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Dry high-temperature shearing in the fossil Hercynian lower crust of Calabria (southern Italy)

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ABSTRACT. — In the Hercynian lower crust of Calabria, post-Hercynian shearing occurred during retrograde amphibolite to sub-greenschist facies conditions and was localised in mm to cm thick shear zones. With the aid of (*i*) petrological and microstructure-based thermobarometry, (*ii*) textural analyses, and (*iii*) infra-red spectroscopy, the timing, P-T conditions and kinematics of shearing are exemplified. (1) The general top-to-the-NW transport of the lower crustal section is reflected by the shear zone. (2) A period of annealing is bracketed by a stage of high-temperature and one of low-temperature shearing. (3) Even in ultramylonitic layers and in the greenschist facies, shearing occurred in dry conditions. (4) The shear zone was active for approximately 60 million years. (5) After the first formation of a shear zone, the concentration and continuation of movements was not necessarily governed by the infiltration of fluids but by the presence of a texturally weak zone. (6) Despite the general problems of exchange-reaction-based thermobarometry in shear zones, in specific cases – like the one presented here – reliable results may be obtained.

RIASSUNTO. — La crosta inferiore Ercinica della Calabria è stata interessata da deformazioni di taglio (da millimetriche a centimetriche) ass a metamorfismo retrogrado (da facies anfib a facies degli scisti verdi). La successione tem le condizioni P-T e la cinematica di questi p sono state ricostruite tramite: (i) ri termobarometriche di tipo microtessi (ii) analisi strutturali, e (iii) spettro all'infrarosso. Importanti sono risultati i s aspetti: (1) le caratteristiche delle deforn indicano un trasporto generale verso Nor della crosta inferiore. (2) Processo di «ann da alta a bassa temperatura. (3) Le defor responsabili dei livelli ultramilonitici degli scisti verdi) sono avvenute in cor virtualmente anidre. (4) Le deformazioni dovrebbero essersi realizzate nell'arco 60 milioni di anni. (5) Dopo le prime defo di taglio, la concentrazione ed il pro dei movimenti tettonici non sarebbe controllati da infiltrazione di fluidi, ma da debolezza strutturale. (6) I risultati in questo caso specifico, nonostante la di incertezze sulle stime termobarometric su reazioni di scambio in zone di taglio, son affidabili.

KEY WORDS: *Shear zone, dry shearing, P*

INTRODUCTION

In the northern Serre (Southern Calabria, Italy) a 7-8 km thick continuous section of the Hercynian lower crust is exposed (fig. 1). It is subdivided into metapelite and granulite-pyriclasite units. The P-T-t history has been reconstructed by Schenk (1984, 1985, 1990). Subsequent to a Hercynian granulite facies peak of metamorphism at 300±10 Ma, the lower crustal section was isothermally lifted by 5-7 km to a mid-crustal level, then cooled from 700-800°C to about 200°C during the period from about 290 Ma to ca. 25 Ma, and was finally tilted and exhumed during the Apenninic orogeny (Schenk, 1990).

Seismic reflection-refraction measurem (Lüschen *et al.*, 1992) have shown that Hercynian lower crustal section is underlair a zone of low S- and P-wave velocities, wl correlates with a mylonitic gneiss zone cropping to the N (fig. 1). The structural s of the rocks of the former lower crus preserved in most parts (Kruhl and Huntema 1991). Locally, during cooling af the Hercynian granulite-facies peak metamorphism, shear zones developed, mai within the metapelite unit, with a tectonic t to-the-NW transport. Shearing occurred retrograde amphibolite to sub-greensch facies conditions (Kruhl, 1992).

In the present study, we focus on a single

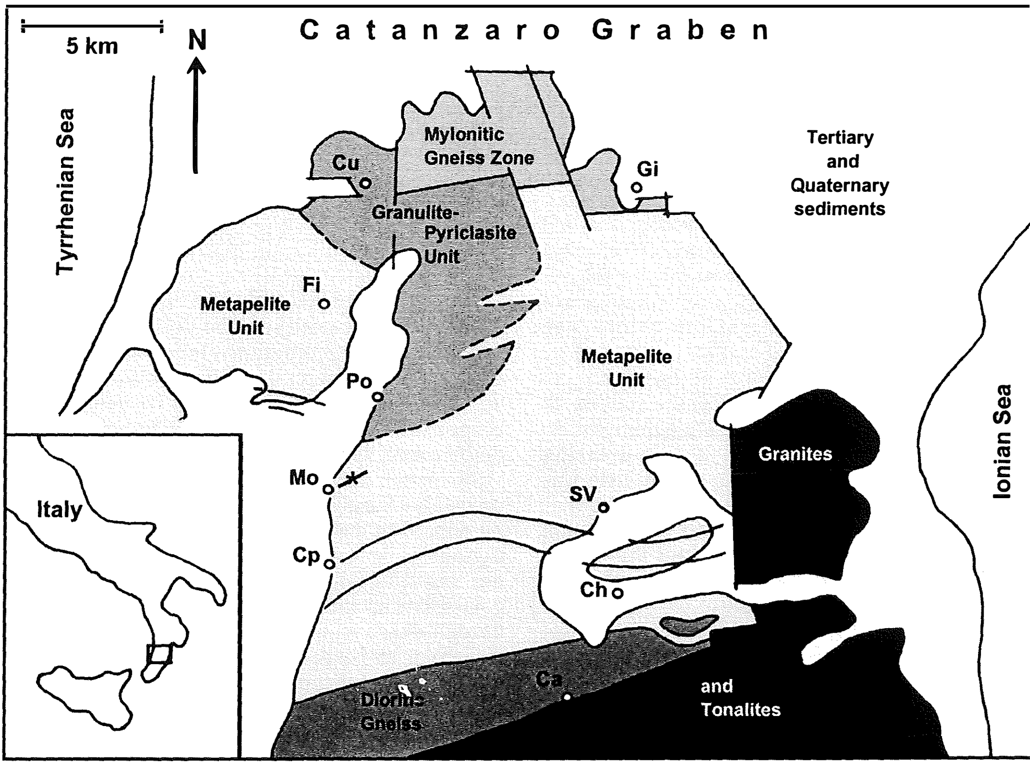

Fig. 1 – Schematic geological map of Hercynian lower crustal section in southern Calabria/Italy, after Amodio-Morell *al.* (1976), Schenk (1985), Appel (1990) and Kruhl (unpubl. data). ==== «univariant zone» (garnet-cordierite-bioti

cm thick shear zone in the central part of the lower crustal section (fig. 1). It is located in a coarse-grained (mm-size) metapelite with quartz, K-feldspar, plagioclase, garnet, biotite, and sillimanite and is oriented sub-parallel to the layering (fig. 2). Shearing is concentrated in a quartz-K-feldspar-rich zone and is partitioned into 50-500 μm thin fine-grained mylonitic layers. SCHENK (1990) proved the common P-T-t history of all parts of the lower crustal section, with generally decreasing temperatures and pressures from former lower to higher parts of the section, i.e. nowadays — due to late tilting — from the northern to the southern boundary of the exposed area. Therefore, the same P-T-t history is assumed for those rocks containing the shear zone and, related to their position within the lower crust, the P-T path of the shear zone can be figured. However, the question as of which P-T conditions obtained while the shear zone was active has to be answered on the basis of additional thermobarometric and miocrostructural analyses of the shear zone. The answer w serve to exemplify the conditions and timing retrograde high-T shearing in the Calabri Hercynian lower crust.

MICROSTRUCTURES

Quartz

Outside the shear zone, quartz is coars grained. It shows c-axis maxima near t preferentially NNE-SSW oriented elongati direction of the granulite facies deformati (figs. 2 and 3A) and a pattern of prismatic a basal subgrain boundaries («chessboar pattern) typical of rocks which underwe deformation in the high-quartz field (Kru 1996). In the shear zone, quartz is dynamica recrystallized (fig. 4A), with average gra sizes of 10-50 μm. The flat faces of t recrystallized grains are commonly oblique the mylonitic foliation. The c-axes of the oblique grains plot as an incomplete gird

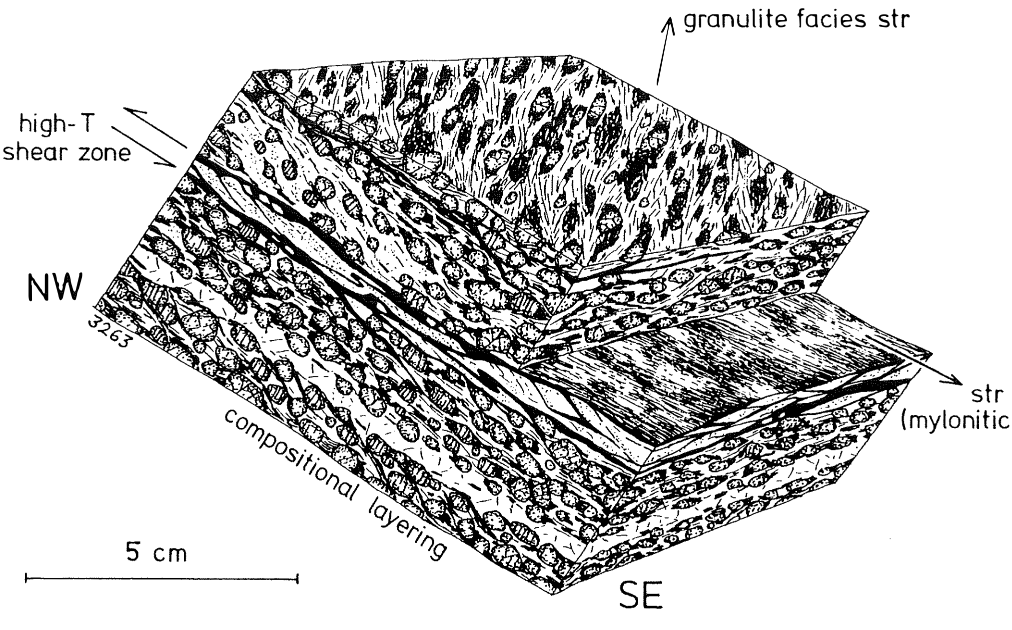

Fig. 2 – Sketch of shear-zone sample (3263) in its original SE-dipping position. Compositional layering consists of garn

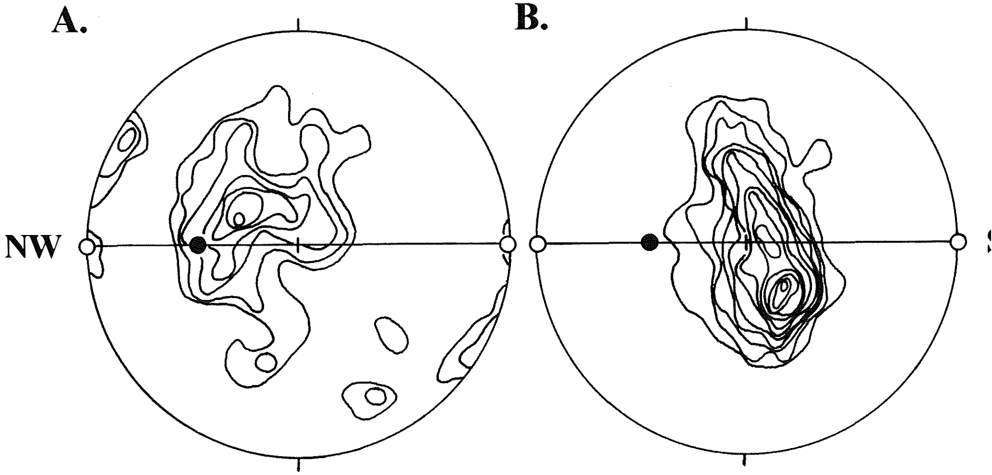

Fig. 3 – Quarz c-axis orientations. 300 measurements per diagram; contour intervals in times uniform distribution, s with 2; counting circle 1% of hemisphere; foliation = solid line; stretching lineation = open circles; granulite lineation of wall rocks = filled circle. **A.** Coarse crystals from granulite adjacent the shear zone. **B.** Dynam recrystallized quartz grains in shear zone, from a domain where flat faces of quartz are oblique to mylonitic fol similar to domain shown in fig. 4A.

around the mylonitic foliation plane and are perpendicular to the mylonitic stretching direction (fig. 3B), indicating dominant prism-<a> glide. Within regions of extreme flattening, quartz is kinked into thin ribbons without recrystallization. In low-strain regions near coarse garnet and feldspar clasts, relics of coarse quartz grains still show «chessboard» subgrain patterns (fig. 4B).

K-feldspar

Coarse K-feldspars (host grains) show wavy extinction and, along fractures and grain margins, are recrystallized and polygonized to smaller grains and subgrains of variable size. Recrystallization is partly dynamic and partly static. The size distribution of the recrystallized grains is bimodal. Those grains which developed along the margins of the K-feldspar hosts against fine-grained mylonitic quartz layers, are predominantly in the range of 5-10 μm and show transitions to subgrains (fig. 5). Grains between the K-feldspar hosts are in the range of 50-100 μm (fig. 4C). They bear albite

aggregates with predominantly equilib angles at triple points. K-feldspar is fo between the fragments of coarse prism sillimanite dismembered during shearing. feldspar hosts from the shear zone and fro host rocks and in the large recrystallize feldspar grains, perthitic exsolution developed (figs. 4A, C, D and 5). Ther three different types of exsolutions: large 50 μm diameter), medium (about 10 μm) small (2-3 μm). The large and me exsolutions occur in the large K-feldspa the shear zone and of the wall rock. The exsolutions, however, do not occur in th feldspars of the wall rock but only in the zone, where they are predominantly prese the boundaries of the larger exsolution recrystallized K-feldspars, only s exsolution blebs are found. Within the arr these recrystallized grains, there are lo small albite grains (fig. 4C). All this indi that (*i*) the small exsolutions developed d shear zone activity, (*ii*) were partly induce strain concentration around the

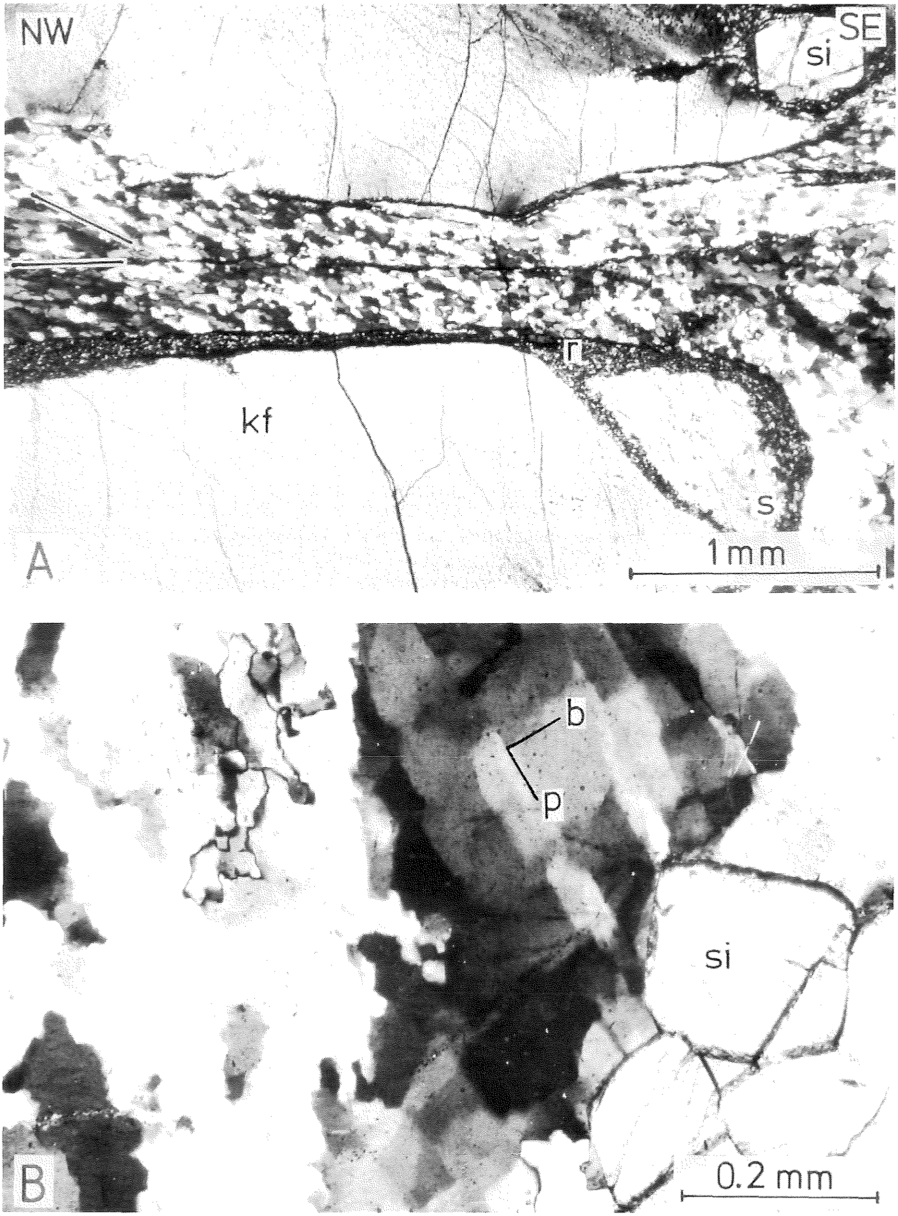

Fig. 4 – Thin section photographs of shear zone; sample 3263. **A.** Dynamically recrystallized quartz ribbons between coa K-feldspars (kf). A «S-C» texture (bars) is figured by long axes of quartz grains and thin layers of very small K-felds grains. K-feldspar shows wavy extinction, albite exsolutions (dots), and rims of recrystallized grains (r) and subgrains (s similar size. Recrystallization is enhanced around coarse sillimanite grains (si). Crossed polarizers (X pol). **B.** Margir

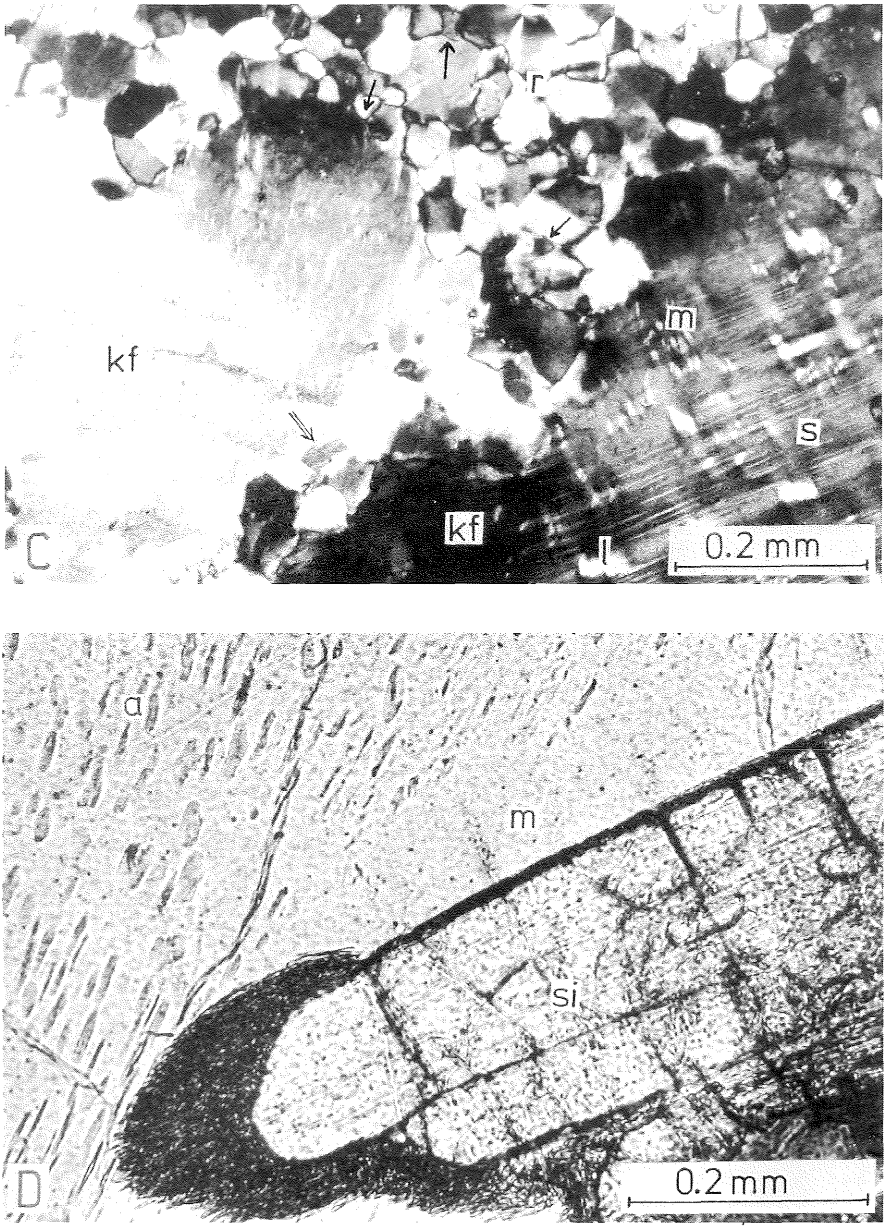

Fig. 4 – **C.** Coarse K-feldspar crystals with statically recrystallized grains (r) along their margins. Old grains exhibit types of exsolutions with different crystallographic orientations: large (l), medium (m), and small (s). Within recrysta grains, only small exsolutions (double arrow) are developed. Small albite grains (arrows) within region of recrystalliz

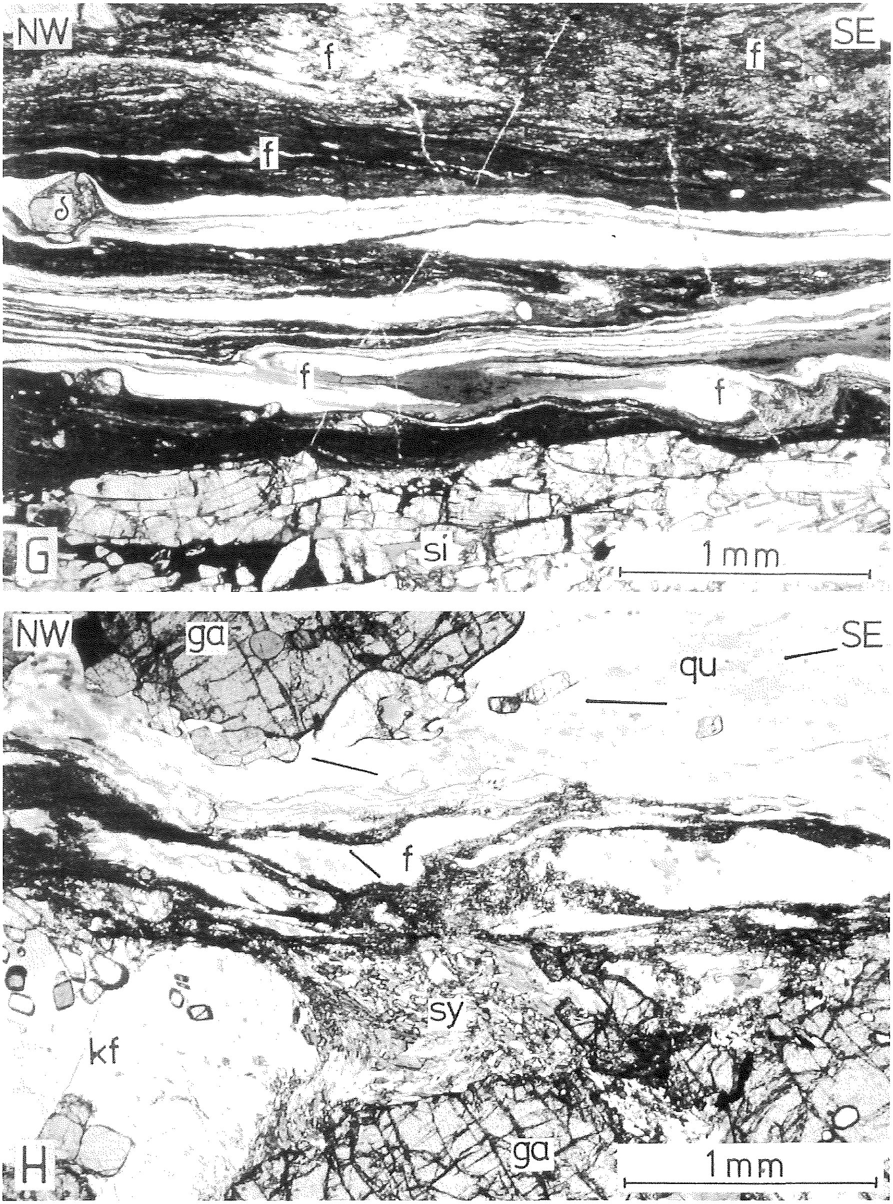

Fig. 4 – **G.** Mylonitic layering (light: recrystallized quartz grains; dark: predominantly opaque minerals, sillimanit biotite, optically determined), locally folded (f) in sense of shear (top-to-the-NW). Sillimanite forms delta clas indicating same sense of shear. Lower part of photograph occupied by coarse sillimanite (si) of wall rock; ppl. **H.** C garnet (ga) (with quartz and biotite inclusions) and K-feldspar (kf) (with sillimanite inclusions) of granulite facies Garnet locally altered to sillimanite-biotite-quartz symplectite (sy) which — in sense of shearing (top-to-the-NW)

present large and medium exsolutions was reorganized and concentrated among the recrystallized K-feldspar grains, and (*iv*) the large and medium exsolutions formed prior to the small ones. However, the margins of the coarse K-feldspars and around inclusions do not show exsolution (fig. 4D), possibly due to original compositional zoning. Microcline twinning frequently occurs in regions of enhanced deformation. Coarse feldspar crystals are commonly surrounded by 5 to 500 μm wide mylonitic layers composed of predominantly K-feldspar or an optically well determinable sillimanite-biotite-quartz mixture. In ribbons of fine-grained dynamically recrystallized quartz, feldspar may be fractured and rotated in the sense of shearing.

Plagioclase

The few plagioclase crystals exhibit the same deformation textures as the K-feldspar, i.e. fractures, wavy extinction, subgrains, recrystallized grains.

Sillimanite and kyanite

In the host rock, sillimanite forms coarse euhedral crystals (fig. 4G). In the shear zone, they are either rotated with the long axes commonly parallel, but rarely perpendicular to the stretching direction, or are bent, fractured and dismembered parallel to the stretching direction (fig. 4E). Locally, the coarse crystals are fractured and rotated in the sense of shearing (fig. 4F). Between the fragments new biotite and K-feldspar are formed. In the pressure shadows of coarse crystals, newly grown fibrolitic sillimanite is oriented with the long axis parallel to the mylonitic stretching direction (fig. 4D). It overgrows the exsolution-free margins of K-feldspar. The ultramylonitic layers are formed of a sillimanite + feldspar or a sillimanite + biotite + quartz ± opaque mixture with grain-sizes ≤ 5 μm (fig. 4F, G). Despite the small grain sizes, the minerals can and grain shapes. In these layers, prismati[c] sillimanite is polygonized, dynamicall[y] recrystallized and, at rare sites, inverted t[o] kyanite along the margins (as shown b[y] electron plus light microscopy). Th[e] ultramylonitic layers are refolded in the sens[e] of shearing (fig. 4G). Apart from the inversio[n] to kyanite, sillimanite does not show any othe[r] signs of retrograde alteration, even in th[e] finest-grained mylonitic layers. Only in ver[y] few places is sillimanite replaced by whit[e] mica.

Garnet

The garnet grain-size is the same (up to [1] mm) both inside and outside the shear zone. [It] is marginally replaced by quartz-biotite sillimanite symplectite which in turn i[s] deformed along mylonitic layers (fig. 4H). Th[e] garnet crystals are fractured and dismembere[d]. The fractures are oriented predominantl[y] perpendicular to the stretching direction of th[e] shear zone. In the fractures pale biotite and K-feldspar are formed. Along the shear plane coarse biotite exsolved rutile and is kinked.

White mica

White mica is rare. It occurs as the fine-grained alteration product of sillimanite o[r] locally overgrows the mylonitic foliation as u[p] to 100 μm large crystals only show wea[k] deformation.

KINEMATICS OF SHEARING

The principal strain axes in the shear zon[e] clearly differ from those outside. In the hos[t] rock, coarse prismatic sillimanite and th[e] pressure shadows of garnet are oriented NNE-SSW, parallel to the main elongation directio[n] of the high-temperature deformatio[n] characteristic of the Hercynian lower crusta[l] section of Calabria (Kruhl and Huntemann 1991). However, within the shear zone, th[e]

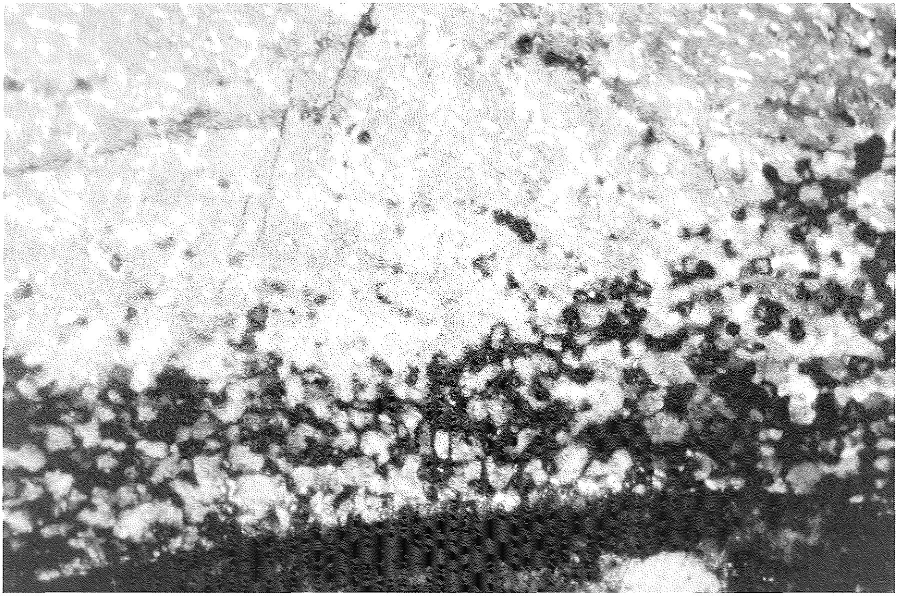

Fig. 5 – Microphotograph of deformed rim of a K-feldspar phenocryst from shear zone centre (sample 3263, sect
Core of K-feldspar host (bottom of photograph) with albite exsolutions is surrounded by rim of recrystallized pol
grains (top). Change from host to recrystallized grains is gradual through subgrains. Long side of photograph = 400 μ

2). Prismatic sillimanite is also generally aligned in this direction (fig. 4E) which coincides with the main elongation direction of the retrograde amphibolite to sub-greenschist facies deformation in the lower crustal section (Kruhl, 1992).

S-C textures are developed in fine-grained layers of dynamically recrystallized quartz ± sillimanite and K-feldspar (fig. 4A). Coarse feldspars and sillimanite form sigma- and delta-clasts. Locally, the mylonitic foliation is tightly folded, with the flat faces of the dynamically recrystallized quartz grains parallel to the fold axial plane (fig. 4H). Quartz c-axis orientations form a girdle oblique to the mylonitic foliation and approximately perpendicular to the mylonitic SE-NW oriented stretching direction (fig. 3B). All these textures are consistent with the interpretation that the mylonitic stretching direction is the direction of local tectonic transport. Moreover, S-C textures, sigma- and delta-clasts, monoclinal

stretching direction – a top-to-the-NW ser
shear, in agreement with the general sen
retrograde shearing in the lower crustal se
(Kruhl, 1992).

WATER

Presence of water (henceforth used
collective term for undifferentiated typ
water, see e.g. PATERSON, 1989) du
shearing may be inferred (*i*) by d
measurement of H_2O contents of nomi
H_2O-free minerals and (*ii*) by the occurren
OH-bearing minerals related to the shearin

(*i*) The water contents of host
recrystallized K-feldspars within the shear
were qualitatively determined by micro i
red spectroscopic measurements (fig
Feldspars show absorption bands at 3
3620, and 3383 cm^{-1}, indicating H-bondec
and OH-stretching vibrations (e.g. Bane

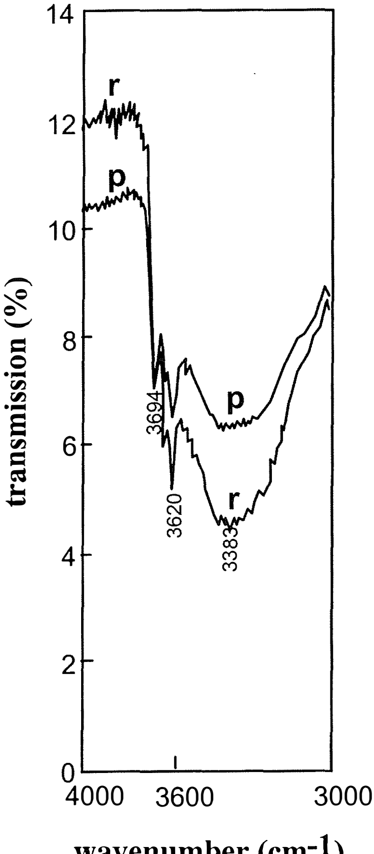

Fig. 6 – Two infra-red transmission spectra of undeformed K-feldspar crystals (p) and recrystallized grains (r) (not a single grain but a group of grains); resolution approx. 20 μm.

recrystallized feldspar grains. Moreover, the relatively weak absorption indicates that the water contents were very low.

On the basis of these results, it may be argued that the very low water contents of the K-feldspars, which can be interpreted as a consequence of granulite facies metamorphism, did not increase during shear-zone formation and recrystallization. However, it cannot be totally excluded that the water contents of the recrystallized grains were slightly higher during

(*ii*) The rare formation of symplectit biotite, together with quartz and sillimanite, c of garnet, occurred exclusively before or duri shearing, suggesting that locally at least son water was present. The rare white mica, up 100 μm in diameter, which overgrows the fin grained mylonitic layers, may be interpreted a result of post-mylonitic hydration. Only very few fine-grained and a few microns wi mylonitic layers did sillimanite partly rea with white mica. In general, even in the layers and in contact to K-feldspar, silliman is not altered (fig. 4H). Consequently, duri shearing even in the fine-grained mylonit layers which should represent good diffusi pathways, no water was present. Otherwis sillimanite and K-feldspar would have be changed to quartz + white mica during the conditions of shearing (see next section).

TEMPERATURE-PRESSURE CONDITIONS DURING SHEARING

On the basis of the P-T evolution of t Calabrian lower crustal section, reported Schenk (1990), and considering the position the shear zone in the metapelite unit (fig. the P-T path of the shear zone can established (fig. 7). The structural a petrological changes in it were induced shearing and reflect the P-T conditions duri shearing. These conditions may be estimat on the basis of microstructures, newly form minerals, feldspar thermometry, Fe-M exchange thermometry of garnet and biot and garnet – Al_2SiO_5 – plagioclase-quar barometry. Estimates vary considerat according to chosen method and mine location in the shear zone. Nevertheless, th are in general agreement with the P conditions derived from the P-T evolution pa given by Schenk (1990). This shows th despite the problems of exchange-reactio based thermobarometry in shear zones (e Altenberger, 1995c), in the present case,

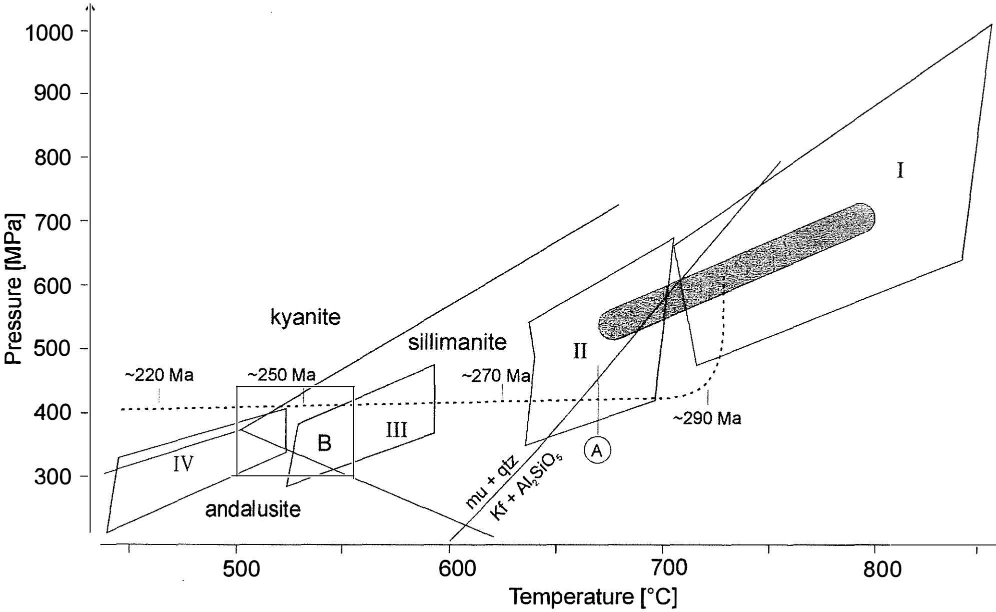

Fig. 7 – P-T development of studied shear zone. Shaded area: P-T peak conditions of Hercynian lower crustal sectio Calabria (after Schenk, 1990); dotted line and numbers: P-T path of shear zone and ages (in Ma), inferred from P-T and mineral cooling ages given by Schenk (1990). (A) Upper temperature limit of albite exsolutions in K-feldspar. Field of two-feldspar thermometry of recrystallized grains. Fields I to IV: P-T ranges estimated by garnet-bi thermometry and garnet-plagioclase barometry. White mica + quartz breakdown reaction after Spear and Cheney (19 Al_2SiO_5 polymorph fields after Holdaway and Mukhopadhyay (1993).

which give a reliable picture of P-T conditions during movement of the shear zone.

Feldspars

On the basis of chemical compositions of the large host grains, combined with the total volumes of the three exsolution generations, the K-feldspar compositions before unmixing are calculated. Using the experimental data and calculated solvi at 500 MPa from Luth *et al.* (1974), a temperature of 650-670°C may be inferred for the beginning of unmixing.

Within the mylonite, K-feldspar and the rare plagioclase recrystallize along microfractures and grain margins, showing wavy extinction and subgrain structures. The dynamic recrystallization of K-feldspar and plagioclase to new grains with An > 15 mol% indicates minimum temperatures of ca. 500°C during metamorphism plagioclase starts to recrystal near the oligoclase boundary, i.e. around 5 530°C – a temperature which increases sligl with increasing pressure – and K-feldspa somewhat lower temperatures of 490-500 However, this recrystallization temperature been calibrated for «normal» strain rates and prograde metamorphism with approximat $P_{H_2O} = P_{total}$. Although experiments indicate t in dry conditions, recrystallization starts higher temperatures (e.g. Paterson, 1989), th are no data available from natural syster Deformation experiments (Tullis *et al.,* 19 and comparisons with natural conditions sh that an increased strain rate increases recrystallization temperature (Kruhl, 199 Therefore, a recrystallization temperature feldspars above 500°C, due to increased str rates during shearing, can not be tota occurs, causing the formation of microcline twins due to ordering of Al, which lowers feldspar symmetry. Similar features are described by Reischmann *et al.* (1990) from an amphibolite facies shear zone of comparable composition. The chemical composition of adjoining dynamically recrystallized K-feldspar and plagioclase in mylonitic layers (Table 1) was used for the ternary two-feldspar thermometer of Fuhrman and Lindsley (1988). Pressures during feldspar recrystallization may be estimated as approximately 300-450 MPa (by GASP barometry; see below). On the basis of such P-conditions, temperatures between 445°C and ca. 555°C with average temperatures of 518°C at 300 MPa and 521°C at 400 MPa are estimated. However, a recrystallization temperature for feldspar of approx. 500°C is the lower temperature limit (fig. 7, field B).

Al_2SiO_5 polymorphs

The formation of fibrolitic sillimanite in the pressure shadows of coarse prismatic sillimanite (fig. 4D) and at the expense of K-feldspar porphyroclasts, as well as the crystallization of the quartz-biotite-sillimanite symplectites in the mylonite, indicate that shearing was active at least partly within the stability field of sillimanite. Additionally, some prismatic sillimanite in the mylonitic layers is dynamically recrystallized along subgrain boundaries. In addition, in these grains kyanite grew along the margins, indicating that the P-T path cuts the sillimanite-kyanite boundary, however, not necessarily during shearing. Moreover, the presence of kyanite and the absence of andalusite probably indicates that the P-T path does not run through the andalusite field, in accordance with the P-T path of the shear zone inferred from the lower crust P-T evolution given by Schenk (1990) (fig. 7).

Garnet-biotite Fe-Mg exchange thermometry and garnet-Al_2SiO_5-plagioclase-quartz (GASP) barometry

significantly different temperatur
Temperature determinations are based on t
thermometers of Thompson (1976), Goldm
and Albee (1977), Holdaway and Lee (197
Ferry and Spear (1978), Ganguly (197
Perchuk *et al.* (1985), Hoinkes (1986) a
Bhattacharya (1992). Pressure is estimated
the GASP barometer (Koziol and Newto
1988), applying the composition of coexisti
garnet and plagioclase (Table 1). Th
barometer is based on the net transfer reacti
anorthite = grossular + kyanite + quar
Activity/composition relations are given
Aranovich and Podlesskii (1980), Hodges a
Spear (1982) and Spear (1993). Four coexisti
garnet-biotite-plagioclase assemblages we
analysed, which correspond to differe
structural stages (Table 1, fig. 7):

Stage I. Cores of coarse garnet, plagiocla
and biotite in the undeformed protolith indica
P-T conditions before shearing;

Stage II. Rims of coarse garnet, plagiocla
and biotite in the protolith adjacent to the sh
zone, which indicate a step of the cooling pa
and probably also the beginning of sheari
(see further below).

Stage III. The fine-grained, syntectonica
recrystallized biotite-plagioclase assemblage
the mylonitic layers and the rims of coa
garnet in contact with them; specifically, bio
is oriented in the mylonitic foliation, indicat
syntectonic formation.

Stage IV. Rims of syntectonica
dismembered garnets and biotite form
between fragments.

The validity of applying mineral equilib
thermobarometers in high-strain zones in
conditions is still under discussion. Howev
the fact that the studied shear zone does
show a significant increase in (OH)-bear
minerals does not rule out the fact that (O
bearing fluids were active during some sta
of shearing. The partial breakdown of biot
which was in part preserved during the wh
shearing process may be a source for suffici
(OH). It was also possible that other non-(O

TABLE 1

Mineral analyses of major phases of shear zone sample (3263).

T I			T II			T III			T IV			T_{fsp}	
biotite	garnet	plag.	biotite	garnet	plag.	biotite	garnet	plag.	biotite	garnet	plag.	K-feld	plag.
38.84	38.27	63.03	39.43	38.59	63.18	39.43	38.42	65.02	42.85	38.16	63.23	64.93	62.45
2.5	0.02		5.08	0.04		2.77	0.01		3.76	0.03			
19.84	21.66	23.65	17.68	21.54	24.04	18.13	21.46	21.83	18.35	21.51	23.47	18.96	23.27
10.34	31.11	0.34	12.3	30.93	u.d.l.	8.57	31.26	0.08	10.4	32.79	0.19	0.01	0.14
u.d.l.	0.8		u.d.l.	0.8		0.05	0.7		u.d.l.	0.92			
13.91	7.33		12.34	7.34		16.86	6.98		11.48	5.59			
	1.09	4.57		1.02	4.81		1.07	3.45		1.3	4.90	0.05	4.72
0.09		8.90	0.06		8.60	0.04		8.73	0.2		8.57	0.88	8.77
9.44		0.10	9.64		0.14	9.63		0.31	10.37		0.10	15.85	0.15
0.14			0.08		0.03	0.01			0.34		0.13	0.1	0.04
0.34			0.26			0.9			0.34				
95.30	100.21	100.65	95.72	100.23	100.79	96.07	99.98	99.44	97.07	100.03	100.60	100.81	99.40
2.8095	2.9883	2.7820	2.8410	3.0084	2.7697	2.8225	3.0107	2.8712	3.0136	3.0425	2.788	2.946	2.7777
0.1367	0.0011		0.2750	0.0022		0.1490	0.0006	0.1988	0.0018				
1.0538	0.0106	1.2186	0.8840	0.0000	1.2422	1.0285	0.0000	1.1364	0.7876		1.2140	1.014	1.2198
0.4199	1.9832		0.6161	1.9785		0.5008	1.9819		0.7732	1.9964			
0.0000	0.0000	0.0125	0.0000			0.0000			0.0000	0.0000	0.0069		0.0052
1.0033	2.0317		0.7413	2.0161		0.5129	2.0481	0.0028	0.6118	2.1590			
0.0073	0.0526		0.0002	0.0527		0.0032	0.0465		0.0005	0.0613			
1.4170	0.8536		1.3751	0.8527		1.7995	0.8157		1.2032	0.6562			
0.0000	0.0916	0.2141	0.0000	0.0852	0.2261	0.0047	0.0903	0.1631	0.0000	0.1100	0.2297	0.428	0.2251
0.0099	0.0000	0.7543	0.0083	0.0000	0.7313	0.0505	0.0050	0.7473	0.0270	0.0060	0.7271	0.077	0.7561
0.9553	0.0000	0.0056	0.8859	0.0000	0.78	0.8798	0.0011	0.176	1.2032	0.0050	0.0054	0.918	0.0080
	67.08			67.08			68.27			72.32			
	28.18			28.37			27.19			21.98			
	1.74			1.75			1.55			2.05			
	2.97			2.73			2.98			3.6			
	0.03			0.06			0.02			0.05			
		77.44			80.53			80.53			75.57	9.8	76.17
		21.98			17.58			17.58			23.87	0.3	22.91
		0.58			1.89			1.89			0.56	89.9	0.82

of the minerals show intensive element redistribution is given by the chemistry of the analysed minerals. Moreover, the small mineral grain size in the shear zone would enhance diffusivity and, therefore, element exchange and equilibration.

EVOLUTION OF SHEAR ZONE

Considering the above petrological and fabric evidences a P-T path of the shear zone can be constructed (fig. 7). The thermal peak of the host granulite-facies rock is given by the coarse-grained assemblage K-feldspar-garnet-sillimanite-plagioclase-biotite-rutile. Thermobarometric calculations (fig. 7, field I) give P-T conditions in the range of 850°C/1 GPa and 715°C/480 MPa with mean values of about 780°C/740 MPa. These data fit the P-T estimates of Schenk (1985, 1990) who gives peak metamorphic conditions for the Calabrian lower crust in the range 670°C/530 MPa and 800°C/740 MPa.

The retrograde evolution of granulite is shown by several stages of fabric and mineralogical changes within the shear zone. Its development began with coarse-grained recrystallization of granulite-facies K-feldspar, now preserved only in the host rock at the margin of the shear zone and between porphyroclasts in it. Coarse albite exsolutions in granulite-facies K-feldspars, which do not occur in the recrystallized grains, indicate unmixing temperatures of 650-670°C and, therefore, recrystallization below this temperature. This early deformation, which led to coarse-grained recrystallization, probably also induced the chemical changes in the mineral rims of the granulite adjacent to the shear zone, as documented in field II of fig. 7. Consequently, shearing started after peak temperature conditions, probably after isothermal decompression (Schenk, 1990) and during isobaric cooling. This is in agreement with the pre-shearing formation of the biotite-

1984). The stability of the assemblage feldspar + sillimanite instead of the low grade assemblage muscovite + quartz requi absence of water during this high-temperat shearing stage, which occurred outside the feldspar-Al_2SiO_5-stability field. After t recrystallization in the early high-temperat stage of shearing, annealing possibly occurr as suggested by granoblastic polygo microstructures in recrystallized K-felds grains.

After the annealing phase, sheari continued in lower P-T conditions. The grov of fibrolitic sillimanite, which formed in (*i*) pressure shadows of prismatic sillimanite, coarse recrystallized K-feldspar grains, and (exsolution-free margins of K-feldspa indicates that shearing was active within stability field of sillimanite. The myloni layers show evidence for further dynam recrystallization of K-feldspar and plagiocl and are additionally characterized by stability of garnet, biotite and sillimani Thermobarometric calculations, based on stability of sillimanite, garnet, biotite a plagioclase, indicate P-T conditions of ab 560°C and 350 MPa (fig. 7, field III).

Lower P-T conditions of shearing suggested by the dynamic recrystallization sillimanite and its inversion to kyanite in rims. Formation of kyanite in an earlier st would have resulted in an exotic P-T pa considering the well-constrained fields I to Fig. 7 indicates that the P-T path could ha cut the sillimanite-kyanite boundary above triple point. The results of felds thermometry on small recrystallized feldsp in the mylonitic layers overlap field III and sillimanite-kyanite transition in low-press conditions, but show a somewhat lower m temperature of about 520°C (fig. 7). T preferred orientation of small white mi formed at the expense of recrystalliz feldspars and the new growth of pale bio between the fragments of dismembered garn indicate excess water and equilibration in

DISCUSSION AND GEOLOGICAL CONSEQUENCES

The P-T stages of the shear zone are in good agreement with the P-T path constructed by Schenk (1990) for the Hercynian lower crust exposed in the nappe pile of Calabria (fig. 7). This argues for the validity of the applied thermobarometers even in a «dry» shear zone. This may be due to: (1) the small mineral grain-size and consequently high density of grain boundaries in the shear zone, which support volume as well as boundary diffusivity; (2) the possibly long period of activity of the shear zone (see further below) may have improved the process and favoured completion of ion-exchange reactions.

The development of the shear zone started after isothermal decompression of the Hercynian lower crust of Calabria, i.e. subsequent to its first uplift to a mid-crustal level, according to the P-T history given by Schenk (1990). The host rocks of the mylonite zone were locally hydrated prior to shearing, as indicated by the biotite-sillimanite-quartz symplectites. However, shearing started in nearly dry conditions. Water activity only increased during the late stage of shearing, as indicated by the formation of biotite and muscovite in some thin mylonitic layers. Schenk (1985, 1990) suggested that local hydration, which is mainly observed in the upper metapelite unit, is the result of water released from cordierite during early isothermal decompression. As indicated by the very limited hydration, such a small water reservoir was soon consumed. Instead, there is no evidence for a fluid phase rich in CO_2 or with a different composition. Neither is there any substantial growth of new minerals or significant differences in water contents between host and recrystallized grains of K-feldspars.

Most high-temperature shear zones reported from other regions describe increased fluid infiltration with retrogressive mineral reactions

Also, most shear zones in the lower c appear to be fluid pathways during shea (Brodie and Rutter, 1987; Newton, 1 Altenberger, 1991, 1995a,b). However, studied shear zone, which was active at a crustal level, was obviously not in connec with a fluid reservoir. Water started infiltrate the host rocks at the decline shearing, i.e., in uppermost greenschist fa conditions.

Shear zone formation started earlies temperatures and pressures of ca. 700°C/ MPa and declined in P-T conditions lower ca. 520°C/380 MPa, i.e. shearing w continuously or discontinuously, active u amphibolite to uppermost greenschist fa conditions. Relics of textural equilibration, equilibrium interfacial angles or undefor sillimanite-biotite-quartz symplectites aro garnet, are preserved in the high-tempera part of the shear zone. This observation and gap between P-T-fields II and III (fig suggest at least two stages of deformation high and low temperature, with a br between them. On the basis of the P-T-t given for the Hercynian lower crustal sectio Calabria (Schenk, 1990; fig. 8), shea started at about 280 Ma and declined at a 220 Ma. As given by regional-scale tecto (Kruhl, 1992), during this time approxima horizontal top-to-the-NE tectonic trans occurred along this cm-thin shear zone.

Activation of narrow shear zones over a period of time, in addition to relatively conditions during shearing, indicate that, the first formation of a shear zone, the loca of movements was not necessarily governe the infiltration of fluids, but by ot parameters, such as grain-size reduction shape and preferred crystallogra orientations, i.e. the presence of textur weak zones. Moreover, even at mid-cru levels the former lower crustal sectio Calabria did not generally behave ductilely as a rigid block, elsewhere unable

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