Pergamon

Journal of Structural Geology, Vol. 16, No. 11, pp. 1495 to 1507, 1994

0191–8141/94 $07.00+0.00

0191–8141(94)00054–9

Deformation and metamorphic evolution of a large-scale fold in the lower crust: the Dumbara synform, Sri Lanka

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(*Received* 7 *September* 1993; *accepted in revised form* 21 *April* 1994)

Abstract—In the granulite facies Highland Complex of Sri Lanka, large scale folds are formed after the peak of metamorphism (800–850°C, 8–9kbar) at slightly lowered temperatures (700–750°C). The folds formed with axes parallel to a marked stretching lineation which is a continuous element throughout the whole deformational sequence.

In the central parts of the Highland Complex a lithostratigraphic sequence derived from shelf-type sediments can be established. Numerous laccolithic or sill-like intrusions with granitic to basic composition thickened the pile. A pronounced foliation and stretching lineation (S_1/L_1) parallel to compositional layering formed during peak metamorphic conditions and continued towards lower temperatures. Isoclinal folding with axes parallel or normal to this stretching lineation locally complicates the lithological sequence (D_2/D_3). Deformation involving grain-size sensitive mechanisms led to an equalizing of viscosities between layers of differing composition, and thus to very homogeneous deformation during the highest temperature stage. The Dumbara Synform is one of several large scale fourth folds which formed at slightly decreased temperature. The fold mechanism changed from the outer, deeper level towards the core of the fold. Stretching parallel to fold axis continued during folding causing small scale extensional shear zones, which formed after return to marked viscosity contrasts between compositionally different layers. Deformation ended in the higher amphibolite facies and was followed by static annealing and slow strain-free exhumation.

INTRODUCTION

The granulite facies rocks of Sri Lanka are exposed in an arcuate belt extending from the north-northeast across the central Highlands to the south-southwest (Fig. 1), the Highland Complex. For lithological reasons this granulite belt was subdivided by Cooray (1962) into a southwestern part, the SW-Group, with a more pelite dominated sequence, and a northern part, the Highland Series. Yet, both subunits share a common structural and metamorphic history and differences are not very marked. Kröner *et al.* (1991) introduced the term 'Highland Complex' to combine both units.

The Highland Complex was thrust over the Vijayan Complex in the east and southeast along a subhorizontal thrust plane which underlies large parts of the Highland Complex (Kleinschrodt 1994). The Vijayan Complex differs in lithological, structural, metamorphic and geochronological respects. It is essentially made up of granite to granodioritic gneisses, migmatites and subordinate metasediments and was metamorphosed under upper amphibolite facies conditions.

Thrusting occurred under upper amphibolite facies conditions and was followed by slow cooling without further deformation. Within the Highland Complex deformation ended at granulite facies conditions. Therefore, Sri Lanka offers an excellent opportunity to study deformation in the lower crust free from later overprint.

In the central Highlands, in the area around Kandy, large folds with wavelengths of about 7–10 km and an exposed length parallel to their axes up to 50 km offer excellent insight in the lithology, metamorphism and structural evolution of the granulite facies Highland Complex. This study is dedicated to one of the largest of these folds, the Dumbara Synform to the east of Kandy. It is exposed for about 40 km along and 8 km normal to axial trend.

THE DUMBARA SYNFORM

The Dumbara Synform is one of the largest of several doubly-plunging synforms in the granulite facies Highland Complex of central Sri Lanka. It is situated east of Kandy and extends more than 40 km north from Rigilagaskada nearly to Matale (Fig. 1). We chose this synform as a model case for the evolution of large folds in the deeper continental crust, and its structural evolution can be extrapolated to large parts of the central Highlands of Sri Lanka.

Lithological sequence

A lithological profile through the southwest-limb of the Dumbara Synform (Fig. 2) is exposed along a new road at the southwest-shore of the Victoria Dam Reservoir (southeast of Kandy).

The high-grade rocks derive from a sedimentary sequence, dominated by psammites and psammopelites. Four marble horizons and several quartzites are intercalated. The sedimentary sequence was intruded by granitic and basic intrusives. These intrusives are parallel to the compositional layering and probably formed sills or laccolithic bodies. In large parts of the Highland Com-

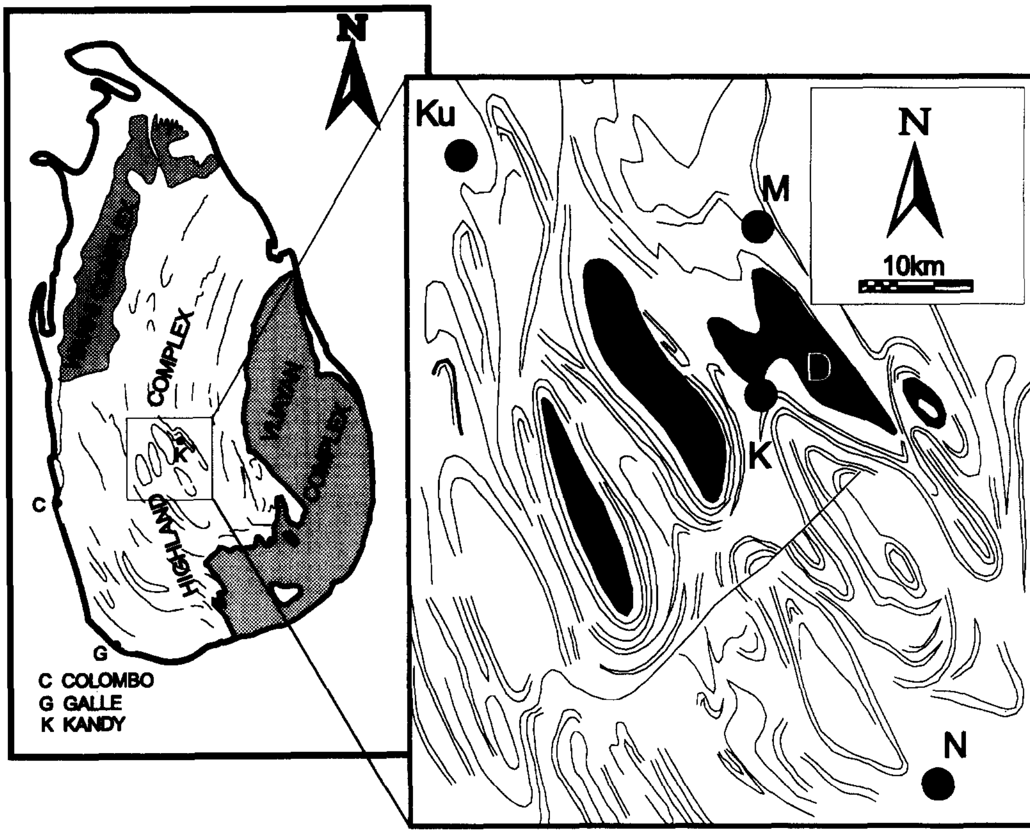

Fig. 1. Structural trend lines compiled from the Geological Map of Sri Lanka (1982). The main units are shown on the left side. The main structural trends are traced from marble horizons, quartzites and granite sills. The basic complex filling several of the large synform is marked with dark signature. The Dumbara Synform is marked by D. To the east follows the Hulu Ganga Synform; to the west the Gadeladenia and Aranayaka Synforms. Locations are: K—Kandy, Ku—Kurunegala, M—Matale, N—Nuwara Eliya.

plex (e.g. in the south and southeast) the total amount of granites is much higher than in the described sequence. The top of the lithological profile is formed by basic rocks, which are interpreted as remnants of a large layered basic intrusive complex (Kleinschrodt *et al.* 1991). These rocks form the filling of several large scale synforms in the central Highland Complex. The base of these rocks is formed by a layer of granitic gneisses ('pink feldspar granite'). The roof is not exposed in the Dumbara synform.

In the next neighbouring synform to the east, the Hulu Ganga Synform, we found similar basic rocks in the same lithostratigraphic position. There, the roof of the intrusion is exposed. The total thickness of the intrusion in this synform is about 200–250 m (Kleinschrodt *et al.* 1991). It may be thicker in the Dumbara Synform.

The sequences within the layered basic rocks of the Hulu Ganga Synform and the Dumbara Synform are similar. For example the position of the layered sequence within the intrusion, which we described as a 'critical zone' in the Hulu Ganga synform (Kleinschrodt *et al.* 1991) is identical. One remarkable, 50 cm wide layer close to the base of the layered sequence contains large garnets with diameters up to 12 cm. This is the only layer within the basic intrusion containing garnet and appears at exactly the same position in both synforms, about 6 m above the base. The single large garnets are separated by garnet-free amphibolites, and the distance between the garnets seems to correlate with their size.

Sedimentary layers can be compared very well between the two neighbouring synforms and maybe even to the synforms further west. The individual marble horizons, forming excellent marker beds, are found all around the southern half of the synform with their individual characteristics (Fig. 3). However, the thickness of sedimentary units differs considerably. For example, the psammopelite sequence below the layered intrusion is 2–3 times thicker in the Dumbara Synform than in the Hulu Ganga Synform.

The granite layers within the sedimentary sequence are perfectly parallel to the layering of the metasediments. They usually part the sedimentary layers perfectly and do not incorporate fragments of the wall rocks. Within the Dumbara Synform two of these granites were found (Fig. 3) (just one of them is exposed along the new road from where we have described the lithological profile). They have a characteristic position within the lithostratigraphic sequence. Their thickness varies between 20 and 60 m.

The upper of these sills ('Digana Granite') shows the following characteristics: about 40% of K-feldspar-phenocrysts, now more or less recrystallized to feldspar-clusters; dispersed graphite content (secondary); no basic layers. The lower one ('Victoria Granite') is

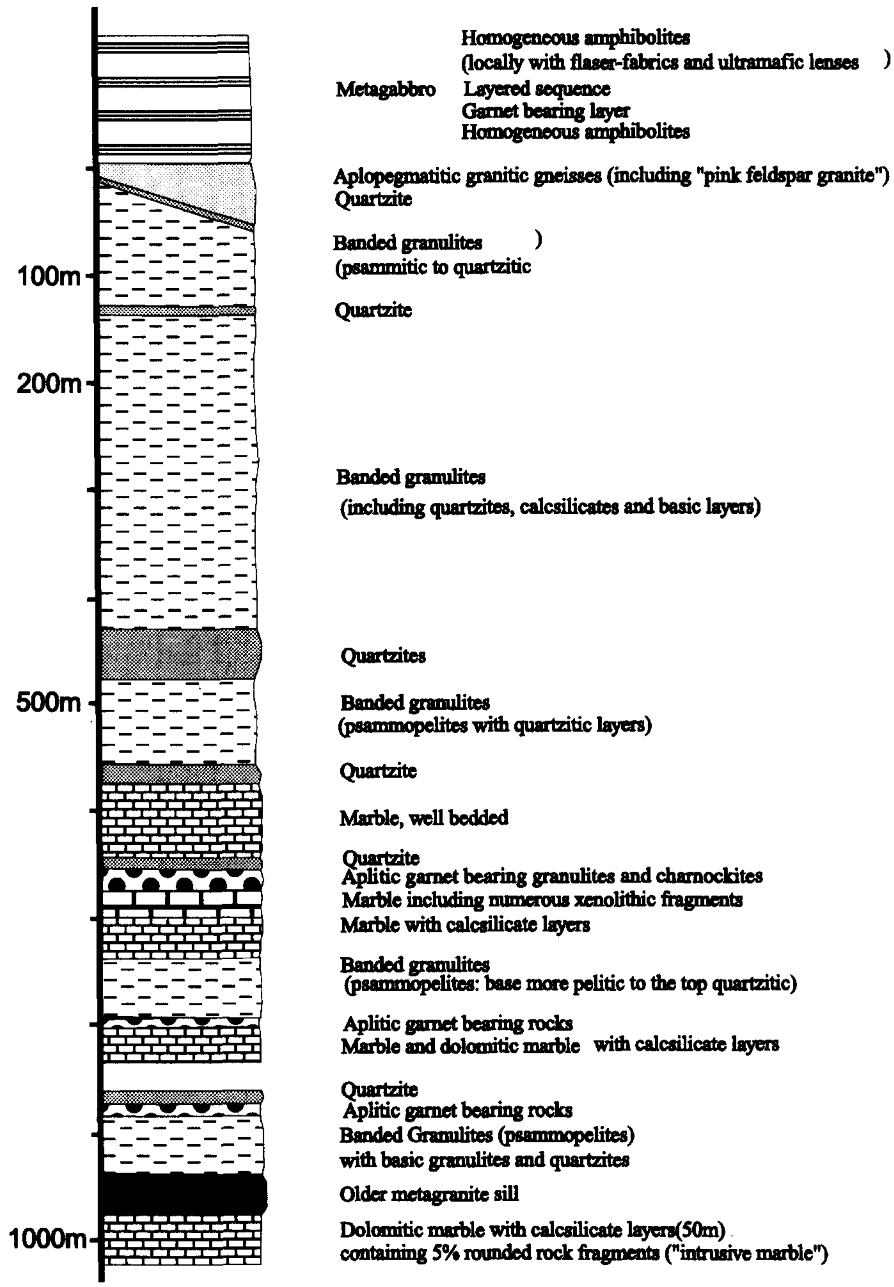

Fig. 2. Lithostratigraphic section as exposed in a cross-section through the western limb of the Dumbara Synform along the new road southeast from Kandy along the southwest-shore of the Victoria Dam reservoir, between Kandy/Mahaweli Bridge and the junction with the main road from Kandy to Hanguranketa.

characterized by: less or no K-feldspar-phenocrysts; less garnet compared to the Digana Granite; no graphite and numerous basic layers. Both granite sills are internally quite homogeneous, and show no compositional layering, except for the basic layers in the Victoria Granite. The latter result from basic sills that intruded the granite early during or before all deformation. The two granite sills are separated by a thin (one to a few meter-thick) lamina of psammopelitic rocks.

Both granites can be traced around large parts of the synform (Fig. 3). All along this distance (i.e. about 30 km along strike) these granites remain in the same lithostratigraphic level.

The fact that one of these granites is intruded by basic sills and the other is not suggests a temporal relationship of basic and granitic intrusives. Some of the older granite sills seem to post-date the intrusion of the basic rocks. Yet, this has not been proven by radiometric dating.

Kröner *et al.* (1991) discussed whether this lithostratigraphic pile is a primary sequence or a tectonostratigraphic succession. They argued that the sequence resulted from complete transposition of structurally complicated rocks, so that all primary relationships were destroyed. As we show later, it is clear from field evidence that the succession is strongly deformed. Yet there are a number of hints that the rocks have suffered no complex deformation prior to the granulite facies event, so that the succession still reflects primary relationships. Locally isoclinal folds may repeat parts of the succession, but repetitions by large isoclinal folds

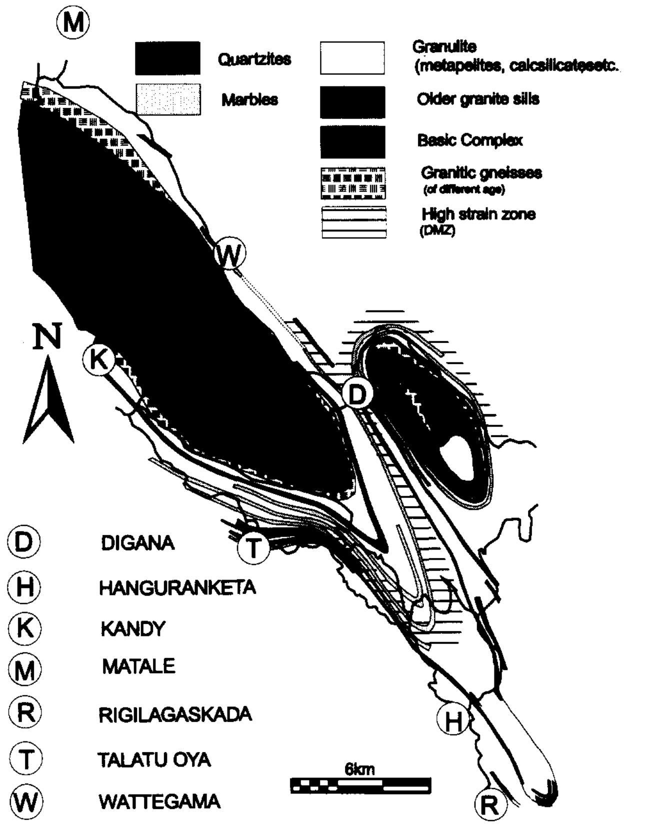

Fig. 3. Geological map of the Dumbara Synform and the neighbouring Hulu Ganga Synform.

can be excluded, as marker horizons like the marble beds and the granites have diagnostic features with individual characteristics.

STRUCTURAL EVOLUTION AND METAMORPHISM

D_1: *first strong stretching and flattening*

A first cleavage (S_1) is everywhere parallel to compositional layering. Flattening parallel to this first cleavage is strong but the amount is difficult to measure, as reliable strain-markers are missing and there has been later overprint by the big synforms. Together with S_1, a pronounced stretching lineation L_1 is formed. In psammopelitic rocks, S_1 is marked by (001) of biotites, flat sides of quartz, (100) of hornblendes, and L_1 by long axes of quartz grains or recrystallized aggregates and hornblende-*c*-axes. In quartzites, the individual quartz grains show no shape preferred orientations, as these were destroyed by extensive post-deformational grain growth. Yet in layers where grain growth was inhibited by feldspars or heavy minerals (which are enriched in layers, still marking sedimentary bedding) a shape preferred orientation in S_1 is visible. Sillimanite usually is strongly oriented with *c*-axes parallel to L_1. In pure marbles, shape preferred orientation is obliterated by later grain growth, causing 'foam-cell-textures' to form. In more impure marbles, (100) of diopside and (001) of phlogopite mark S_1. The deformed granite sills show quartz ribbons with flat sides in S_1, and long axes in L_1. The same is valid for feldspar recrystallization aggregates derived from phenocrysts. Wherever one can clearly identify sedimentary bedding (S_0) and S_1, they always are strictly parallel. We never found first folds folding S_0, to which S_1 is the axial plane cleavage. Most of the strong and penetrative deformation occurred close to peak metamorphic conditions. Garnet–pyroxene assemblages were stable throughout this deformation. There are only a few structures which point to a start of this deformation during prograde metamorphism, already at greenschist-facies conditions (Voll & Kleinschrodt 1991). Dolomite layers within marbles sometimes show seams of calc-silicate minerals, protruding in a stylolitic manner into dolomite layers. These could be relics of early pressure solution. Boudinage of dolomite layers between ductile deformed calcite layers indicates temperatures still below about 450°C (Voll 1968), conditions at which dolomite behaves brittley and calcite is highly ductile. No calcsilicate rims are found at the boudin fractures, precluding an interpretation of the calcsilicates as reaction rims.

In a few places inclusions in garnets have preserved a pre-existing crenulated foliation. This indicates that at least locally the prograde metamorphism was connected with rotational deformation. Relics of isoclinal folds in S_1 are extremely scarce. Where they occur they are commonly folded thin basic layers, which could also have been generated by simple flattening of dykes oblique to S_1.

The strong deformation also affected early pegmatitic dykelets, which intruded the rocks early or prior to D_1. They essentially consist of KNa-feldspar and quartz and commonly do not bear minerals like mica or hornblende. They seem to have suffered all the deformation together with the surrounding metasediments. Usually their thickness now is a few millimetres up to 3 cm and they provide one of the rare criteria to estimate the amount of flattening during D_1. Where KNa-feldspar phenocrysts are present within the dykelets, especially in thin dikelets, the thickness is much greater, as these acted as rigid inclusions and escaped flattening. Taking the diameter of this phenocrysts to reflect at least a minimum thickness of the undeformed dyke, and comparing it with the thickness of the dykelet where quartz and feldspars are recrystallized, a strain ratio of 1:20–1:25 is obtained. As the thickness of the original dyke may have been even higher than the diameter of the relict phenocryst, the real amount of strain could be even higher.

Another possibility for strain estimation is provided by the older granite sills, where they are rich in KNa-feldspar phenocrysts. In the Digana Granite the phenocrysts are strongly deformed and recrystallized, with few relicts. The recrystallized aggregates form markedly oblate ellipsoids, with a flat side in S_1, and long axes in L_1. Axial ratios are difficult to measure exactly, but commonly vary about X:Y:Z = 20–28:4–6:1, i.e. the strain ellipsoid is oblate, but not far from plane strain.

Close to Digana there is a zone of significantly increased strain, the Digana Movement Zone (DMZ) (Voll & Kleinschrodt 1991), in which stretching parallel to L_1 is strongly increased. Quartz ribbons of deformed pegmatoid dykelets have axial ratios up to 100:4:1. The DMZ can be traced around the southern hinge of the synform for a distance of 25 km with a thickness of up to 100 m.

Maximum strain is concentrated in a 10–15 m wide zone in the center of the DMZ. Foliations and lineations in the DMZ are strictly parallel to S_1 and L_1 everywhere, so that the increased deformation is a condensation of D_1- strain in discrete zones. Within the investigated area, the high strain zone does not climb to a deeper or higher lithostratigraphic level in any direction. It is found always between the Digana Granite, in parts affecting it, and below the next higher marble, so it is not a thrust ramp. It could be the flat part of a thrust system following a ramp/flat geometry and just a flat exposed in the synform area, which would mean that any ramps must be separated by more than 30 km. Or it could be the deeper, layer-parallel part of a large scale listric thrust.

Metamorphic conditions during D_1

Basic layers with garnet–orthopyroxene–plagioclase parageneses within the Dumbara Synform have been used for calculating pressure and temperature conditions. These minerals were stable during D_1-deformation, hornblende and orthopyroxene being oriented with their *c*-axes parallel to L_1. Using the calibrations of Harley (1984) to calculate temperature and Newton & Perkins (1982) to calculate pressure yield 850°C at 8kbar for core compositions of co-existing minerals (Fig. 4), similar to the results of Schuhmacher *et al.* (1990) and Schenk *et al.* (1991) for the central Highlands. These could reflect reequilibration conditions during cooling from even higher temperatures (Schuhmacher *et al.* 1990). No garnet–pyroxene parageneses were found in the layered basic intrusion in the core of the synforms. Pyroxene appears in ultramafic layers, sometimes with co-existing orthopyroxene and clinopyroxene pairs. The mineral parageneses found in the basic rocks in the core of the synforms are compiled in Table 1.

Relict fabrics found close to Wattegama reveal that most of the hornblendes probably formed from pyroxenes. The rock is the least deformed found throughout the basic intrusion in the Dumbara Synform. The mafic clusters have aspect ratios of about 5:2:1. In the cores of the mafic clusters orthopyroxene is still present as recrystallized grains, completely surrounded by a mantle of amphibole (Fig. 5a). In more strongly deformed rocks, which still show flaser fabrics, the pyroxenes are completely replaced. Most of the rocks are even more strongly deformed, so that the flaser fabric is replaced by an equigranular hornblende–plagioclase gneiss.

Thus, the rocks reached granulite facies conditions without a strong or complicated deformation, as most of the D_1 structures formed during granulite facies conditions. Furthermore, the absence of a hydrous fluid phase seems to be a necessity, because there is no sign of partial melting during or prior to D_1. Geochemical investigations by Pohl & Emmermann (1991) also prove that the rocks have preserved their primary signatures and show no restitic character.

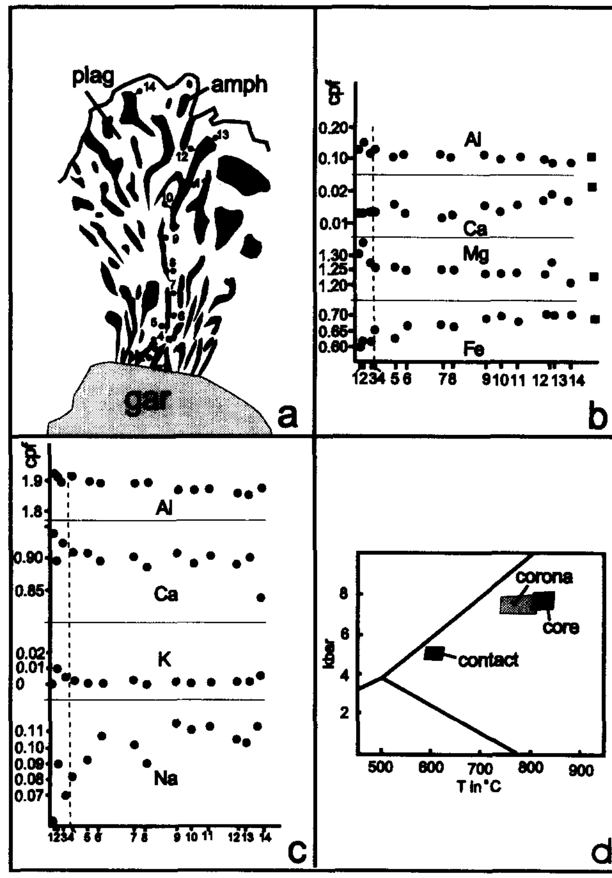

Fig. 4. (a) Sketch of orthopyroxene–plagioclase corona around garnet. Location of measured points from b and c. (b) Orthopyroxene-composition in the corona, black squares at right mark core compositions of orthopyroxene outside the corona. (c) Plagioclase-composition in the corona. (d) *P/T* estimates according to Harley (1984) and Newton & Perkins (1982) for core composition of garnet and orthopyroxene outside the corona, from the outer margin of the corona and of contact parageneses (quartz is missing in the contact, so that pressure calculation according to Newton and Perkins gives minimum pressures).

Table 1. Parageneses found in the basic to ultrabasic rocks in the central part of the Dumbara Synform

Mineral parageneses in the Layered Basic Complex
grt–hbl–pl–bi–mag/ilm
hbl–pl–bi–mag
hbl–pl–mag
hbl–cpx–mag
hbl–opx–pl–mag
hbl–cpx–opx

D_2 *and* D_3

In the area of the Dumbara Synform we have not found large second or third generation folds. However, these are present in the eastern and southern parts of the Highlands, where they refolded S_1 and formed folds with axes parallel (F_2) and normal (F_3) to the stretching direction (Cooray 1962, Berger & Jayasinghe 1976, Voll & Kleinschrodt 1991) during granulite facies conditions.

D_4: *large scale upright folds*

The spectacular large synforms and antiforms in the central Highland Complex are the result of D_4. The Dumbara Synform is one of the largest of these synforms. Folding is associated with a cleavage, S_4, which is penetrative in the central parts of the synform and becomes less intense in the outer, deeper levels (Fig. 6). S_4 is absent specially in the south, where the fold axis is nearly horizontal and the fold is open. Generally, S_4 is nearly vertical in the center of the synform, is steeply dipping to the west on the west-limb, to the east on the east-limb (Fig. 6d).

S_4 is associated with a reorientation of minerals, which acquire shape preferred and lattice preferred orientations. Quartz rods with axial ratios of about 25:6:1 are formed in quartzitic rocks during D_1, and are preserved in pegmatoid layers and in quartzitic rocks, which have enough feldspar to prevent the quartzite from exaggerated grain growth. In both rocks the rods are surrounded by feldspar grains, so that their shape is preserved during static annealing. Commonly one rod is formed from one quartz grain, and where two rods are in contact they are often formed from the same grain, which involved extensive grain growth. In places where S_4 starts to overprint the S_1-fabric (e.g. in the hinge zone of the Dumbara Synform exposed along the new road at the southwest-shore of Victoria Dam reservoir), S_1 and the flattened quartz rods are crenulated by S_4. The hinges of these small folds are thickened and the limbs reduced, during continued deformation, so that the former quartz rod is replaced by a few smaller, isolated grains derived from the hinges, now with their flat sides in S_4. During the same deformation biotite (001), hornblende (100) and sillimanite (010) are reoriented in S_4. The granulite facies assemblages (garnet, clinopyroxene, hornblende and plagioclase; garnet, orthopyroxene, hornblende and plagioclase, in basic rocks) remained stable throughout this deformation. Orthopyroxene and clinopyroxene are also reoriented during D_4 with (100) in S_4 without any sign of retrogression (Kleinschrodt *et al.* 1991).

The core of the Dumbara synform is formed by basic to ultrabasic rocks of variable composition which we identified as part of a large layered intrusion (Kleinschrodt *et al.* 1991, Voll & Kleinschrodt 1991). In the basal part of the intrusion, where magmatically layered rocks are common, S_1 is dominant and parallel to the layering. Approaching the core region, S_4 becomes more pronounced and S_1 is strongly folded between S_4 planes. Folds develop in the cm–dm range.

Early during D_4 a high amount of melt intruded the rocks in the central synform, where basic to dioritic hornblende–plagioclase gneisses predominate. These melts often intrude parallel to S_4. Especially in the area around Katugastota, north of Kandy, the amount of these melts can make up more than 50% of the rock volume. The existence of fragments and 'schollen' in the melts proves they were highly mobile. Intrusion of melts occurred prior to, during and at the end of D_4, but no clues for a melt-supported deformation were found. All deformation was solid state, being combined with formation of S_4 in the intrusives. Depending on their orientation (i.e. if they are intruded parallel to S_4 or parallel to S_1 or anywhere in between), the intrusives are folded, thinned or thickened. Intrusives parallel to S_4 are thinned, those parallel to S_1, which is about normal to S_4 in the central synform, are folded or buckled.

Frequently layers differing only slightly from neighbouring layers, such as in hornblende-content, are boudinaged. These layers are commonly between 10 cm and 1 m thick. Extension marked by separation of the boudins parallel to the fold axis is about 10–15%. Often boudinage is asymmetric and associated with left-lateral shearing in the steeply dipping limbs. Aplitic to pegmatoid mobilisates, commonly bearing biotite, occupy the voids and cracks in the extension zones. Still these are slightly affected by D_4, with biotites reoriented in S_4, proving active extension parallel to the fold axis during folding.

Metamorphic conditions during and after D_4

Within the basic rocks filling the core of the synform, P/T estimates are lacking because no sensitive parageneses, especially no garnets, are found in these rocks. Metabasics in the underlying psammopelitic sequence have developed parageneses including garnet, plagioclase and orthopyroxene. This mineral assemblage is stable throughout the formation of S_4. Core compositions of these minerals still reflect peak conditions (850°C, 8 kbar), even though these rocks are affected by S_4 (orthopyroxene reoriented with (100) in S_4 and garnet developing tails in S_4). Evangelakakis *et al.* (1991) provide arguments that prior to the formation of the synforms temperature had already decreased to 700–750°C at the level of the DMZ. Schuhmacher *et al.* (1990) found that, based on garnet reaction textures, a temperature decrease from about 850°C to 700–750°C occurred essentially isobarically. We attribute this isobaric cooling to the time between D_1 and D_4.

In garnet–amphibolites intercalated in the psammopelitic sequence below the basic complex (Fig. 2), garnets have alteration coronas with symplectitic intergrowth of orthopyroxene and plagioclase (Figs. 4 and 5b). Replacement of garnet by these coronas is very variable. They may form just thin rims, but in some layers garnet is totally replaced. The orthopyroxene forms stems with 10–15 μm diameter at the outer rim to about 5 μm diameter near the contact to the garnets or in the central parts of totally replaced garnets. This indicates their formation during decreasing temperature (Voll 1982). Within the symplectitic rim, orthopyroxene and plagioclase have preserved a compositional zona-

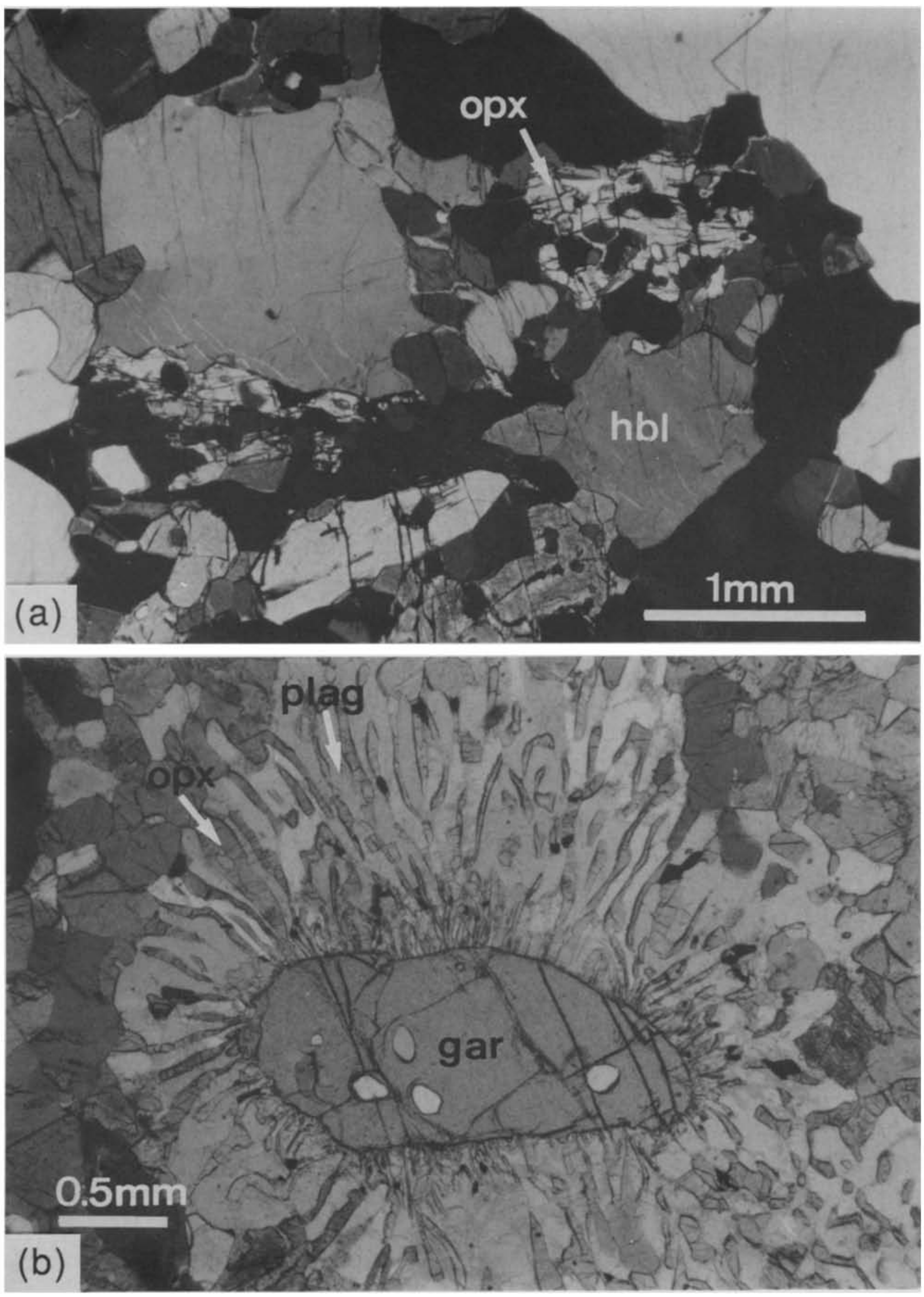

Fig. 5. (a) Orthopyroxene preserved in the core of mafic flaser structures in less deformed parts of the basic complex. The orthopyroxene is recrystallized during D_1, no magmatic relics are preserved. (b) Metabasite with posttectonic (post-S_4) plagioclase/orthopyroxene coronas around garnets.

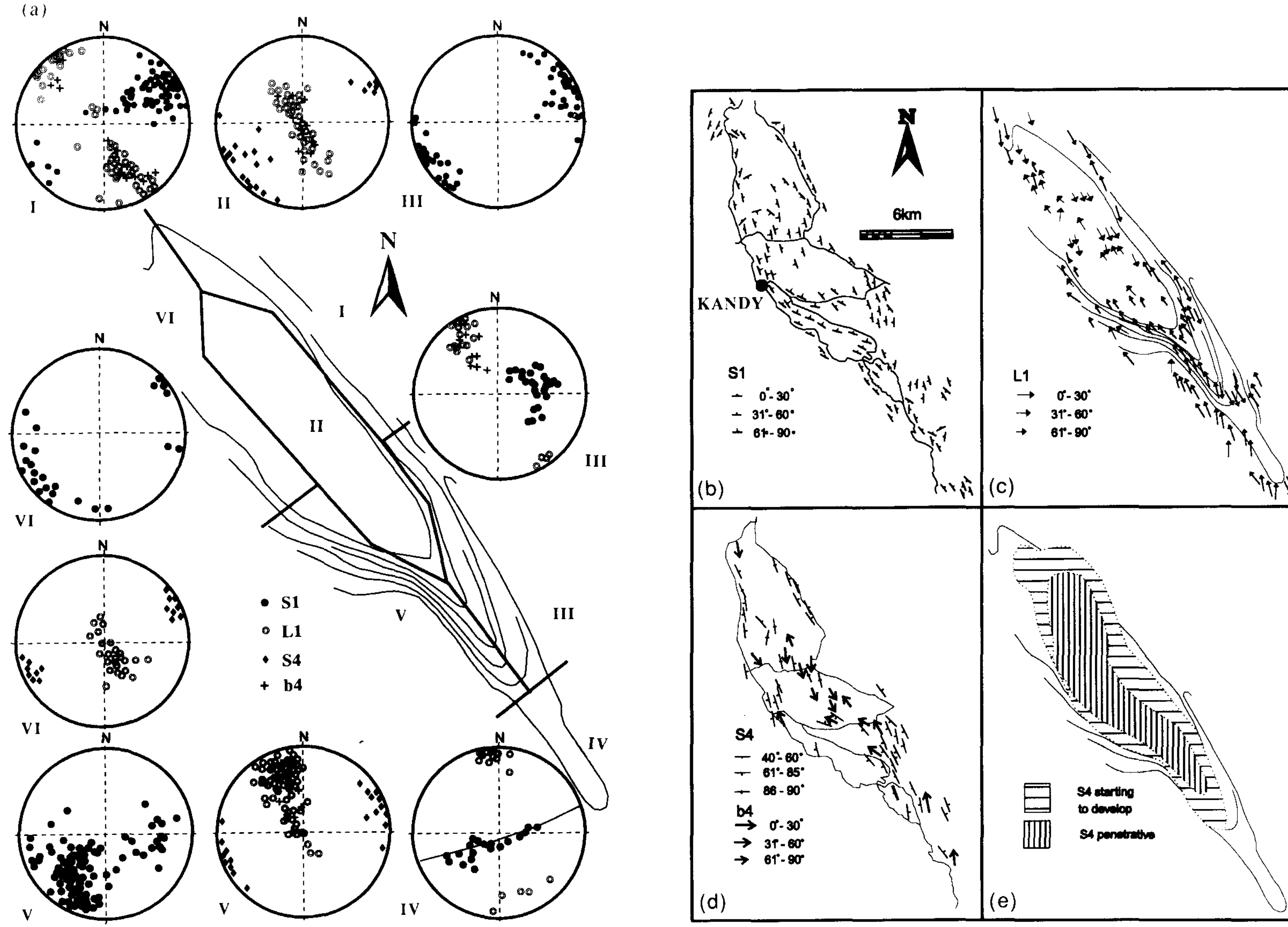

Fig. 6. (a) Tectonic data from the Dumbara Synform (Schmidt net, lower hemisphere). (b) First foliation orientation, S_1. (c) First stretching lineation orientation, L_1. (d) Foliation orientation, S_4. (e) Distribution of S_4.

Table 2. Compilation of U–Pb zircon ages and Rb–Sr-biotite ages from the Dumbara Synform or adjacent areas in the Highland Complex of Sri Lanka according to data published by Hölzl *et al.* (1991)

Radiometric ages—Highland Complex (data from Hölzl *et al.* 1991)		
Sedimentation:	Minimum age:	1894 ± 22Ma
	Maximum age:	unknown
	U–Pb-upper intercept of clastic zircons:	1982 ± 290
		2063 ± 133
		1868 ± 87
	Rb–Sr whole rock age (protolith?)	2384 ± 48
		2287 ± 108
Intrusives:	(U–Pb upper concordia intercepts)	
	'Older Granites'	1900–1850Ma
	'Pink feldspar granite' (pre-D_4)	580 ± 7Ma
	Post-tectonic granites	about 550Ma (558 ± 14, 552 ± 8, 550 ± 3)
Metamorphism:	Peak metamorphism	
	(U–Pb lower intercepts)	610–590Ma (608 ± 3, 591 ± 40, 611 ± 22)
	Cooling ages	
	Biotite Rb–Sr	470 ± 440Ma

tion (Figs. 4b & c). This zonation seems to reflect the compositions acquired at the reaction front with garnet. In the orthopyroxene stems, the increase of Al and Mg content and the decrease in Ca and Fe content from rim to core follow a linear trend. Slight compositional changes are evident in the plagioclase component with an increase of Ca and Al towards the core and a correlated decrease in Na. This also indicates the formation during continuously decreasing temperature. Clearly there is no step in the compositional evolution, which is consistent with the continuous decrease of the width of the symplectite phases. Together this argues for undisturbed cooling during their formation. Diffusive exchange after the reaction front had stopped its progress into the garnet is limited to a 15–20 μm wide zone at the reaction front. Calculation of pressure (Newton & Perkins 1982) and temperature (Harley 1984) from contact parageneses yields temperatures of 580–630°C at a minimum pressure of 5–5.5kbar (Fig. 4d).

At the outer margin of the corona, the composition of the orthopyroxenes is close to the composition of orthopyroxene outside the corona, which was affected by D_4 and reoriented in S_4. As these still yield temperatures higher than 800°C if one calculates temperatures using core compositions of the garnets, they have obviously not been adapted to the lower temperature during formation of the fold, probably because of the lack of a fluid phase.

All corona structures are completely undeformed (Figs. 4a and 5b). Sometimes a slight shape preferred orientation of coronas in S_4 is visible, but these are pseudomorphs after garnets slightly elongated in S_4.

The coronas therefore formed after the formation of the large synforms and reflect the start of a long period of slow static cooling from temperatures close to 700°C. The static grain growth destroyed most intracrystalline deformational features. In monomineralic layers (marbles, quartzites) exaggerated grain growth caused a strong increase in grain size. In other rocks, like calcite marbles, the calcites form very well developed foam-cell textures. In the outer parts of the coronas, pyroxene grain-boundaries have rearranged and the stems form idiomorphic crystal faces.

Late pegmatites

Late pegmatites intruded the rocks not only of the Dumbara Synform, but all over Sri Lanka. They cross-cut all structures, but are slightly deformed in a late compressive field, with the main compressive axis about normal to the axial plane of the synform, suggesting their intrusion at a late stage during its formation. Usually deformation is concentrated at the pegmatite rims and is manifested by recrystallization of quartz and feldspar grains. In the inner parts, coarse pegmatite minerals (feldspar, quartz, biotite, hornblende, scapolite, allanite, etc.) are well-preserved. The pegmatites mark the last intrusive event in the Highland Complex and their slight deformation is the very last evidence of ductile deformation in the Highland Complex.

TIMING OF MAGMATIC AND METAMORPHIC EVENTS

Data from recent geochronological work on the rocks of Sri Lanka are compiled in Table 2. Zircons from one of the older granite sills occurring in the Dumbara Synform yielded U–Pb ages of 1894 ± 22 Ma (Hölzl *et al.* 1991). Consequently, the sediments intruded by these granites must be older. Based on U–Pb zircon dating, Hölzl & Köhler (1989) and Hölzl *et al.* (1988, 1991) found evidence for high-grade metamorphism between 610–550 Ma with peak metamorphism around 610–600 Ma (lower concordia intercepts). From their data they see no arguments for previous metamorphic events as postulated by Crawford & Oliver (1969), De Measschalck *et al.* (1990) and Cordani & Cooray (1989). The

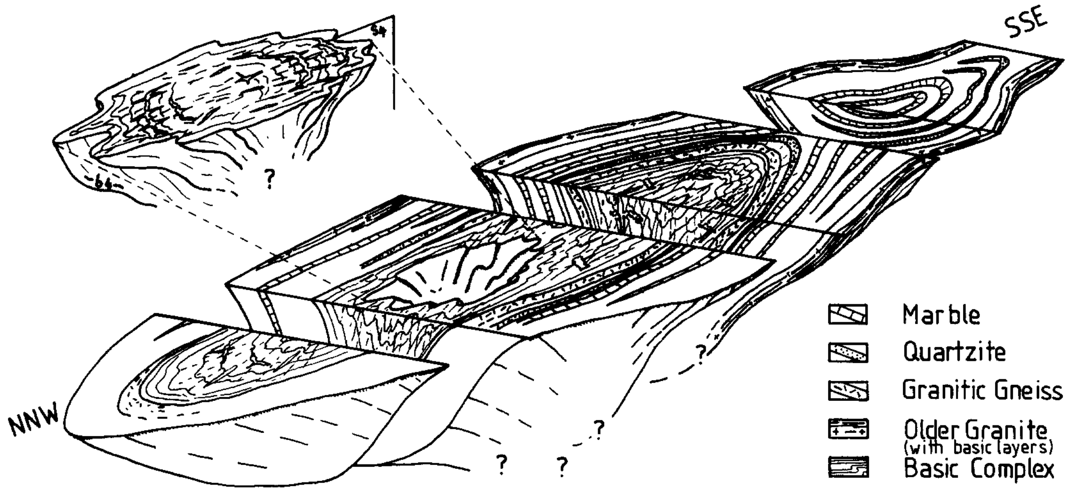

Fig. 7. Schematic block diagram of the Dumbara Synform, not to scale. View from the northwest towards southeast. Note the change in fold style from the outer part to the center.

meaning of single zircon ages around 1100 Ma (Kröner *et al.* 1987), which was interpreted in the original paper as a separate metamorphic event, is unclear and the interpretation was retreated (Baur *et al.* 1991). These dates are now interpreted as discordant ages.

Late granites, which are less deformed than both the sediments and the older granites, yield ages of 560–600 Ma (Hölzl & Köhler 1989). One of these granites, the 'pink feldspar granite' is situated within the Dumbara Synform, directly underneath the Basic Complex. It yielded an age of 580 ± 7 Ma. Cooling after granulite facies metamorphism to under 300°C is dated by Rb–Sr-biotite ages between 474 and 439 Ma (Hölzl *et al.* 1991).

Thus, there is a time period of at least 1200 Ma between sedimentation and metamorphism. Most of the deformation seen in the rocks was acquired during granulite facies conditions and the rocks were statically annealed. Deformation had ceased, when the rocks were still under high temperature. Cooling of the rocks after the granulite facies metamorphism seems to be very slow, as indicated by the Ordovician (480–440 Ma) biotite cooling ages. At the base of the Highland Complex an average cooling rate of 3–4°C/Ma was derived for the temperature interval between 650°C and 300°C (Kleinschrodt 1994).

FOLD GEOMETRY OF THE DUMBARA SYNFORM

In map view (Fig. 3) the Dumbara Synform forms an elongate ellipse with a long axis of about 40 km, a short axis of about 8 km. It is embedded in an array of en échelon arranged synforms and antiforms, which are well exposed in the central Highlands (Berger & Jayasinghe 1976). The fold-axis trends about NNW/SSE. The plunge of the fold axis steepens towards the center of the synform (Figs. 6a and 7) and becomes subvertical in the central part. The fold axis is parallel to the stretching lineation. It is nearly impossible to get an estimate how much of the total stretching seen in the rocks was acquired during D_4 deformation. The boudinage structures formed during D_4 document that during F_4 folding stretching parallel to L_1 continued. No new stretching lineation was developed. This proves that although folding happened after significant cooling (i.e. after a significant time interval), the kinematic framework did not change during this cooling.

Penetrative S_4 cleavage is lacking in the outer parts of the Dumbara Synform, especially on the west-limb and in the southern hinge zone (Fig. 6e). S_4 cleavage becomes penetrative towards the central parts of the synform above the level of the second marble horizon (Figs. 2 and 6e). This contrasts with Berger & Jayasinghe (1976), who report a 'lack of axial plane cleavage' in these folds. S_4 becomes the dominant schistosity in the central parts of the synform and forms a slightly convergent fan across the synform (Fig. 6d). Also on S_4 planes the stretching lineation L_4 (marked by hornblende *c*-axes) is still parallel to L_1.

Towards the core of the synform the fold shape changes. The outer parts fit into class 1 of folds of Ramsay (1967), with slightly converging dip isogons, as is expected for flexural slip or flexural flow folds. The central parts of the synform show significant hinge thickening, and the fold geometry changes to that of similar folds with parallel or even divergent dip isogons. This strong thickening of the hinge zones in the core of the synform may be one reason for the vast extent of the layered basic intrusion there. Folding in this core region is different from that in the outer, deeper level (Fig. 7). In the outer parts, very little parasitic folding is developed, and the layers are openly undulating with wavelengths of 5–20 m. Approaching the basic core, intense folding with a dominant wavelength in the dm to meter-scale starts. The fold axes become steeper and curve. If the fold as a whole acts like a flexural flow fold, it is clear that the inner parts of the fold must be more compressed. This stronger compression produced the S_4 cleavage and modified the earlier S_1/L_1 fabric. With stronger compression, the central parts, where S_1-layering is nearly normal to the compression axis, are intensely folded. Thus the structural differences between outer and inner parts of the folds are caused by the fold mechanisms and do not reflect differences in pre-D_4 deformation as argued by Kriegsmann (1991).

Hinge-thickening and intense S_4-cleavage development provided pathways for melts to intrude. As already mentioned, the core regions of the Dumbara Synform is impregnated by quartz–feldspar-rich melts during D_4. These melts migrated along S_4 planes, but still were deformed by D_4. As the total amount of melts can be more than 60% of the rock volume, the field-appearance of the basic rocks is changed dramatically.

One marked feature of these folds is that their axes are parallel to the stretching direction. This stretching direction is a continuous element during all deformation in the Highland Series. The main difference between D_1 and D_4 is that, while the extension direction remains constant in orientation, the intermediate and contractional axes are exchanged. During D_1, in a plane normal to the extension direction, the shortening direction is about vertical, leading to flattening of the pre-existing horizontal layering. During D_4 this vertical shortening is less significant than E–W directed shortening.

DEFORMATION MECHANISMS

D$_1$

During the formation of D_1 foliations and lineations, it is remarkable that there was very little viscosity contrast between layers of very variable composition. Quartzites, psammopelitic layers, granitic sills and even basic sills are deformed together, without significant strain partitioning. Basic sills in granitic sills are not boudinaged at all, nor are basic sills in quartzites. Exceptions are boudinaged or fractured basic sills or dolomite layers in marbles, but most probably this deformation occurred during lower temperature conditions on the prograde path of metamorphism. During D_1 no hydrous fluid phase was involved, as no pressure shadows are found around rigid porphyroclasts. Furthermore, at higher temperatures and pressures a hydrous fluid phase would cause instantaneous partial melting. Minerals like feldspar and quartz acquire a distinct lattice preferred orientation, arguing for dislocation creep mechanisms being active. If this were the only active deformation mechanism, there should be significant viscosity contrasts between the lithologies involved in the synforms, as flow strengths of quartzites, granites and basic rocks are quite different during dislocation creep. Grain-size sensitive deformation mechanisms probably contribute to the total deformation as coarser-grained layers are harder than finer-grained layers of about the same composition.

Especially remarkable is the lack of viscosity contrast between the layered basic intrusion in its basal parts and the underlying rocks with granitic composition. It appears that, during high-grade metamorphism (in contrast to lower temperature deformation), there seemed to be no marked contrast in flow strength between 'granitic' and 'basic' rocks.

D$_4$

Evidence for grain-size sensitive mechanisms are also found in D_4-structures. Coarser grained hornblende rich layers within the layered part of the basic complex are boudinaged between similar, but finer-grained layers. If dislocation creep were the dominant deformation mechanism, the coarser-grained layers should be the more ductile ones.

The absence of parasitic folds in the outer part of the synform with numerous layers on all scales proves the low viscosity contrast between these layers, even though temperatures had decreased already to 700–750°C at the time of deformation.

During D_4, boudinage and pinch and swell structures are more common in compositionally layered sequences. Viscosity contrasts reappear at these conditions.

All the minerals show distinct crystallographic preferred orientations. Shape preferred orientation is only preserved where post-tectonic grain growth is restricted. In these cases quartz is commonly the weakest phase, as it forms ribbons in the matrix dominated by feldspar. Hornblende and plagioclase have intermediate and about equal strength, and pyroxenes are the strongest phase. This argues for dislocation creep being the dominant deformation mechanism under these conditions.

STRUCTURAL EVOLUTION MODEL

One of the puzzling results is that metasedimentary rocks reached granulite facies conditions without complicated deformation, so that original sequences are still preserved. The original thickness is reduced during the formation of the granulite facies foliation and stretching lineation and primary discordancies between sediments and intrusives may be obliterated by the strong deformation. Kröner *et al.* (1991) regard the compositional layering seen in the rocks as a result of extreme deformation and complete transposition of complicated pre-existing structures.

However, our detailed mapping within the synforms reveals the following:

(1) A lithostratigraphic sequence can be compiled, which is comparable between neighbouring synforms. Characteristic horizons reappear in identical positions, often with individual characteristics, which exclude repetition of layers by isoclinal folding (for example the marble horizons in the Dumbara Synform).

(2) Even thin layers, such as the garnet-bearing layer near the base of the layered intrusion can be found within neighbouring synforms in identical positions.

(3) Sedimentary characteristics, such as heavy mineral layers in quartzitic rocks, are preserved without complications by folding.

In some areas isoclinal D_2 or D_3 folds are more common than in the Dumbara Synform and repeat parts of characteristic sequences.

One considerable contribution to crustal thickness is provided by the numerous granitic and basic sills in the Highland Series. The layered basic intrusion at the top of the sequence now has a thickness of at least 300 m (as exposed in the neighbouring Hulu Ganga Synform), but its original thickness may have been much higher (Kleinschrodt *et al.* 1991). Numerous granite sills were found south of the big synforms, in the area between Kandy and Nuwara Eliya (Voll & Kleinschrodt 1991), which exceed the amount of metasediments. For most of these rocks there are no radiometric age data available. As the numerous older granite sills are much older than the high-grade metamorphism (Table 2), they cannot be related to crustal thickening in causal relation to high-grade metamorphism. Up to now we do not know if this also counts for the other magmatic sills or if some of them contribute to thickening and maybe also to temperature increase during the Panafrican event.

The basic rocks filling several large synforms (usually termed hornblende-gneisses and hornblende migmatite gneisses, e.g. Geological Map of Sri Lanka 1982) have been one of the main problems in Sri Lankan geology. We have earlier shown (Kleinschrodt *et al.* 1991, Stosch 1991, Voll & Kleinschrodt 1991) that these rocks represent the relicts of a large layered basic intrusion, which are preserved just in the cores of the synform, as they form the highest part of the lithostratigraphic succession. A small section of roof rocks occurs only in the Hulu Ganga Synform (Kleinschrodt *et al.* 1991). We do not have clear evidence for the depth at which these basic rocks were intruded. We can only say that they intruded prior to or early during D_1 deformation. The absence of contact metamorphism and partial melting in the neighbouring rocks may be evidence that these rocks were already dry granulites at the time of intrusion.

Basic intrusions like those found in the synforms of central Sri Lanka may be quite common in the granulite facies lower crust; possibly, they have not yet been interpreted in this way because of strong reduction of thickness by deformation and metamorphic overprint.

Most of the D_1 structures formed during high grade conditions. D_1 must be directly related to the thermal perturbation causing high-grade metamorphism. The strong deformation during D_1 produced a pronounced stretching lineation. From D_1 structures alone it is impossible to decide whether this is connected with an overall extensional or collisional regime, which both could produce high-grade metamorphism (England & Thompson 1984, Sandiford & Powell 1986). The direction of stretching is maintained throughout D_1 to D_4. D_2 and D_3 seem to be shortening/collisional structures, as they clearly cause thickening of the lithologic pile. If D_1 were an extensional event, it would have to be followed by an exactly reversed movement direction during subsequent shortening. Thus, the prominent stretching lineation probably does not mark the extension direction in an active extension regime, but marks the transport direction in a shortening/collisional regime. The Digana Movement Zone cannot be responsible for crustal thickening, as it clearly post-dates peak conditions and does not apparently duplicate the section. So we have to speculate on the presence of thrusts above the presently exposed surface level (now eroded), acting during or prior to D_1.

Exhumation of the Highland Complex followed after thrusting of the Highland Complex along a subhorizontal thrust plane over the Vijayan Complex (Kleinschrodt 1994). This happened after significant, isobaric cooling to temperatures of about 700°C. The stretching lineation which formed during thrusting is parallel to L_1.

During all deformational events the kinematic axes remain constant. This and the geochronological data of Hölzl *et al.* (1991) argue for a monocyclic metamorphic/deformational event in the Highland Complex of Sri Lanka.

In the area of the Dumbara Synform, deformation ended after the formation of the Synform (D_4) at 700°C at a pressure of about 8 kbar. Static annealing caused a tremendous increase in grain size and erased most of the intracrystalline deformation features. Overthrusting at the base of the Highland Complex probably started at similar conditions and ended at a temperature of about 650°C and 6–7 kbar (Kleinschrodt 1994). Final uplift is not accompanied by deformation.

The homogeneous deformation across the boundary of the basic layered intrusion and the underlying well layered, compositionally variable rocks during D_1 and D_4 proves that there is no or very little difference in flow strength of these rocks. So if the lower crust really consists of a deeper part with basic granulites and a higher part with granitic granulites as often assumed for crustal modelling, these two parts will not show much difference in deformational behaviour, at least in the transition from middle to lower crust. Deformation mechanisms may be different during deformation with comparatively high natural strain rates, such as in shear zones.

These observations offer an explanation for the simple style of deformation in the lower crust of Sri Lanka, and maybe for the lower crust in general. If all different lithologies have about the same flow strength, this part of the crust will act as one more-or-less homogeneous layer. This style of deformation contrasts with deformation in the upper crust (greenschist and amphibolite facies conditions), where differences in flow strength caused by differing deformation mechanisms of the main rock-forming minerals cause numerous complications.

Acknowledgements—This study was supported by the Deutsche Forschungsgemeinschaft (DFG). Field work was performed during the Priority Research Projekt "Composition, Structure and Evolution of the Lower Continental Crust". Furthermore R.K. was supported by research grant Kl 851-1/2. Special thanks to P. W. and R. Vitanage for the warm hospitality we received at their home in Kandy and for their invaluable help. The manuscript was significantly improved by through reviews of C. Simpson and an anonymous reviewer.

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