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The peridotites of the Kukës ophiolite (Albania): structure and kinematics

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Abstract

Structural study of the Kukës ultramafic massif which belongs to the Albanian Mirdita zone is of great interest for reconstruction of Alpine tectonic environments. From bottom to top, the following lithological succession is observed in the ultramafic-mafic section: harzburgites, harzburgites with dunitic lenses, dunites, an intermediate zone and gabbros. The ultramafic rocks are apparently weakly deformed. The textures indicate a high-temperature deformation and slight locally superimposed low-temperature deformation. This is confirmed by olivine lattice-preferred orientation which indicates that (010)[100] was the activated glide system in olivine (i.e., at high temperature). The kinematics of deformation have been determined from the lattice-preferred orientation of olivine; an inversion of the shear sense occurs at the base of the dunite zone and is interpreted as being due to asthenospheric flow. A simple geometric reconstruction indicates that the Kukës ultramafic massif could correspond to the eastern flank of a NNW-SSE-orientated paleo-ridge.

1. Introduction

The Mirdita zone (Albania) corresponds to a large alpine ophiolitic complex in which two alignments are recognized and distinguished on the basis of petrology of the ultramafic section (Shallo et al., 1985). The eastern belt belongs to the harzburgite ophiolite type (HOT, Rocci et al., 1975; Boudier and Nicolas, 1985) and contains several chromite and platinum group mineral deposits. The western belt is of the lherzolite ophiolite type (LOT, Rocci et al., 1975; Boudier and Nicolas, 1985). Both belts are in continuity under the central synclinorium (Frashëri et al., 1991) which is filled by thick volcanics and post-Jurassic sediments (Fig. 1). Thus, the Mirdita zone is the continuation of the two ophiolitic belts recognized by Pamic (1983) to the north of the Shkodër–Pejë (Dinarides) and to the south of Albania (Hellenides) and is a unique case in which two ophiolite types are juxtaposed in the same belt.

To date, a very good petrological and petrographical database has been available on the two ophiolitic belts of the Mirdita zone (Shallo et al., 1985). However, there have been no structural nor kinematic studies even though the eastern belt is of great economic interest for exploration and exploitation of chromite deposits. The geometry of the mineralized bodies, including the alignment and elongation of chromite lenses, cannot be understood without a structural study of the deformed surrounding peridotites: the determination of the plane and direction

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of high-temperature flow allows the alignment of mineralized bodies to be determined and thus provides some guide for exploration and exploitation of chromite deposits (Cassard, 1980). Moreover, such a structural study may shed some light on the kinematics of mantle deformation before the emplacement of the ophiolites onto the continent by comparison with what is known in Oman and in other ophiolitic massifs (see Nicolas, 1989).

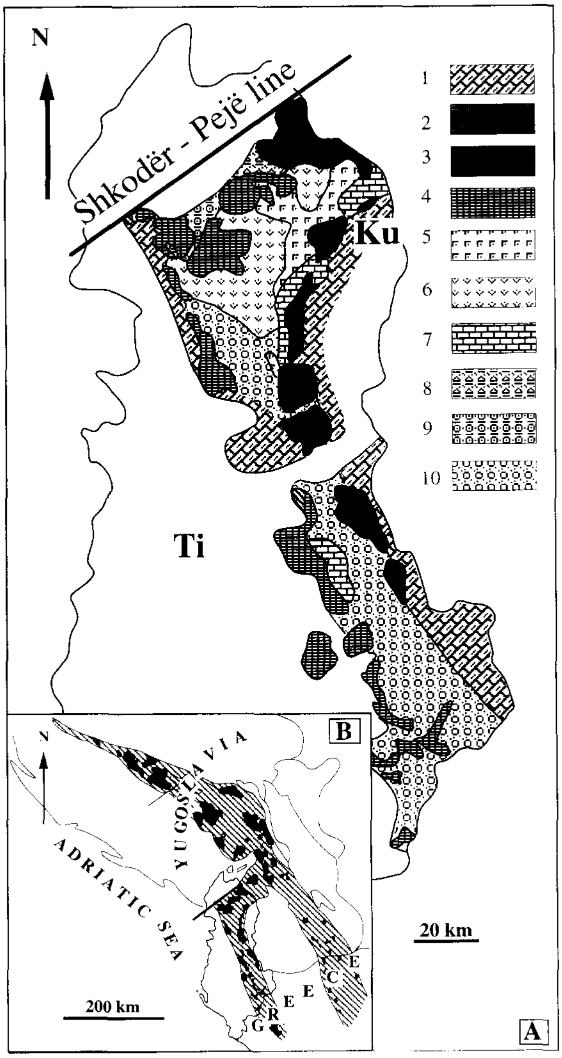

Fig. 1. (A) The ophiolitic complex of the Mirdita nappe (simplified from the Geological map of Albania, 1983). *1* = Upper Triassic–Lower Jurassic limestones; *2* = volcano-sedimentary complex; *3* = eastern belt (HOT = harzburgitic ophiolite type); *4* = western belt (LOT = lherzolitic ophiolite type); *5* = gabbro-norites; *6* = Volcanic rocks (basalts, spilites, keratophyre, dacites, andesites, etc...); *7* = Lower Cretaceous limestones; *8* = Tithonian–Lower Cretaceous breccia with ophiolitic elements; *9* = heterogenous ophiolitic melange; *10* = Paleogene–Quaternary sedimentary rocks (limestones, sandstones, flyshs and conglomerates). *Ti* = Tirana; *Ku* = Kukës. (B) Place of the Mirdita nappe in the Dinaro–Albano–Hellenide Alpine Belt.

This paper is concerned with the Kukës massif which belongs to the eastern belt of the Mirdita zone (Figs. 1 and 2). The Kukës massif contains all the components of an ophiolite (Anonymous, 1972). It shows a complete cross-section of oceanic lithosphere, from mantle rocks (6–7 km under the paleo-Moho) to mafic crust (gabbros and gabbro-norites, Fig. 3). This study will focus on structures of the ultramafics and mafics as mapped in the field and observed in thin section, and on the kinematics that may be deduced from lattice-preferred orientations of minerals such as olivine.

2. Geological environment of the Kukës massif

The Upper Triassic–Lower Jurassic limestones (Fig. 2) southeast of the Kukës ultramafic massif are composed of grey neritic limestones, locally containing stromatolites and fragments of megalodontes. The thickness of the limestones is on the order of 1 km. The degree of recrystallization increases upwards in the formation, i.e., northwestwards.

The volcano-sedimentary complex lies in stratigraphic continuity above Triassic–Jurassic limestones (Kodra et al., 1993). This complex, of Late Jurassic age, has a large extension along the eastern boundary of the Mirdita ophiolites and may be compared to the "diabase-chert" formation in the Dinarides (Ciric, 1984). The complex is comprised of green shales, which become black (organic matter) or reddish (iron oxides) towards the summit of the cross-section. Some 10-m-thick reddish pelagic limestones are locally observed. Volcanics are interlayered in these sedimentary rocks and are represented by effusive basic rocks, such as spilites, diabases and variolites, which have suffered different degrees of chloritization, epidotization, carbonatization and pyritization.

The metamorphic sole of the Kukës massif is observed locally between the ultramafics and the volcano-sedimentary complex with which it appears in continuity. It is constituted of two types of rocks: (1) Finely layered amphibolites showing a metamorphic foliation, a stretching lineation and a granolepidoblastic texture (Bard, 1990). The thickness of these amphibolites varies from 200 to 500 m. The metamorphic paragenesis is amphibole (tremolite-actinolite), epidote (clinozoïte) and albite. (2) Micaceous quartzites and micaschists with garnet developed during low-grade regional metamorphism (Turku, 1987). These rocks exhibit a lepidoblastic texture and their paragenesis is garnet, quartz, muscovite, biotite and plagioclase. Their thickness varies

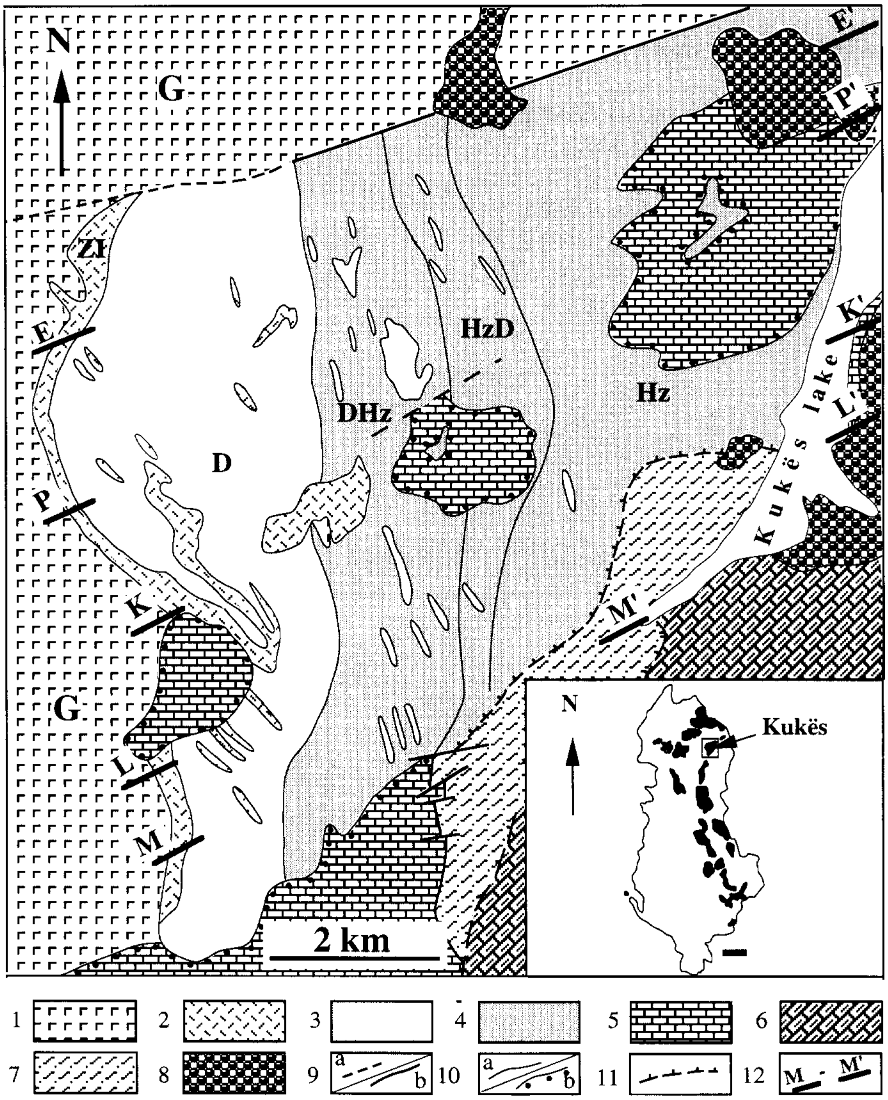

Fig. 2. Schematic geological map of the Kukës ultramafic massif. *1* = gabbro-norites; *2* = intermediate zone between the dunites and the gabbro-norites; *3* = dunites; *4* = harzburgites; *5* = Lower Cretaceous limestones; *6* = Upper Triassic–Lower Jurassic limestones; *7* = volcano-sedimentary complex and metamorphic sole; *8* = Alluvium; *9* = fault (*a* = supposed, *b* = observed); *10* = lithologic boundary (*a* = between different lithological facies, *b* = stratigraphic unconformity); *11* = thrust; *12* = cross-sections used to construct the logs in Fig. 3. Lithological zones: *Hz* = Harzburgites; *HzD* = harzburgites-dunites; *DHz* = dunites-harzburgites; *D* = dunites; *ZI* = intermediate zone; *G* = gabbros.

from 20 to 30 m, but may locally reach 100–150 m due to folding.

The northeastern and southern parts of the massif are covered by transgressive limestones of Early Cretaceous age. Therefore, the continental emplacement of the Kukës massif occurred at the Upper Jurassic–Lower Cretaceous boundary.

3. The ultramafic and mafic rocks

The Kukës ultramafic massif is subdivided into five N–S-elongated lithological zones which are from bottom to top or from east to west (Fig. 2): the harzburgite zone (Hz); the harzburgite-dunite zone (HzD); the massive dunite zone (D); the intermediate zone (Zi) constituted by intercalated pyroxenites, wehrlites and lherzolites; and the gabbros (G).

3.1. The harzburgite zone (Hz)

The contact between the metamorphic sole and the harzburgites is topographically easily visible but difficult to observe in outcrops. The average composition of the harzburgites is 80% olivine (forsterite), 17% orthopyroxene (enstatite), 3% clinopyroxene (diopside) and 2% spinel. The harzburgite zone presents some dunite lenses (less than 10% orthopyroxene, Streckheisen, 1973). Its thickness increases from south to north, where it reaches 2500 m (Fig. 3).

3.2. The harzburgite-dunite zone (HzD)

Depending on the relative amount of dunite and harzburgite, the harzburgite-dunite zone may be divided into two sub-zones: (1) the lower harzburgite-

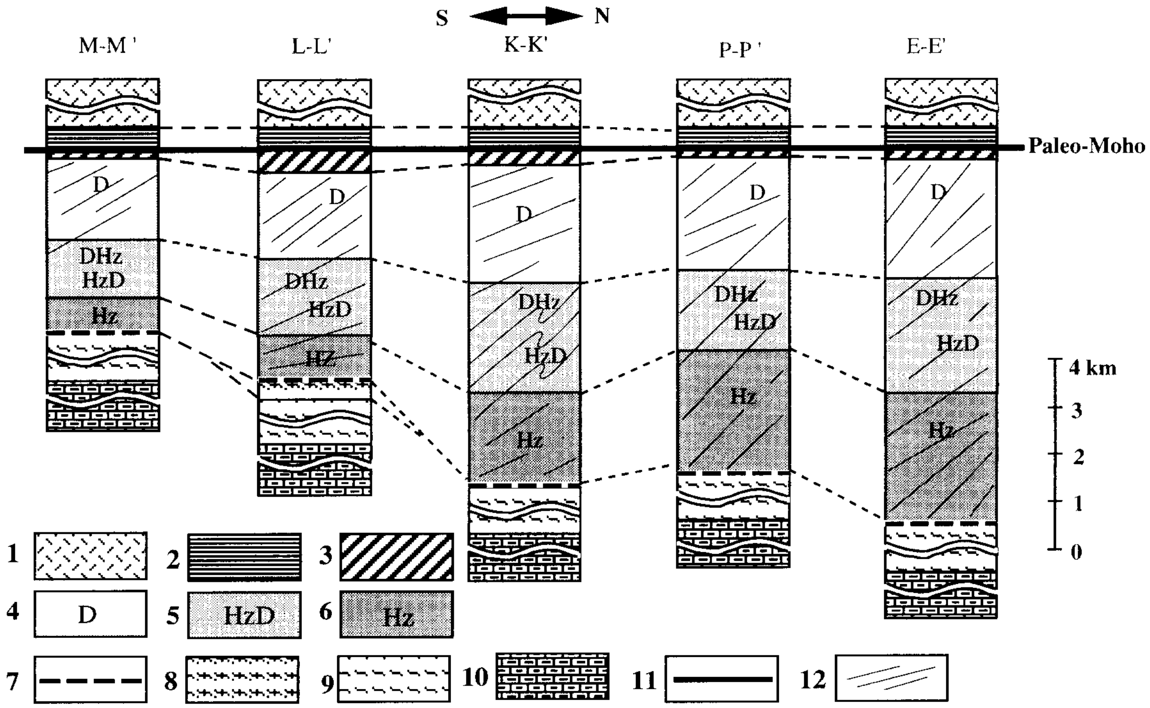

Fig. 3. Correlation between different lithological logs in the Kukës ultramafic massif. *1* = isotropic gabbros; *2* = layered gabbros; *3* = intermediate zone between dunites and gabbros; *4* = dunites; *5* = harzburgites-dunites; *6* = harzburgites; *7* = tectonic contact; *8* = amphibolites and garnet-micashists; *9* = volcano-sedimentary complex; *10* = Upper Triassic limestones; *11* = Paleo-Moho; *12* = trace of the foliation plane. *M–M'*, *L–L'*, *K–K'*, *P–P'*, *E–E'*: lithological logs constructed from the cross-sections in the Kukës massif (see Fig. 2 for location).

dunite zone (HzD) where one finds some dunite lenses within the harzburgite; and (2) the upper dunite-harzburgite zone (DHz), in which the abundance of dunite lenses increases upwards. Both zones do have the same mineral paragenesis (olivine, orthopyroxene, clinopyroxene and spinel). The thickness of the harzburgite-dunite zone increases from 1000 m in the south to 2000 m in the north (Fig. 3). The chromite-rich deposits in this zone are of great economic interest.

3.3. The dunite zone (D)

The dunites are well developed in the Kukës massif. They contain some chromite layers, bodies or lenses that are currently exploited. In the northern part of the massif, very fresh dunites are mined for refractories. The dunites are massive and composed of olivine, spinel (chromite in the mineralized bodies), orthopyroxene (2–3%) and interstitial clinopyroxene (less than 3%). Locally, a secondary amphibole develops along orthopyroxene cleavages. The chromitic layering is locally folded or displaced along normal shear zones (Hoxha, 1993; Hoxha et al., 1993). The thickness of the massive dunite zone varies from 1.5 km to the south to 2.5 km to the north (Fig. 3).

3.4. The intermediate zone

The intermediate zone (Zi) corresponds to a progressive transition between the massive dunites (D) and the gabbros (G). This zone is comprised of intercaled olivine-clinopyroxenites, wehrlites, olivine-websterites and lherzolite lenses. Some dunite lenses identical to the massive dunites are also observed. The thickness of the intermediate zone varies from 300 to 450 m.

3.5. The gabbros

The gabbros are concordant on the top of the ultramafic unit to the west of the massif, but, to the north, the contact is faulted and vertical (Fig. 2). The gabbros at the base are layered, and become more isotropic toward the top (Fig. 3). The 1- to 10-cm-thick layering is defined by variation of the amounts of olivine, clinopyroxene, orthopyroxene and plagioclase. Therefore, some gabbros, gabbro-norites and plagioclase-bearing ultramafic rocks may be recognized (Streckheisen, 1973). Some secondary amphiboles are locally developed at the expense of pyroxenes.

4. Structural study

The structural study of the Kukës ultramafic massif has been conducted using the methods proposed by Nicolas et al. (1972), Nicolas and Poirier (1976) and Nicolas (1989). The orientations of planar (layering, foliation), linear (lineations) and tabular (veins) structural elements have been systematically measured (Figs. 4 and 5). The definition of spinel or pyroxene lineations used in the present work is the one proposed by Darot and Boudier (1975).

The layering observed in the different lithological zones is subparallel to the foliation. The structural maps show that the planar and linear structures are homogeneous in the upper part of the massif (Fig. 4). The average orientation of the foliation plane is

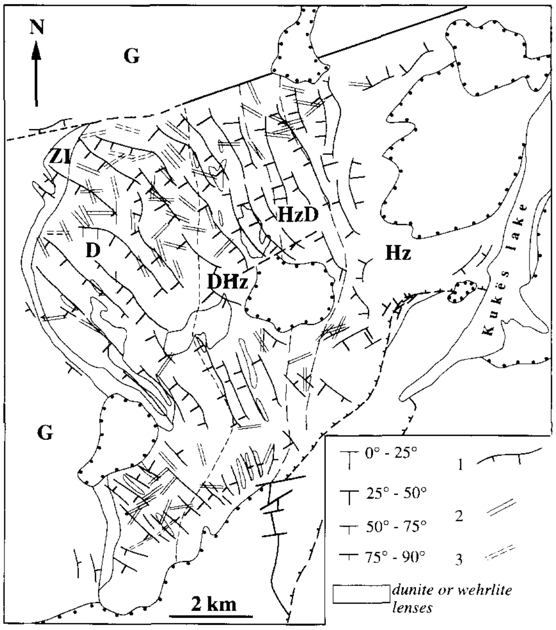

Fig. 4. Map of foliation trajectories in the Kukës ultramafic massif. *1* = high-temperature foliation trajectories, the barbed lines showing the direction and the dip value; *2* = pyroxenite dyke; *3* = gabbro dykes.

145°W45°, i.e., parallel to the Albanides, whereas the average orientation of the stretching lineation is 60°W40° to 60°. In most cases, the lineation is down dip in the foliation plane. The foliation is secant on the boundaries of the pyroxenite lenses in the dunite zone and is truncated at the base of the gabbro zone.

In the northern part of the massif, the foliation is slightly deflected in a narrow zone along the boundary fault and the pyroxene lineation defines a fan in the harzburgite zone (Fig. 5). In the southern part of the massif, some perturbations affect the foliation plane which shows some dip inversions (Fig. 4). This indicates the presence of some folds at the map scale but they have not been observed in the field. The lineation is regular in that area and has an average 55°W75° orientation (Fig. 6).

Numerous clinopyroxenite veins with comb-like texture intrude the harzburgites and dunites. Gabbro dykes are observed in the upper part of the ultramafics, near the base of the layered gabbros, and postdate the extensional structures (normal faulting) in the chromitic dunites. The average orientation of the ultramafic and mafic intrusive veins is 115°N40°, i.e., at a high angle to the stretching lineation.

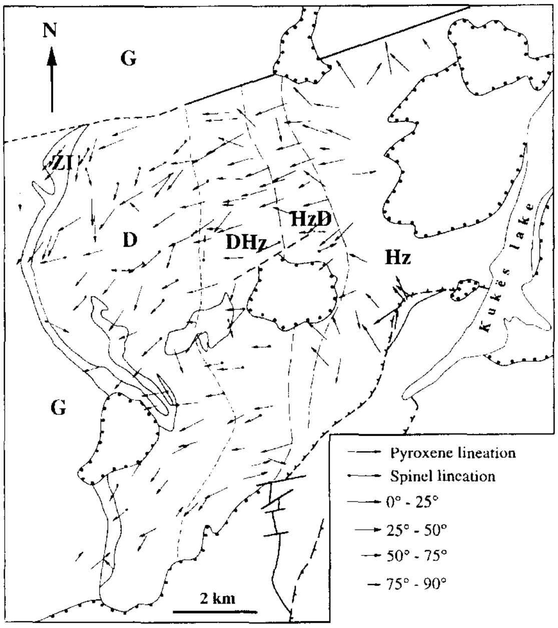

Fig. 5. Map of stretching lineations in the Kukës ultramafic massif.

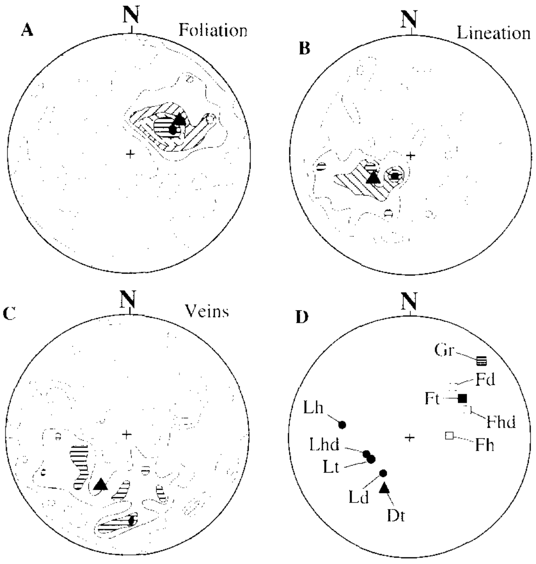

Fig. 6. Stereographic projection of the structural elements from Kukës ultramafic complex (Schmidt diagram, lower hemisphere projection). (A) Foliation (555 measurements) in the ultramafic unit. Contours at 0.18, 1.26, 2.58, 3.78, 5.05 and 6.31%. (B) Lineation (195 measurements) in the ultramafic unit. Contours at 0.15, 1.54 and 3.08%. (C) Poles of pyroxenite and gabbro dykes in the ultramafic unit (168 measurements). Contours at 0.59, 2.38 and 3.57%. ▲ = best axis of the distribution. (D) Best axes of the ditribution of remarkable planes and lines following different lithological zones in the Kukës ultramafic massif. *Fh* = foliation in harzburgite zone (140 measurements); *Fhd* = foliation in harzburgite-dunite zone (247 measurements); *Fd* = foliation in dunite zone (207 measurements); *Ft* = foliation in the whole massif (555 measurements); *Gr* = layering in the gabbros (35 measurements); *Dt* = dykes and veins (168 measurements); *Lh* = stretching lineation in the harzburgite zone (58 measurements); *Lhd* = stretching lineation in the harzburgite-dunite zone (70 measurements); *Ld* = stretching lineation in the dunite zone (83 measurements); *Lt* = total stretching lineation in the whole massif (195 measurements).

5. Textural study

The description of the textures observed in the Kukës massif is based on the classifications proposed by Harte (1977) and Mercier (1985). Two main types of textures have been recognized.

5.1. High-temperature textures

Most of the ultramafic rocks have equant or tabular granoblastic textures (Hoxha, 1993; Hoxha et al.,

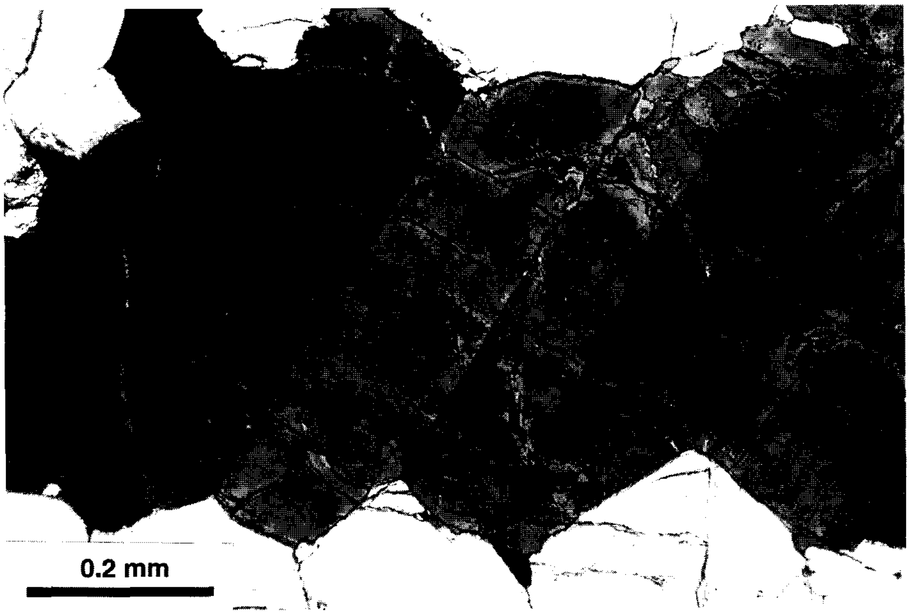

Fig. 7. Photograph of wide subgrains in olivine with straight (100) subgrain boundaries indicating a high-temperature deformation. Sample no. 45. The section is normal to the foliation plane with the lineation parallel to the long side of the photograph. Crossed nicols. Compare with Fig. 9.

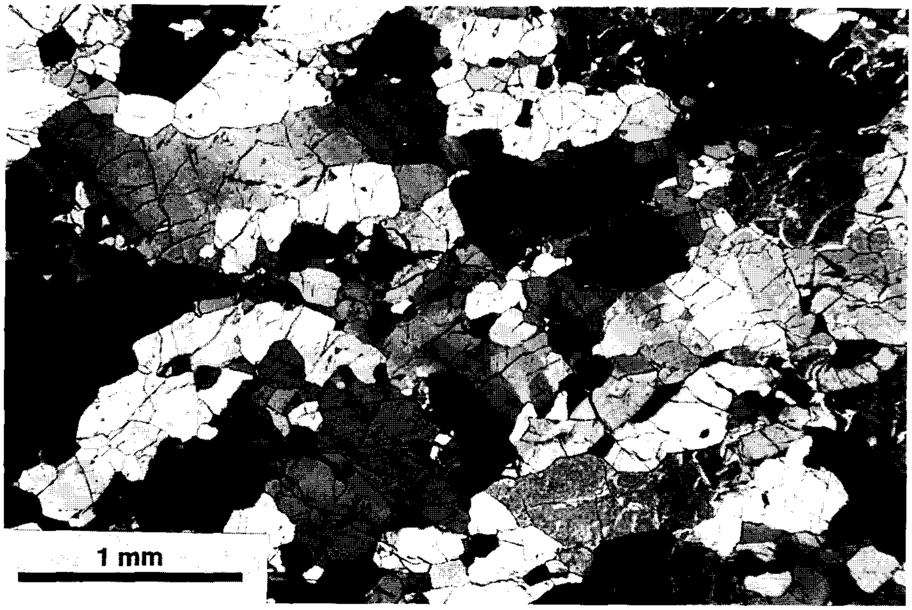

Fig. 8. Photograph of a porphyroclastic texture with a bimodal grain size of olivine (small neoblasts and large deformed porphyroclasts). Sample no. 44. The section is normal to the foliation plane with the lineation parallel to the long side of the photograph. Crossed nicols.

1993). Olivine crystals are coarse grained (1–3 mm) and contain sharp, straight and 100- to 150-μm-spaced (100) subgrain boundaries (Fig. 7). In some samples from the northern boundary of the massif, the grain size distribution is bimodal and typical of a porphyroclastic texture, i.e., large olivine porphyroclasts are surrounded by recrystallized grains (Fig. 8). The size of these neoblasts (0.1–0.15 μm in diameter) suggests high-temperature or low-stress deformation (Mercier et al., 1977; Ross et al., 1980; Guéguen and Darot, 1980; Karato et al., 1980; Karato, 1984). In the granoblastic dunites, spinel is locally included in the olivine crystals, suggesting that olivine grain growth occurred by grain boundary migration.

The intermediate zone and the base of the gabbro unit also exhibit high-temperature-deformation textures (kink bands in orthopyroxene, wide subgrains in olivine, deformed plagioclase).

5.2. *Low-temperature textures*

The first evidence of low-temperature deformation in the ultramafic rocks is the development of narrow (5- to 10-μm-spaced) and fuzzy (100) subgrain boundaries in olivine leading to undulatory extinction (Fig. 9). It generally corresponds to very low strain which does not induce elongation among the recrystallized grains formed at high temperature nor obliterate the previous high-temperature textures; deformation never reaches the porphyroclastic stage (no fine-grained recrystallized olivine). This low-temperature deformation is localized in a 2- to 3-km-wide band along the northern boundary of the Kukës massif.

6. Strain estimation

Strain has been estimated in the Kukës peridotites by measuring the elongation of minerals in the field or in *XZ* thin sections. Depending on the amount of serpentine present and on the grain size of the rock, 20 to 60 grains have been measured in each sample and the arithmetic mean calculated. The measured minerals show slight (1.5–2.8 for olivine) to moderate (1.3–3.6 for spinel) elongation throughout the whole massif. The values are somewhat irregular in

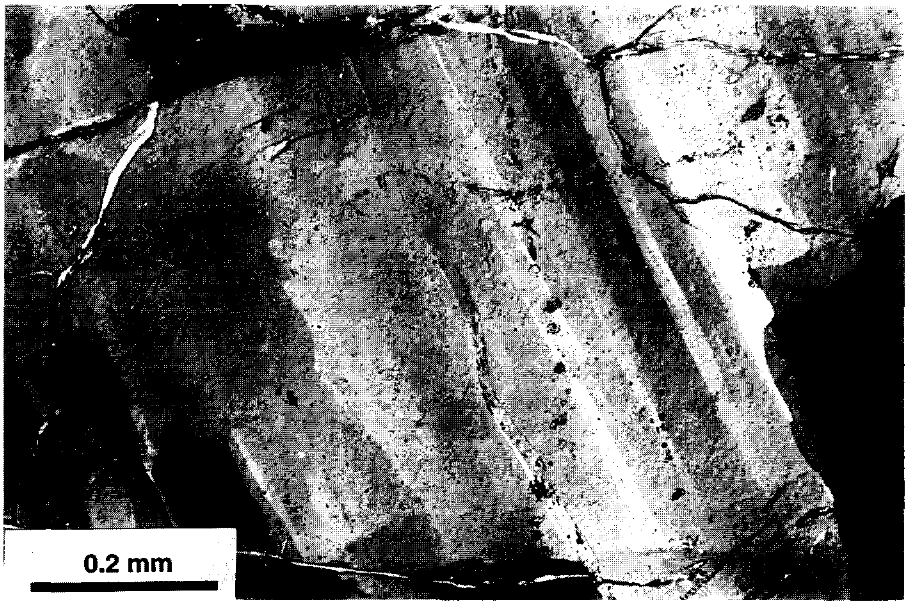

Fig. 9. Photograph of narrow olivine subgrains with fuzzy (100) subgrain boundaries indicating a low-temperature deformation. sample no. 45. The section is normal to the foliation plane with the lineation parallel to the long side of the photograph. Crossed nicols. Compare with Fig. 7.

the harzburgite and harzburgite-dunite zones (especially for spinel), but regular in the dunite zone where elongation decreases progressively upwards. Therefore, a significant elongation limit may be placed at the bottom of the dunite zone, above which olivine and spinel elongation becomes lower than 2.

7. Kinematic analysis of the Kukës massif

In peridotites, the lattice-preferred orientation of minerals (L.P.O.) acquired by intracrystalline deformation is strongly dependent on the active glide system (Nicolas et al., 1973), itself dependent on the

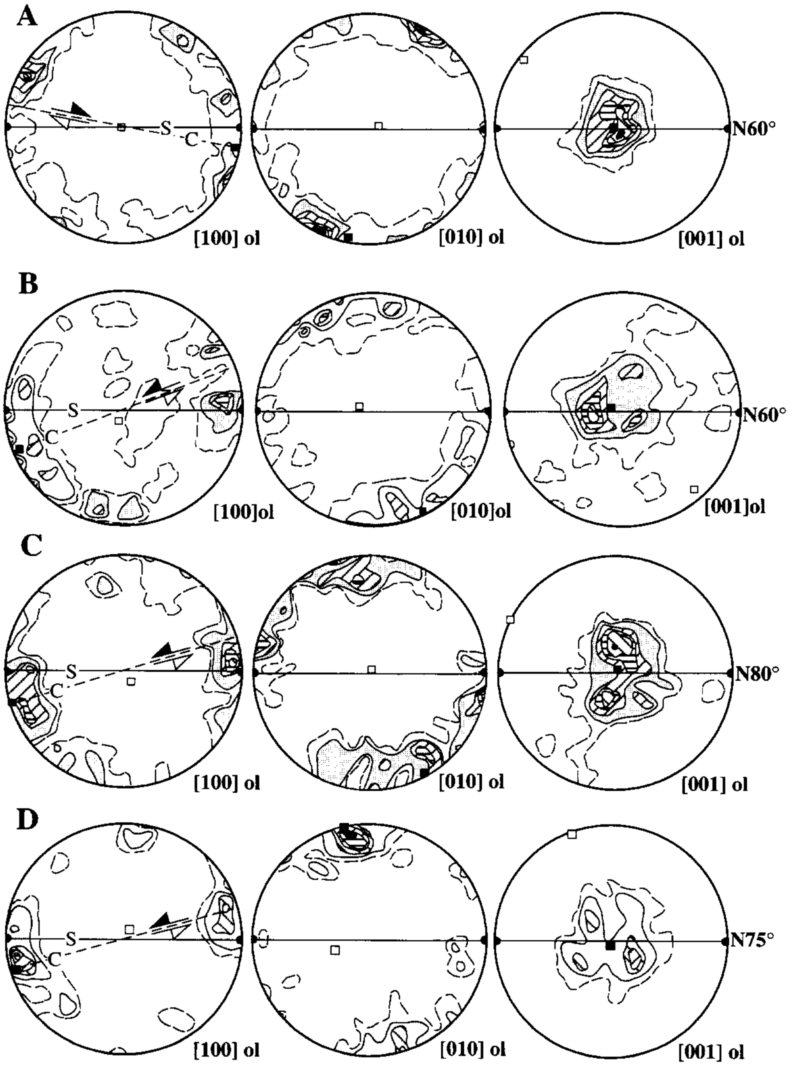

Fig. 10. Lattice-preferred orientation of olivine in some representative samples from the Kukës massif. Schmidt canevas, lower hemisphere. ■ = best axis of the distribution; □ = pole of the best plane of the distribution. The foliation (solid line) and the stretching lineation (●) are E–W oriented. Localization of the samples on Fig. 11. (A) Sample no. 237 (Hz zone), olivine, 70 measurements; contours at 1.42, 4.28, 7.14, 10.10, 12.85 and 15.71%. (B) Sample no. 233 (HzD zone), olivine, 100 measurements; contours at 1, 3, 5, 8 and 10%. (C) Sample no. 26 (DHz zone), olivine, 80 measurements; contours at 1.25, 2.5, 5, 7.5, 10 and 15%. (D) Sample no. 44 (D zone), olivine, 60 measurements; contours at 1.66, 3.33, 8.33 and 11.66%.

temperature and strain rate (Carter and Ave Lallemant, 1970). In order to determine the active glide system in olivine and the shear sense in peridotites, the olivine L.P.O. was measured in the *XZ* plane for nineteen samples distributed randomly over the whole massif. For each sample, the foliation plane was defined either in the field or on HF-bleached sections by the elongation of minerals such as spinel and orthopyroxene; it was then compared with the elongation of olivine porphyroclasts in thin sections. We will describe below the stereograms of some representative samples (Fig. 10; see Hoxha, 1993 for a complete study). For samples showing a porphyroclastic texture, only the porphyroclasts have been measured, since they are better kinematic indicators than the recrystallized neoblasts (Karato, 1987). However, some comparative measurements of both olivine neoblasts and porphyroclasts have been performed in a single sample and confirm that no significant difference exists between the L.P.O. of both types of grains.

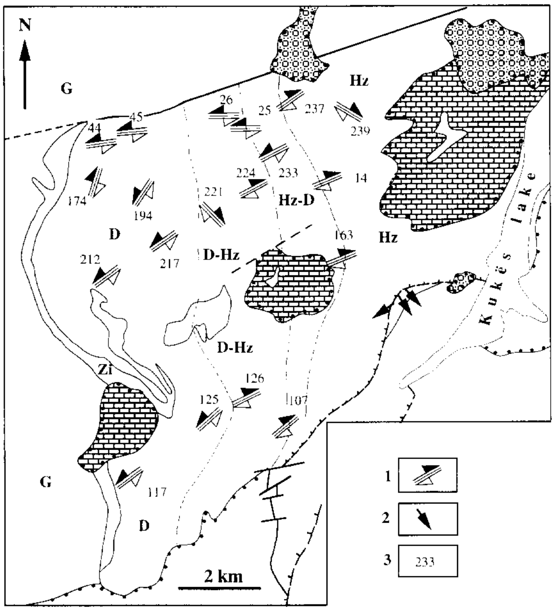

Fig. 11. Map of shear sense in the Kukës ultramafic massif. *1*: black arrows show the sense of movement of the upper block relative to the lower one in the peridotites. *2*: Sense of movement of the upper block relative to the lower one in the metamorphic sole. *3*: Number of the samples studied for lattice-preferred orientation.

In this study the shear senses in peridotites of the Kukës massif were determined from the obliquity between the best axis of the [100] olivine distribution (indicative of the shear direction or C-plane) and the foliation-lineation trace in *XZ* thin sections. Because the foliation has an average westward dip and the lineation is generally down dip in the foliation, the shear sense is described by considering the relative movement of the upper block relative to the lower block (Fig. 11).

7.1. Harzburgite and harzburgite-dunite zones

Sample no. 237

This harzburgitic sample has a porphyroclastic texture showing incipient (20%) high-temperature recrystallization. Evidence of low-temperature deformation is also observed. A 15.71% [001] concentration appears in the foliation plane (150°SW20°) and perpendicular to the lineation (240°SW20°, Fig. 10A). The axes [010] and [100] are distributed in girdles at the periphery of stereograms. The angle (α) between the best axis of [100] and the lineation direction is 13° and indicates a top-to-the-N60° shear sense.

Sample no. 233

This sample comes from a dunite lens within harzburgites (Fig. 11). The rocks have a granoblastic tabular texture with an incipient dynamic recrystallization. The direction of the foliation plane is 150°W30°, whereas the lineation direction is 240°W30°. The olivine L.P.O. is similar to that in sample no. 237 except that [100] and [010] are more concentrated in the outer girdle (Fig. 10B) and show a different symmetry relative to the lineation. Both samples 233 and 237 display similar deformation features (2.1 elongation of olivine). The angle between the [100] best axis and the foliation-lineation trace is 23° and the sense of shear is top-to-the-N240°.

Sample no. 26

This sample is a fresh harzburgite from the upper harzburgite-dunite zone located near the northern boundary fault (Fig. 11). Its structure is medium grain-size tabular granoblastic. The olivine L.P.O. is strong: [001] is perpendicular to the lineation in the foliation plane (Fig. 10C). The pole of (100) is close to the lineation. The shear sense is top-to-the-N260°.

7.2. The dunite zone

Sample no. 44

This sample is a fresh dunite with a porphyroclastic structure (Fig. 8). The olivine porphyroclasts display wide kink bands and a few wavy low-temperature kink bands. [100] defines a maxima near the lineation and [010] is well concentrated (11.66%) near the pole of the foliation plane. The [010] and [100] maxima lie in the outer part of the stereograms. Their position relative to the foliation-lineation trace indicates a top-to-the-N255° shear sense.

All the measured olivine L.P.O. in the Kukës peridotites originated from the activation of the (010)[100] glide system and corresponds to high-temperature deformation (Carter and Ave Lallemant, 1970; Nicolas and Poirier, 1976). The (100) measured orientation of the subgrain boundaries is consistent with that inferred glide system. However, it was not possible to determine the glide plane from the orientation of the subgrain boundaries and the external rotational axis due to the imprecision of the universal stage measurements for small rotation angles.

The L.P.O. measured in the Kukës massif are somewhat different from those described in other ophiolites such as in Lanzo (Boudier, 1976), Oman (Boudier and Coleman, 1981), Bay of Island (Girardeau, 1979; Girardeau and Nicolas, 1981) and Xigaze (Girardeau and Mercier, 1988). In most peridotite massifs described by these authors, the observed olivine lattice-preferred orientations show a maximum of [100] axis close to the lineation but girdles of [010] and [001] subperpendicular to the lineation. Such a girdle disposition is interpreted by the authors as produced by the activation of the (0kl)[100] glide system in olivine which is activated at a lower temperature than the (010)[100] system (Carter and Ave Lallemant, 1970).

Concerning the kinematic interpretation of the olivine L.P.O., the Kukës samples generally indicate a top-to-the-E shear sense in the harzburgite and harzburgite-dunite zones, except for two samples close to the northern boundary fault in which the shear sense is reversed. On the contrary, most samples in the dunite zone except one, show a top-to-the-W shear sense. Therefore, the base of the dunite zone is not only a lithological limit but also a kinematic boundary.

The angle between the foliation and the shear plane, as indicated by the angle between the elongation lineation and [100] maximum on the L.P.O. stereograms, varies between 13 and 23° in the samples described in this paper. This angle α is theoretically linked to the shear strain γ by the relationship $\gamma = 2 \cot 2\alpha$ (Nicolas et al., 1973) and corresponds to γ values of 4.1 to 1.9. These high γ values are inconsistent with the low elongation measured for spinel and olivine crystals in the ultramafics.

Concerning the shear sense in the metamorphic sole, only three samples (Fig. 11) have been studied, giving conflicting results based on rotated garnets or average orientation of subgrain boundaries in quartz (top-to-the-SSE or SSW shear sense). Therefore, it is not yet possible to determine the sense of movement of the ophiolitic nappe.

8. Discussion

The Kukës ultramafic massif represents a classical ophiolite as described by Nicolas (1989). The harzburgites present quite irregular foliation trajectories (Fig. 4) that could correspond to the deformation induced by a rising asthenospheric diapir close to the ridge (Ceulener et al., 1988). However, the dimensions of the massif do not allow confirmation of this hypothesis (Hoxha et al., 1993). All structural elements (foliation, lineation, dykes) are more coherent above the harzburgite zone, indicating homogeneous flow whose kinematics may be approached by the study of olivine lattice-preferred orientations.

The foliation trajectories are deflected along the northern faulted contact between the ultramafic and mafic units (Fig. 4). This fault, which extends to the west and east outside of the mapped area, is parallel to the Shkodër–Pejë line (Fig. 1) and is underlain by a strong asthenospheric deformation (presence of porphyroclastic structure, Hoxha and Boullier, unpublished data) and by a local inversion of shear sense in the northern part of the harzburgite-dunite zone (Fig. 11). A low-temperature lithospheric deformation is also observed along this fault which could therefore be described as a transform fault (Prinzhofer and Nicolas, 1980; Sécher, 1981) whose orientation is more or less subparallel to the flow direction (lineation). Taking into consideration (1) the parallelism with the Shkodër–Pejë (Skadar–Pec) transform fault (Pamic, 1983) and (2) the orientation of different veins in the vinicity of this shear zone compared to those documented by Prinzhofer and Nicolas (1980) and Nicolas (1989), this fault can be interpreted tentatively as a dextral transform fault (Fig. 13).

In general, the Kukës peridotites are apparently weakly deformed although L.P.O. seems to indicate very strong deformation. The X/Z olivine and spinel elongation is variable in the harzburgite and harzburgite-dunite zones but slightly and regularly decreases from the bottom to the top of the massif. The low elongation of spinel could be explained by the refractory character of this mineral and by its low ductility (Doukhan et al., 1984): chromite grains frequently display extensional fractures where olivine crystals are plastically deformed. On the contrary, the low elongation of olivine is not compatible with the small α angle or high γ values determined from the L.P.O. of olivine crystals. Therefore, these conflicting observations are interpreted as resulting from annealing and recrystallization processes in olivine (grain growth or grain boundary migration) occurring at high temperature in the dunite zone. Numerous spinel grains entirely enclosed in olivine crystals in dunites are also indicative of such processes. Percolation of magma through harzburgites and dunites (Hoxha, 1993) could have enhanced recrystallization by grain boundary migration. Another consequence of recrystallization by grain boundary migration is that the relationship between olivine L.P.O. (α angle) and shear strain is not reliable.

The systematic study of olivine lattice-preferred orientation (Hoxha, 1993) suggests that the (010)[100] glide system was activated in olivine. Therefore, deformation occurred at high temperature. In some samples, the [100] and [010] olivine crystallographic axes are scattered in outer girdles (Fig. 10A and C). Such a disposition could be explained by the superimposition of two deformations: a high-temperature deformation and a low-temperature deformation characterized by the activation of a (110)[001] or (100)[001] glide system (Carter and Ave Lallemant, 1970). The low-temperature deformation was of variable intensity but was never sufficient to crush its own foliation. In all cases, the high-temperature deformation predominates.

The lattice-preferred orientation that we have determined is relatively comparable with that determined by Girardeau et al. (1988) at the Galicia margin, Mercier and Nicolas (1975) in the peridotite xenoliths from basalts and Boullier and Nicolas (1975) and Boullier (1977) in the peridotite xenoliths from kimberlites. Extrapolating the experimental data for olivine (Raleigh, 1968; Carter and Ave Lallemant, 1970; Raleigh and Kirby, 1970; Nicolas et al., 1973) to geological deformation rates (e.g., 10^{-14} s^{-1}), we may deduce that the Kukës peridotites were deformed at a minimum temperature of or some hypersolidus condition ($\geq$ 1200°C, Nicolas and Vialon, 1980).

Taking into consideration the remarkable homogeneity of different structures such as foliation, lineation and veins (in pyroxenites, gabbros, dunites, etc...) as well as the absence of major perturbations (later strike-slip faults) in the different lithological zones, one can integrate the structural data in a kinematic framework for the creation of oceanic lithosphere in the Kukës massif. In order to do that, we used the approach of Juteau et al. (1977), Nicolas (1986) and Nicolas (1989) in defining the paleo-horizontal plane and the direction of the paleo-ridge axis. In ophiolites, the paleo-horizontal plane may be defined as the base of the mafic unit, generally composed of layered gabbros. This limit, corresponding to the seismic Moho in the oceanic lithosphere (Nicolas, 1989), was considered as paleo-horizontal in this study. The transport direction in the dunites and the harzburgites is given by the stretching lineation. Its average direction is east-northeast–west-southwest after rotation of the paleo-Moho to the horizontal (Fig. 12). Nicolas and Poirier (1976) consider that the lineation is at a high angle to the ridge axis, whereas Vogt and Johnson (1975) and Bird and Philips (1975) described some cases of obliquity between the plastic flow and the accretion zone; this obliquity was found also in other ophiolitic massifs (Nicolas, 1989). In order to define the accretion geometry of the lithosphere (ridge axis), the orientation of the sheeted dyke complex may be used as well, as suggested by Nicolas and Poirier (1976) and Nicolas (1989). However, no structural data on the

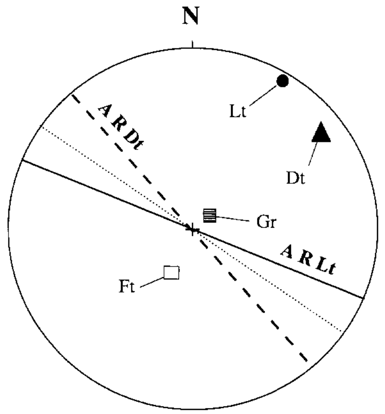

Fig. 12. Structures in the Kukës ultramafic massif after rotation of the base of the layered gabbros (paleo-Moho) to the horizontal (Schmidt canevas, lower hemisphere projection). The ridge axis is supposed to be parallel to the average orientation of veins in the ultramafic unit. *Ft* = total foliation plane in the ultramafic unit; *Lt* = total stretching lineation in the ultramafic unit; *Gr* = layering in the gabbros; *Dt* = pyroxenite and gabbro veins; *ARDt* = direction of paleo-ridge estimated from the average orientation of the veins; *ARLt* = direction of paleo-ridge estimated from the average orientation of the stretching lineation.

sheeted dyke complex are available in the Kukës massif. In the absence of these data, we equated the orientation of the veins within the ultramafics (pyroxenites and gabbros) with the orientation of the paleo-ridge axis. The veins are orientated in a regular manner within the massif (Fig. 2) and their average orientation after rotation of paleo-Moho to the horizontal becomes N142° (Fig. 12). This direction is subparallel to the sheeted dyke complex of the Mirdita ophiolitic nappe (Geological map of Albania, 1983). In the Kukës massif, the veins are at a high angle to the lineation (Fig. 12).

After rotation of the paleo-Moho to the horizontal plane, it can be observed that (Figs. 12 and 13):

(1) In the layered gabbros, the plane of magmatic accumulation is weakly inclined (5–10°) to the southeast, i.e., towards the axis of the magma chamber under the ridge.

(2) In the peridotites (dunites and harzburgites), the dip of the foliation plane is in the opposite sense and increases downwards. In the dunites, the foliation plane is subparallel to the major lithological contact (the paleo-Moho).

(3) The lineation is at a high angle to the veins and thus to the constructed ridge axis. That disposition gives a classic image of an accretion zone.

(4) The shear sense deduced from olivine lattice-preferred orientation, and its inversion at the base of the dunite zone, are consistent with the model of

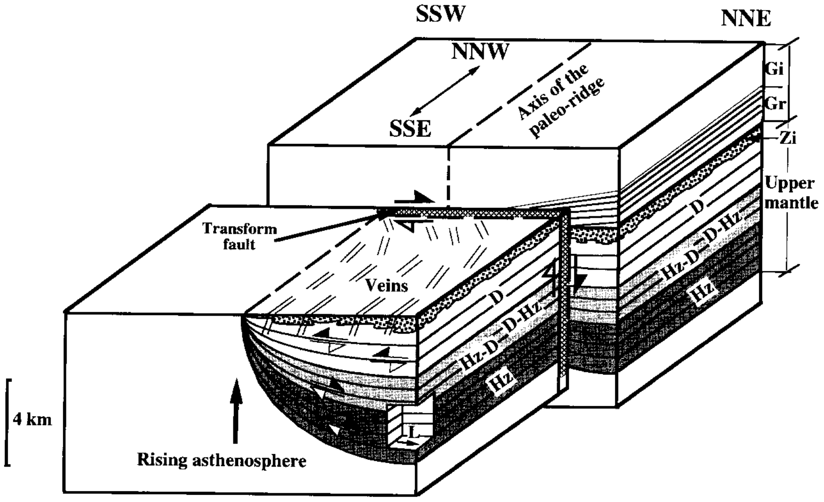

Fig. 13. Kinematic model for the creation of oceanic lithosphere in the Kukës ultramafic massif established from structural data after rotation of the paleo-Moho to the horizontal. *Gi* = isotropic gabbros; *Gr* = layered gabbros; *Zi* = intermadiate zone; *D* = dunite zone; *D-Hz* = dunite-harzburgite sub-zone; *HzD* = harzburgite-dunite sub-zone; *Hz* = harzburgite zone; *L* = stretching lineation in the ultramafic massif; black arrow = displacement of the upper block relative to the lower one as determined from the lattice-preffered orientation.

forced plastic flow diverging from a diapiric intrusion beneath the ridge (Rabinowicz et al., 1984).

From the observations and the reconstruction presented above, it appears that the Kukës ultramafic massif could correspond to the eastern flank of a NNE–SSE-oriented ridge.

9. Conclusion

In the Kukës ultramafic massif, the olivine lattice-preferred orientation induced by (010)[100] intracrystalline gliding indicates that high-temperature flow was predominant. A low-temperature deformation occurred on the northern margin of the massif but was never sufficient to obliterate the previous high-temperature structures. As indicated by the foliation trajectories, the flow is irregular at the base of the section and regular in the dunite zone. The apparent low finite strain of olivine is interpreted as the result of high-temperature recrystallization and fluid-enhanced grain boundary migration. An inversion of the shear sense is observed along the dunite-harzburgite boundary and is attributed to a forced plastic flow due to a velocity gradient change below the paleo-Moho (Rabinowicz et al., 1984).

After rotation of the paleo-Moho to the horizontal plane, the structural data show that the Kukës ultramafic massif could represent the eastern flank of a NNW–SSE-oriented paleo-ridge. Complementary structural studies of the other ultramafic massifs in the Mirdita zone in Albania are necessary to confirm this reconstruction and to better understand the kinematics of low-temperature obduction onto a continental environment.

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References

Anonymous, 1972. Penrose field conference on ophiolites. Geotimes, 17: 24–25.

Bard, J.P., 1990. Microtextures des Roches Magmatiques et Métamorphiques. Masson, Paris, 208 pp.

Bird, P. and Philips, J.D., 1975. Oblique spreading near the Oceanographer fracture. J. Geophys. Res., 80: 4021–4027.

Boudier, F., 1976. Le massif lherzolitique de Lanzo (Alpes Piémontaises). Etude structurale et pétrologique. Ph.D. Thesis, Nantes.

Boudier, F. and Coleman, R.G., 1981. Cross-section through the peridotite in the Samail ophiolite, Oman, southeastern Oman Mountains. J. Geophys. Res., 86: 2573–2592.

Boudier, F. and Nicolas, A., 1985. Harzburgite and lherzolite subtypes in ophiolitic and oceanic environments. Earth Planet. Sci. Lett., 76: 84–72.

Boullier, A.M., 1977. Structure des péridotites en enclaves dans les kimberlites d'Afrique du Sud. Conséquence sur la constitution du manteau supérieur. Bull. Soc. Fr. Minéral. Cristallogr., 100: 214–229.

Boullier, A.M. and Nicolas, A., 1975. Classification of textures and fabrics of peridotites xenoliths from South African kimberlites. In: L.H. Ahrens, J.B. Dawson, A.R. Duncan and A.J. Erlank (Editors), Physics and Chemistry of the Earth. Pergamon Press, Oxford, Vol. 9, pp. 467–475.

Carter, N.L. and Ave Lallemant, H.G., 1970. High temperature flow of dunite and peridotite. Geol. Soc. Am. Bull., 81: 2181–2202.

Cassard, D., 1980. Structure et origine des gisements de chromite du massif du Sud (ophiolites de Nouvelle Calédonie). Guides de prospection. Doct. 3ème cycle, Nantes.

Ceulener, G., Nicolas, A. and Boudier, F., 1988. Mantle flow patterns at an oceanic spreading centre: the Oman peridotites record. Tectonophysics, 151: 1–26.

Ciric, B., 1984. Complexe ophiolitique et formations comparables dans les Dinarides. Mém. Belgrade, 23: 1–44.

Darot, M. and Boudier, F., 1975. Mineral lineation in deformed peridotites: kinematic meaning. Pétrologie, 1: 225–236.

Doukhan, N., Doukhan, J.-C., Nicolas, A. and Sécher, D., 1984. Transmission electron microscope analysis of the deformation of chromites from ophiolites. Bull. Minéral., 107: 777–793.

Frashëri, A., Lubonja, L., Nishani, P., Bushati, S., Hyseni, A. and Leci, V., 1991. Les données géophysiques sur les relations entre les zones tectoniques des Albanides à terre et sur le plateau continental de la mer Adriatique. Colloque sur la Géologie de l'Albanie, Paris, Soc. Géol. Fr.

Geological map of Albania, 1983. 1/200000. Inst. Stud. Proj. Gjeol. Inst. Gjeol. Naftës, Tirana.

Girardeau, J., 1979. Structure des ophiolites de l'ouest de Terre

Neuve et modèle de croûte océanique. Doct. 3ème cycle, Nantes.

Girardeau, J. and Mercier, J.C., 1988. Petrology and texture of the ultramafic rocks of the Xigaze ophiolite (Tibet). Constraints for mantle structure beneath low-spreading ridges. Tectonophysics, 147: 33–58.

Girardeau, J. and Nicolas, A., 1981. Structure in two of the Bay of Islands (Newfoundland) ophiolite massifs: a model for oceanic crust and upper mantle. Tectonophysics, 77: 1–34.

Girardeau, J., Evans, C.A. and Beslier, C.A., 1988. Structural analysis of plagioclase-bearing peridotites emplaced at the end of continental rifting: hole 637D, ODP leg 103 on the Galicia margin. Proc. ODP, Sci. Results, 103: 209–223.

Guéguen, Y. and Darot, M., 1980. Microstructures and stresses in naturally deformed peridotites. Rock Mech. Suppl., 9: 159–172.

Harte, B., 1977. Rock nomenclature with particular relation to deformation and recrystallization textures in olivine-bearing xenoliths. J. Geol., 85: 279–288.

Hoxha, M., 1993. Etude structurale et pétrologique des péridotites de l'ophiolite de Kukës (Albanie): cinématique de la déformation et géométrie de la ride. Doct. Univ. I.N.P.L., Nancy.

Hoxha, M., Boullier, A.M. and Zhupi, S., 1993. Structure des péridotites de l'ophiolite de Kukës (Albanie). Bull. Soc. Géol. Fr., 164: 449–458.

Juteau, T., Nicolas, A., Dubessy, J., Fruchard, J.C. and Bouchez, J.L., 1977. Structural relationships in the Antalya ophiolite Complex, Turkey. Possible model for an oceanic ridge. Geol. Soc. Am. Bull., 88: 1740–1748.

Karato, S.I., 1984. Grain-size distribution and rheology of the upper mantle. Tectonophysics, 104: 155–176.

Karato, S.I., 1987. Seismic anisotropy due to lattice preferred orientation of minerals: kinematic or dynamic? In: M.H. Manghnani and Y. Syono (Editors), High-pressure Research in Mineral Physics. Am. Geophys. Union, Washington, DC, pp. 455–471.

Karato, S.I., Toriumi, M. and Fujii, T., 1980. Dynamic recrystallization of olivine single crystals during high-temperature creep. Geophys. Res. Lett., 7: 649–652.

Kodra, A., Vergely, J., Gjata, K., Bakalli, F. and Godroli, M., 1993. La formation volcano-sédimentaire du Jurassique supérieur: témoin de l'ouverture du domaine ophiolitique dans les Albanides internes. Bull. Soc. Géol. Fr., 164: 61–67.

Mercier, J.C., 1985. Olivine and pyroxenes. In: H.R. Wenk (Editor), Preferred Orientation in Metals and Rocks: an Introduction to modern Texture Analysis. Academic Press, New York, NY, pp. 407–430.

Mercier, J.C. and Nicolas, A., 1975. Textures and fabrics of upper mantle peridotites as illustrated by xenoliths from basalts. J. Petrol., 16: 454–487.

Mercier, J.C., Anderson, D.A. and Carter, N.L., 1977. Stress in the lithosphere: inferences from steady-state flow of rocks. Pure Appl. Geophys., 115: 199–226.

Nicolas, A., 1986. A melt extraction model based on structural studies in mantle peridotites. J. Petrol., 27: 999–1022.

Nicolas, A., 1989. Structures of Ophiolites and Dynamics of Oceanic Lithosphere. Kluwer, Dordrecht.

Nicolas, A. and Poirier, J.P., 1976. Crystalline Plasticity and Solid-state Flow in Metamorphic Rocks. Wiley Interscience, London, 444 pp.

Nicolas, A. and Vialon, P., 1980. Les mécanismes de déformation ductile dans les roches. Bull. Soc. Géol. Fr., 10: 127–139.

Nicolas, A., Bouchez, J.L. and Boudier, F., 1972. Interprétation cinématique des déformations plastiques dans le massif de lherzolites de Lanzo (Alpes Piémontaises). Comparaison avec d'autres massifs. Tectonophysics, 14: 143–171.

Nicolas, A., Boudier, F. and Boullier, A.M., 1973. Mechanisms of flow in naturally and experimentally deformed peridotites. Am. J. Sci., 273: 853–876.

Pamic, J., 1983. Considerations on the boundary between lherzolite and harzburgite subprovinces in the Dinarides and northern Hellenides. Ofioliti, 8: 153–163.

Prinzhofer, A. and Nicolas, A., 1980. The Bogota peninsula, New Caledonia: a possible oceanic transform fault. J. Geol., 88: 378–398.

Rabinowicz, M., Nicolas, A. and Vigneresse, J.L., 1984. A rolling mill effect in asthenospheric beneath oceanic spreading centers. Earth Planet. Sci. Lett., 67: 97–180.

Raleigh, C.B., 1968. Mechanism of plastic deformation of olivine. J. Geophys. Res., 13: 5391–5406.

Raleigh, C.B. and Kirby, S.H., 1970. Creep in the Upper Mantle. Mineral. Assoc. Am., 113 pp.

Rocci, G., Ohnenstetter, D. and Ohnenstetter, M., 1975. La dualité des ophiolites téthysiennes. Pétrologie, 1: 172–174.

Ross, J.V., Ave Lallemant, H.G. and Carter, N.L., 1980. Stress dependence of recrystallized-grain and subgrain size in olivine. Tectonophysics, 70: 39–61.

Sécher, D., 1981. Les lherzolites ophiolitiques de Nouvelle Calédonie et leurs gisements de chromite. Déformation de la chromite. Doct. 3ème cycle, Nantes.

Shallo, M., Kote, D.H., Vranai, A. and Përmeti, I., 1985. Magmatisme ophiolitique de la RPS d'Albanie. Inst. Stud. Proj. Gjeol., Tirana, 362 pp.

Streckheisen, A., 1973. Plutonic rocks. Classification and nomenclature recommended by the I.U.G.S. subcommission on the systematics of igneous rocks. Geotimes, 18: 26–30.

Turku, I., 1987. Metamorphites from the Kukës ultramafic massif (northeastern margin of the Mirdita ophiolitic zone, Albania). Ofioliti, 12: 137–150.

Vogt, P.R. and Johnson, G.L., 1975. Transform faults and longitudinal flow below the mid oceanic ridge. J. Geophys. Res., 80: 1399–1428.