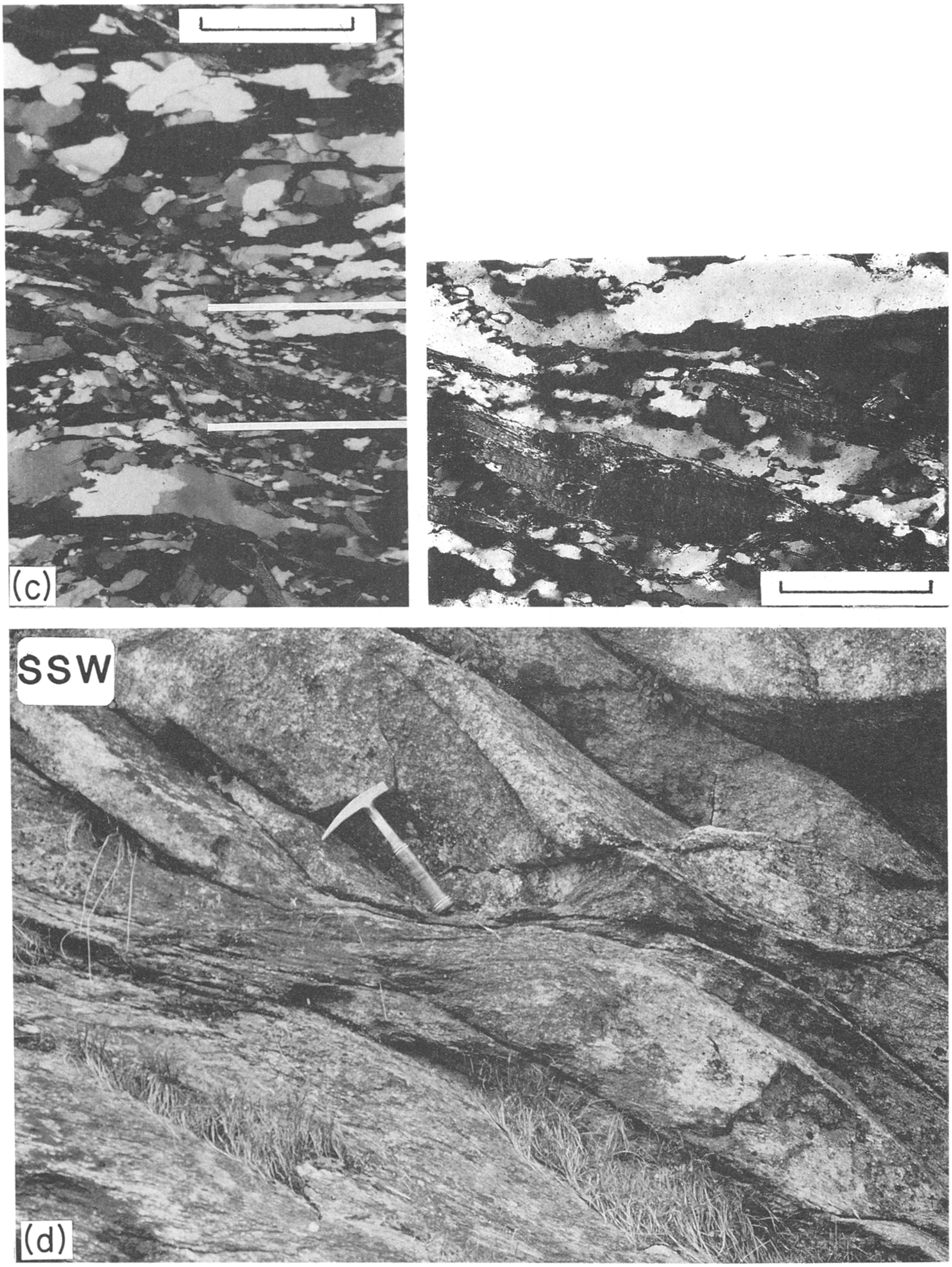


FIG. 8. Kinematic indicators for retrograde deformation. (a) Shear bands at contact between Simano nappe and Leventina gneiss (vertical thin section, E at right, scale bar 1 cm long, see Fig. 9 for locality). Sense of shear is top to W. (b) Shear bands and fold in Baceno schist, Verampio window (vertical thin section, SSW at left, scale bar 2 cm long). Sense of shear is top to SSW. Chlorite content about 20% (c) Shear bands within a quartzitic gneiss near base of Bosco-Gurin thrust (see Fig. 9 for locality). Microstructure of coarse-grained quartz outside a shear band (left) indicates migration recrystallization typical of high temperature deformation. Within same shear band (right), serrate grain boundaries and incipient recrystallization by mechanism of subgrain rotation indicate lower temperatures. Since shear bands represent late instabilities along a continuous strain path, we deduce that shearing occurred while rocks rapidly cooled (compare Fig. 3). Scale bars 0.83 mm long (left) and 0.25 mm long (right). (d) Flat-lying shear zone in Antigorio granite, Lago d'Agaro (locality 8, Fig. 1). Vertical outcrop, SSW at left, hammer gives scale. Shear sense is top to SSW.



over the southern Alps along the N-dipping Insubric mylonite belt, as demonstrated by small scale shear criteria and steep stretch directions (Heitzmann 1987b, Schmid *et al.* 1987). Kinematic studies indicate that backthrusting was followed by dextral shearing in these mylonites. Although this sequence of events can be established locally, both events are best thought of in terms of a continuous process of dextral transpression on a larger scale (Schmid *et al.* this volume). Only during the very latest stages along the Insubric Line (cataclastic deformation at the Tonale and Centovalli lines, greenschist facies mylonitization of Lepontine marble

Laubscher 1971, Fumasoli 1974, Heitzmann 1987a, Schmid *et al.* 1987) did oblique convergence across the belt change into almost pure strike-slip.

Because of the existing N–S temperature gradient during the Insubric phase, it is dangerous to use ambient temperatures to decide on the relative ages of deformation. The retrograde nature of movements N of the mylonite belt is not always obvious, since higher temperatures prevailed for a longer period of time in the northern parts of the Lepontine area (Hurford 1986). In the Valle d'Ossola section, dextral strike-slip (Monte-Rosa nappe at Villadossola near Domodossola, Schmid *et al.* 1987) and backfolding (Vanzone antiform, Klein 1978, Milnes *et al.* 1981) can be attributed to this Insubric phase. Dextral shearing occurred under retrograde conditions all along the southern steep belt (Fig. 6 and Heitzmann 1987b), but also under high-temperature conditions (Villadossola and Ponte Brolla localities). The time relationships between dextral shearing and backfolding are not established yet and it is likely that both components of motion were synchronous in many parts of the southern steep belt, as postulated for the Insubric mylonites. In any case, the vertical uplift associated with backthrusting appears to have reached a maximum between Locarno and Bellinzona.

Ticino subdome

In the W, from Biasca to Bosco–Gurin, the Ticino subdome displays a set of RD stretch lineations of nearly EW trend, superimposed on the HTD stretch lineations (Fig. 7). This second deformation is scarce at the scale of the subdome, but where it occurs it is usually associated with flat-lying ductile mylonites. Thus, mylonites at least 10 m thick occur at the contact between the Leventina gneiss and the Simano sheet (Swiss National Grid 706.2–141.45) and between the Antigorio granite and Bosco–Gurin units (SNG 678.3–128.8). Shear senses determined at outcrop or from thin sections (Fig. 8a) are all top to W.

The retrograde character of this deformation is apparent in thin section. Chlorite has grown preferentially within shear bands that postdate the growth of amphibolite-grade porphyroblasts. However, we have never observed completely retrograde assemblages (as described later for the Verampio window or the Simplon line). Quartz microstructures within microscale shear bands often indicate progressive deformation at decreasing temperatures (Fig. 8c) and this can probably be correlated with the cooling recorded by geochronological data.

Westward shearing may be a key for understanding major structures such as the Maggia steep zone and at the duplication of the Antigorio granite in the Bosco–Gurin area. An E–W cross section through the Maggia steep zone (Fig. 9) shows a large fold overturned towards the W (see Merle & Le Gal 1988 for a fuller description). This structure appears to fold the earlier high-temperature foliation. Retrograde mylonite zones and later brittle faults all show top to W senses of shear or slip. We therefore interpret the overall structure as being modified, if not created, by top-W shearing during the RD event.

In the Bosco–Gurin area, the Antigorio granite is duplicated (Figs 1 and 9). There is a well developed RD stretch lineation striking E–W, parallel to the fold axis of the Wandfluhhorn fold (Hunziker 1966, Hall 1972, see Wandfluhhorn peak located on Fig. 1). In the core of the upper Antigorio sheet, strain is homogeneous and shear indicators are lacking, except at the base, where the granite, in contact with underlying paragneisses (Bosco–Gurin unit), becomes mylonitic in a thin zone dipping gently E. The sense of shear is top to W. Thus, the duplication of the Antigorio granite may be due to overthrusting towards the W (Merle & Le Gal 1988). The northern limit of the Bosco–Gurin thrust remains to be recognized in the core of the Antigorio granite itself, but the Wandfluhhorn fold may be a ductile termination of the Bosco–Gurin thrust. To the S, next to the southern steep belt, the Antigorio granite is refolded in a synformal structure (the Masera synform). On a map, the hinge of this fold curves in a manner consistent with a dextral sense of shear in the southern steep belt.

The westward deformation in the Ticino subdome is considered to be coeval with the formation of the southern steep belt. We believe that both events occurred at the western and southern margin of the uplifting Ticino subdome between 25 and 20 Ma, a period of rapid cooling in the Ticino subdome according to geochronological data (e.g. Werner *et al.* 1976, Wagner *et al.* 1977, Hurford 1986) (Fig. 3).

On the eastern side of the Ticino subdome, within the Adula unit, but still within the limit of Tertiary amphibolite-facies metamorphism, retrograde shear indicators with normal-fault sense (top to ENE) have recently been observed (Fig. 6). The metamorphic conditions of their formation and their age relationships with the shear indicators in the western part of the dome are not yet clearly determined.

Finally, in the steep zone that prolongs the

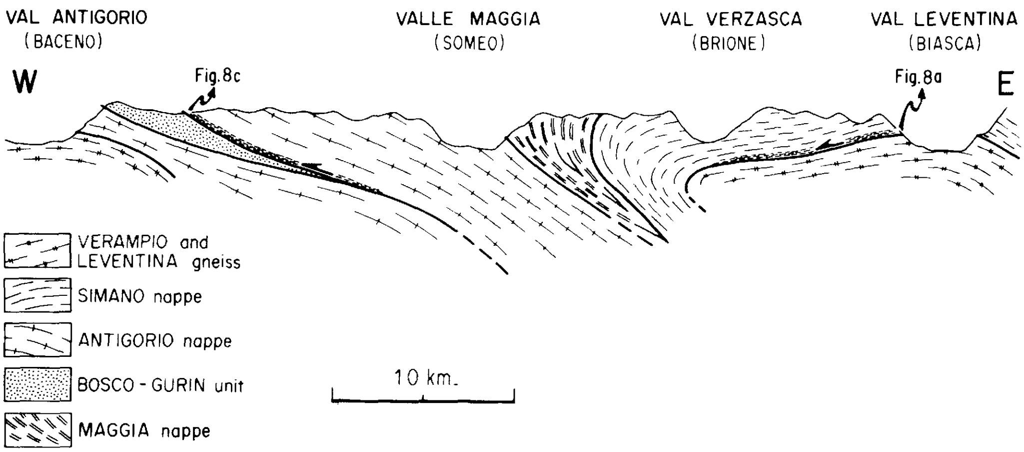


FIG. 9. Schematic geological cross-section running EW from Val Leventina to Val Antigorio across Maggia steep zone and Bosco-Gurin thrust (after Merle & Le Gal 1988). Section line is shown in Fig. 1. The Wandfluhhorn fold cannot be seen as its axis is parallel to the cross section. Arrows show shear senses in retrograde mylonites (see Figs 6, 8a, c).

Engadine Line to the SW (Fig. 1), as well as in the northernmost exposures of the Bergell intrusion, retrograde shear indicators reveal left-lateral wrench components along strike (Fig. 6). Once again, the metamorphic conditions of formation are not yet known, but the structures postdate the Bergell intrusion (30 Ma).

Simplon subdome

Throughout a well-defined triangular area between the Simplon Line and the northern lobe of the Maggia unit, later flat-lying foliations and lineations partially overprint older HTD structures (Steck 1984). The second deformation occurred under retrograde conditions, at temperatures lower than in the Ticino subdome (Merle 1987).

The stretching lineation in the field is mainly defined by an alignment of quartz aggregates, muscovite and chlorite. Coeval shear bands postdate the growth of garnet, staurolite or kyanite porphyroblasts; chlorite aggregates are usually located within microscopic shear bands. Thin sections of samples from the Verampio window and the Simplon line indicate P–T conditions of greenschist facies (Fig. 8b). Shear bands developed within the domain of stability of biotite but generally were contemporaneous with the breakdown of this mineral.

The RD lineations form a consistent, slightly radiating pattern (Fig. 7), restricted to an area of nearly flat-lying units, surrounded by zones of steeper attitude (including the Bosco–Gurin structure). Towards the W, the extent of this area of retrograde deformation beyond the Simplon Line is not yet known (Merle 1987, Mancel & Merle 1987, Steck 1987). The sense of shear, demonstrated by C/S structures, shear bands and asymmetric pressure shadows and porphyroclasts, is top down and SSW in the Verampio window, top down and WSW at the Simplon Line (Fig. 6) (Merle 1987, Mancel & Merle 1987). We correlate this deformation with the Simplon phase described by Mancktelow (1985, 1987) along the Simplon Line. At a few localities in the core and front of the Antigorio granite nappe, the shape of quartz aggregates indicates that the strain ellipsoid is oblate (Greco 1985, Daniel 1986). This strain is due to the retrograde deformation alone, high-temperature aggregates having been observed only at the base and rear of the granite sheet. We suggest that the oblate strain is due to centrifugal radial motion along the divergent RD trajectories of Fig. 7. The strain intensity appears to increase towards the Simplon Line, but is also very high in some ductile units (Baceno schist) exposed in the Verampio window (Merle 1987).

Folds with axes nearly perpendicular to the stretch direction and asymmetries consistent with the bulk sense of shearing are frequently observed in the Simplon subdome. Non-cylindrical folds can sometimes be seen, but the strain is not strong enough for complete reorientation of fold axes into the stretch direction.

The RD in the Simplon subdome must have occurred later than about 15 Ma, an age which represents the end of the amphibolite grade in this area (Jäger *et al.* 1967, Purdy & Jäger 1976, Wagner *et al.* 1977). This age is consistent with the time bracket of 19–6 Ma suggested by Mancktelow (1985) for the deformation recorded along the Simplon Line. On the basis of young ages (Frank & Stettler 1979) in fine fractions of mica, the deformation along the Simplon Line may indeed have lasted until 9 Ma. Thus, it is likely that the Simplon phase postdated the formation of the southern steep belt and the RD deformation in the Ticino subdome (20–25 Ma, see discussion in the previous section).

Northern steep belt

The northern steep belt is a region of backfolding located at the boundary between the Pennine zone and the external crystalline massifs (Huber *et al.* 1980). The backfolds have been described as late structures postdating the high-grade metamorphism in the Pennine zone (Higgins 1964, Chadwick 1968, Cobbold 1969, Thakur 1973, Huber *et al.* 1980, Simpson 1982).

The chronological relationships between high-temperature deformation (HTD) and backfolds are especially clear N of the Ticino subdome (Airolo area and Lucomagno section). Small scale HTD shear indicators (C/S structures in the Tremola series of the southern Gotthard Massif) show an apparent uplift of southern units with respect to northern units along the vertical stretch lineation (Fig. 10). Low-temperature flat-lying shear bands and shear zones (top to S) postdate the HTD indicators and are associated with the backfolding event (Fig. 10). They become prominent in the orthogneiss of the Gotthard Massif which seems to have been less affected by the HTD deformation (see Choukroune & Gapais 1983, Marquer & Gapais 1985). Thus we conclude that the HTD indicators were rotated into an upright position during backfolding. Backfolds postdate well-developed biotites which have been dated at 15 Ma (Steiger & Bucher 1978, Ramsay in Steck *et al.* 1979).

North of the Simplon subdome, the senses of shear associated with the retrograde stretching lineations (depicted in Fig. 6) are reversed between the Berisal synform and the Glishorn antiform (Fig. 1) (Mancel & Merle 1987, Fig. 8). This reversal is interpreted to be not original, but due to backfolding about axes near-parallel to the earlier retrograde lineations. Hence, a very young age of less than 10 Ma is inferred for backfolding within the northern steep belt.

Kinematic history

Using the structural, metamorphic and geochronological data presented or discussed above, we now interpret the kinematics of the Lepontine dome in three stages, which we consider in reverse order (younger to older):

(1) During the youngest stage (about 10 Ma to present), the Lepontine dome appears to have been rigid, except at its margins. In the northern steep belt between Brig and Ilanz, ductile deformation was expressed as backfolding and southwards overthrusting (Fig. 10). Ductile deformation also occurred in the external zones to the N and is responsible for the N6OE striking foliation of the crystalline massifs (Choukroune & Gapais 1983, Marquer & Gapais 1985). To the S of the Lepontine dome, dextral strike-slip motions occurred on the Insubric and Tonale faults. All these ductile and brittle motions we attribute to dextral transpression across the Insubric Line.

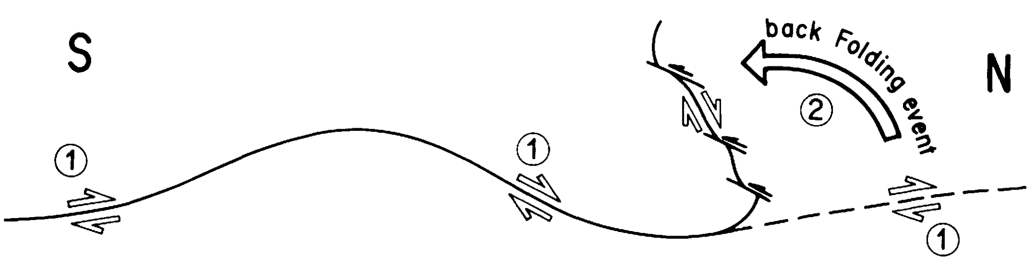


FIG. 10. Schematic N–S section through northern steep belt (Airolo and Lucomagno area, see Fig. 1), illustrating shear indicators (large hollow arrows) of earlier HTD event (1), flat-lying in the S, but steepened and overturned in the N, as a result of later back-folding event (2). Late retrograde shear bands (small arrows) are associated with backfolding.

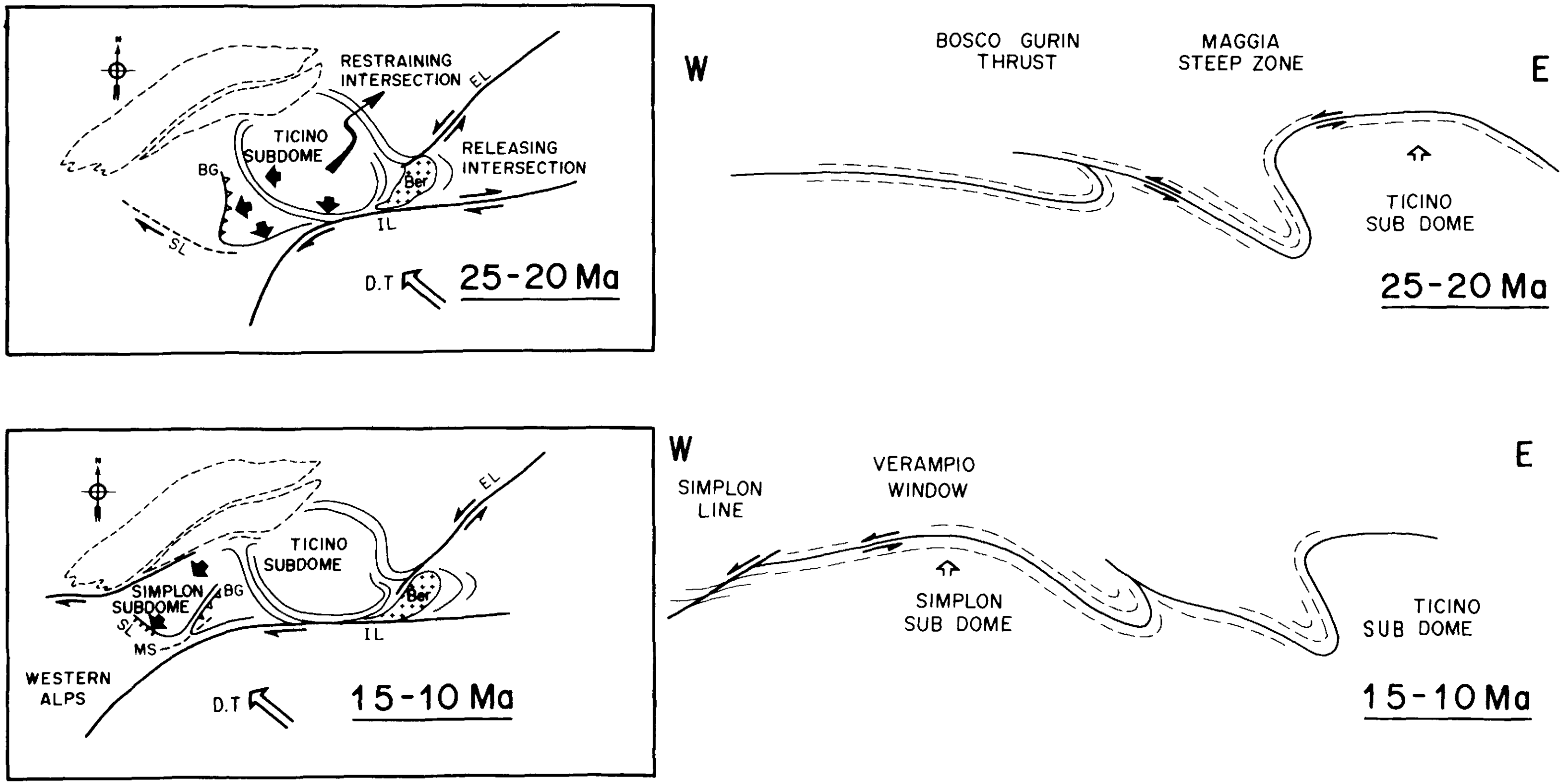


FIG. 11. Tertiary kinematics of the Lepontine dome during uplift (schematic interpretation). Two stages are shown at top (25–20 Ma) and bottom (15–10 Ma), in plan (left) and E–W section (right). In plan, approximately northwestward displacement (DT) of Adriatic subplate relative to Europe results in northwestward shortening within Lepontine area, dextral transpression across Insubric line (IL), and serial uplift of Ticino subdome (top), followed by Simplon subdome (bottom). Present day shape of external massifs merely serves as frame of reference. Full black arrows indicate horizontal components of shearing (roof motion), double arrows indicate wrench components. Ber is Bergell intrusion; BG, Bosco-Gurin thrust; SL, Simplon Line; EL, Engadine Line. The axial-plane trace (dashed line) of Masera synform (MS) curves in a manner consistent with dextral shearing parallel to Insubric Line. Cross sections show enveloping surfaces of foliation, main structures, retrograde shear indicators (black arrows) and uplift of subdomes (empty arrows).

(2) The Insubric phase (about 25–10 Ma) is best considered as twofold (Fig. 11). The kinematics are closely related in time to the uplift history of the Lepontine dome. If we attribute the back-thrusting along the southern steep belt to the converging Adriatic sub-plate, it seems sensible to expect some lateral escape of rocks within the Pennine zone at the same time, especially at releasing intersections of conjugate strike-slip faults such as the Engadine and Insubric Lines. The two regional subdomes of the Pennine zone (see the map of foliation attitude, Fig. 2a) could be due to these major retrograde deformations, coeval with uplift, whose importance seems to have been underestimated until now. An earlier dome (Ticino dome, 25–20 Ma) to the E and a later one (Simplon dome, 15–10 Ma) to the W of the Pennine zone might thus be attributed to continuing dextral transpression at the boundary between both tectonic plates (Fig. 11). The domes could have formed in an area of excess shortening, located in front of the indenting Adriatic sub-plate and at the restraining intersection of the Insubric and Engadine strike-slip fault zones, whereas horizontal extensions were concentrated at the releasing intersections to the E and W of the domes (Fig. 11). The NE–SW direction of extension in the Simplon subdome is kinematically compatible with NW–SE shortening and dextral shearing. The radial stretch trajectories accompanying the extension in the Simplon subdome are perhaps attributable to strong gravitational effects (i.e. a spreading process) contributing to unroofing of the subdome. Physical and numerical models of such gravitational processes characteristically show (a) divergence of stretch trajectories from topographic highs (b) oblate strain ellipsoids resulting from a combination of divergent horizontal shearing and vertical shortening and (c) strain intensities that increase downslope (see Hudleston 1983, Gilbert & Merle 1987).

(3) Before about 25 Ma, the Lepontine area underwent radial shearing (top to N or NW). Restoration of the younger backthrusting along the Insubric Line suggests that the major tectonic units were southerly dipping before backthrusting. If so, the radial motion implies an overthrusting to the N and NW at a level deep enough to sustain amphibolite facies metamorphism. Thus the high-temperature deformation is attributable to Tertiary convergence between Europe and the Adriatic sub-plate. The general direction of this convergence would appear to have been NW–SE, as during the later Insubric phase; but because motion was roughly radial, centrifugal motion of the hangingwall would have led to arc-parallel extension. Radial motions of this kind have been described in experiments scaled for gravity (Davy & Cobbold in press). In such experiments, the hangingwall deforms more readily than the footwall. If we apply these ideas to the Pennine zone, we infer arc-parallel extension at upper levels. Oblate mineral aggregates of amphibolite grade, frequently observed in flat-lying areas of the Ticino subdome, suggest that arc-parallel extension may indeed have operated at this stage, preventing strong crustal thickening and perhaps facilitating initial uplift of the Lepontine dome. The Insubric and Engadine Line may also have been active at this early stage, their releasing intersection facilitating emplacement of the Bergell intrusion some 30 Ma ago (see Trümpy 1980, p. 77).

To conclude, all three stages in the Tertiary kinematic history inferred for the Lepontine dome can be attributed to dextral transpression across the Insubric and Tonale Lines. Uplift of the dome was mainly due to backthrusting along the southern steep belt, but was aided by lateral escape of rocks and an associated component of horizontal regional extension within the Lepontine area.

Concluding remarks

From our structural study of the Lepontine area, we conclude that the central Alps during Tertiary times underwent a tectonic history rather different from those usually described for full frontal collision. Whereas the western Alps appear to have undergone frontal collision with the Adriatic subplate (Menard & Thouvenot 1984), the central Alps were subjected to dextral transpression, with relatively little crustal thickening. This study also suggests that post-collisional late Oligocene and early Miocene deformations have so far been underestimated in the Lepontine area.

A reconstruction of the main tectonic features in N–S section cannot show the effects of strike-slip motion and lateral escape, but it does indicate that backthrusting of the Pennine zone over the southern Alps is one of the main features of central Alpine history during the

Tertiary (see, e.g. Fig. 8 in Schmid *et al.* this volume, and fig. 3 in Milnes 1978). Taking into account the northwestward overthrusting in the external domain, the history of the Pennine zone at this time is that of an erosional pop-up, which is probably the best model for explaining the rapid cooling recorded at around 25–20 Ma and the emplacement of amphibolites into shallow levels of the Earth's crust.

During the closing stages of deformation, a similar but smaller pop-up uplifted the external massifs and formed the northern steep belt, with associated minor backthrusting in the northern Pennine zone.

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