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Strain patterns, décollement and incipient sagducted greenstone terrains in the Archaean Dharwar craton (south India)

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Abstract—The Archaean Dharwar craton is characterized by two greenstone successions: the > 3 Ga Sargur Group and the 3.0–2.5 Ga Dharwar Supergroup. Examples of both successions are described from the region of Jayachamarajapura where they are also distinguished by different tectonic patterns. The younger greenstones have undergone only minor deformation and are only slightly metamorphosed and so provide a good case study of the relative behavior of greenstones in relation to their granite–gneiss country rocks. A detailed structural analysis indicates two strain fields associated with two deformational episodes: D_1 and D_2. The D_1 episode produced dome-and-basin structures and affected merely the older greenstones and the gneisses. The mapped strain field is compatible with the hypothesis that it is associated with the development of diapiric-type gravitational instabilities. The D_2 episode affects only the younger greenstone belt, which has the overall geometry of a complex syncline. It is discordant over a complex of gneisses and older greenstones that was deformed during the D_1 episode. The base of the discordant cover sequence is tectonized and constitutes a décollement surface. Kinematic criteria at this surface have opposite sense and converge towards the belt axis. These structural features are interpreted in terms of progressive deformation compatible with the incipient development of a sagducting trough.

These results are consistent with those obtained from other parts of the craton, where the tectonic evolution appears to reflect mainly relative vertical displacements facilitated by the reheating of continental crust during two major Archaean tectonometamorphic episodes. Copyright © 1996 Elsevier Science Ltd

INTRODUCTION

The mechanisms of Archaean tectonics have been debated for decades (McGregor 1951, Goodwin 1981, Windley 1984). A basic question is whether today's geological phenomena, and especially the orogenic processes due to plate tectonics, are similar to those that prevailed during the Archaean (Glikson 1981, Kröner 1991). Some authors consider that the greenstone belts are relics of marginal basins squeezed between ancient continental margins (Anhaeusser 1975, Burke *et al.* 1976, Tarney *et al.* 1976, Groves *et al.* 1978, Drury *et al.* 1984). Others consider the granite–greenstone patterns (McGregor 1951) as resulting from the interference of crustal-scale folds (Myers & Watkins 1985, Myers & Kröner 1994) or sheets (e.g. Myers 1976, Bickle *et al.* 1980, de Wit 1982, Stowe 1984, Ralser & Park 1992), or else in terms of pre- or post-thickening extension (James & Mortensen 1992, Hammond & Nisbet 1992, Williams & Currie 1993, Kusky 1993, Passchier 1994). According to others, granite–greenstone patterns may be due to the sagging of supracrustal rocks into juvenile crust (Goodwin & Smith 1980) combined with the relative uplift of tonalitic trondjhemitic granodioritic (TTG) gneisses (McGregor 1951, Anhaeusser *et al.* 1969, Gorman *et al.* 1978, Glikson 1979, West & Mareschal 1979, Mareschal & West 1980). On the basis of the experimental studies of Ramberg (1967), Dixon (1975) and Dixon & Summers (1983), some workers have recently claimed a clear distinction of deformation patterns into diapiric strain fields with greenstones on the one hand and granite–gneiss on the other (Bouhallier *et al.* 1993, 1995, Jelsma *et al.* 1993). The driving force behind the vertical tectonic movements are to be sought in (1) the reversed density gradient between supracrustal rocks and material of the juvenile crust and (2) the high degree of partial melting of the TTGs induced by the overall reheating of large segments of the crust.

The structural study of the relations between the greenstones and the granite–gneiss terrains (TTG) should help understand these mechanisms (Windley & Bridgwater 1971, Sutton 1976, Gorman *et al.* 1978, Platt 1980, Park 1982). It should be noted that systematic surveys of strain fields (variations in the local characteristics of the finite strain) are still few in number, but that they are particularly pertinent to deal with this question (Schwerdtner 1990, Choukroune *et al.* 1995).

In the Dharwar craton (Karnataka State, Southern India), the excellent outcrop of the Archaean crust facilitates the detailed structural analysis of granitoid/greenstone relations and the study of the deformation affecting these terrains. Two known occurrences of greenstone belts, both located in the Jayachamarajapura

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area (abbreviated to J. C. Pura), were selected for this study. Their geometric relations can be easily observed (Venkata Dasu *et al.* 1991).

The present study presents (1) the results of field mapping of strain patterns and (2) an analysis of the kinematic criteria that were systematically measured at the base of the greenstones. Since these diachronous sequences have undergone separate histories, the results thus obtained are particularly useful in terms of structural evolution and behavior of greenstones in relation to the foliated crystalline country rocks. The results of this study allow a discussion of current models of granite–greenstone tectonics and the behavior of the continents during the Archaean.

GEOLOGICAL BACKGROUND

The Dharwar Craton (Fig. 1)

Typical bimodal Archaean lithological associations can be easily recognized in the Dharwar craton. The 'Peninsular Gneisses' display the petrological characteristics of tonalitic, trondjhemitic and granodioritic assemblages (TTG). They form the major part of the Dharwar crust, being formed between 3.3 Ga (Beckinsale *et al.* 1980) and 2.5 Ga (Friend & Nutman 1991). The supracrustal rocks have been subdivided into two groups: the Sargur Group and the younger Dharwar Supergroup (Ramakrishnan *et al.* 1976). U–Pb dating performed on detrital zircons from Sargur supracrustal rocks yields ages ranging from 3.0 to 3.3 Ga (Nutman *et al.* 1992), whereas magmatic zircons coeval with belt formation (Holenarsipur area) have been dated at 3.3 Ga (Peucat *et al.* 1995). Whole-rock isochrons obtained from the felsic volcanics of the Dharwar Supergroup give ages ranging from 3.02 to 2.52 Ga (for details, see review in Peucat *et al.* 1995). The end of the Archaean evolution of the craton is characterized by the accretion of a huge quantity of granitic rocks, partly of mantle origin, which makes up the Closepet granitic batholith (Jayananda *et al.* 1995). This activity took place in a strike-slip fault context (Drury & Holt 1980, Jayananda & Mahabaleswar 1990). The Closepet Granite has been dated at about 2.5 Ga (Friend & Nutman 1991).

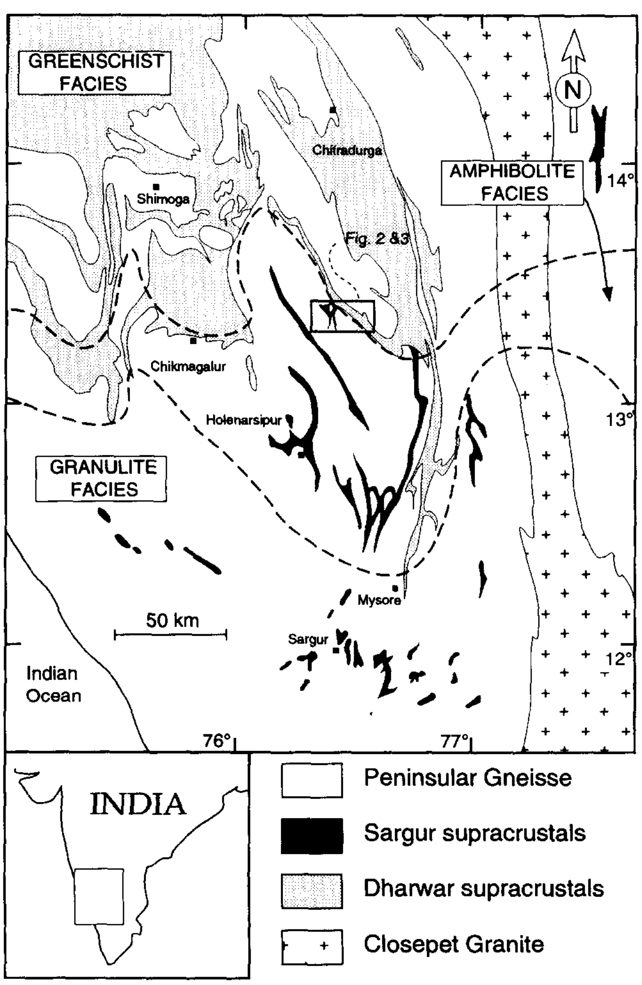

Fig. 1. Geological map of the western Dharwar craton with location of the studied area.

As a whole, the Dharwar craton gives a representative picture of the continental crust at the end of the Archaean. Indeed, a variation in the grade of regional metamorphism is observed along a N–S axis from greenschist to granulite facies (Raase *et al.* 1986). Pressures vary from 2–3 kbar in the north to 7–8 kbar in the south (Newton 1990). These east to west isograds are clear evidence of the conditions prevailing during formation of the granitoids composing the Closepet batholith; the temporal and spatial relations existing between migmatisation and charnockitization in the southern part of the batholith have indeed been noticed for some years (Pichamuthu 1961). Moreover, the age of the granulite facies metamorphism (2.51 Ga) which affects the material of the Closepet batholith, which was itself formed at around the same period, (Friend & Nutman 1991), confirms this close relationship (Peucat *et al.* 1993).

Sargur/Dharwar controversy

Rocks of the Dharwar Supergroup have been clearly identified in the northern half of the craton, in the form of a large basin. This basin is found in several greenstone belts in which the overall deformation is modest (Chadwick *et al.* 1981, 1985, 1989). In this northern part of the craton, it is easier to distinguish the two sequences and unconformities between them have been identified (Venkata Dasu *et al.* 1991). These unconformities separate the Sargur rocks and granite–gneiss terrains, which have undergone amphibolite facies regional metamorphism, from the overlying volcano sedimentary Dharwar successions which have undergone metamorphic transformation to a lesser degree (Raase *et al.* 1986). Both of these supracrustal sequences have also been differentiated according to stratigraphic, lithological and structural criteria (Swami Nath & Ramakrishnan 1981, Chadwick *et al.* 1981, Viswanatha *et al.* 1982, Ramakrishnan & Viswanatha 1983, 1987).

This distinction is not so clear cut in the southern part of the craton because of the intense degree of metamorphic recrystallization. Besides, it is thought that supracrustal rocks from the Sargur group occur predominantly in high-grade terrains in the form of narrow belts with intense deformation (Radhakrishna 1984). Finally, some authors consider that all the craton supracrustal rocks belong to a single unit whose deformation is linked to one long-lived 'Dharwar orogeny' (Drury *et al.* 1984, Pichamuthu & Srinivasan 1984, Naha *et al.* 1986, Srinivasan 1988).

Geology of the J. C. Pura area

The region selected for this study is located in the centre of the craton, in the transition zone between the amphibolite facies and the greenschist facies zones (Fig. 1). There are four distinct major geological units (Venkata Dasu *et al.* 1991): the Peninsular Gneisses in a broad sense, the Arsikere Granite, the J. C. Pura belt and the Kibbanahalli Arm (Fig. 2). Most Peninsular Gneisses are migmatitic and locally show some magmatic trondjhemite facies (e.g. east of KNA). The Tiptur trondjhemite cropping out in the southeastern part of the studied area has been dated at 3.2 Ga (Rogers & Callahan 1988). The Arsikere massif consists of a potassic pluton intruding the gneisses (Subrahamania & Naganna 1972). The average age of this granitic body is 2.59 ± 0.12 Ga (Venkatasubramanian & Narayanaswamy 1974, Meen *et al.* 1992). The J. C. Pura belt mainly consists of ultrabasic rocks, in most cases serpentinized, as well as amphibolites. Detrital zircons contained in a quartzite have yielded U–Pb ages around 3.2 Ga (Ramakrishnan *et al.* 1994). Many intrusive pegmatites concordant with the stratification can be observed within this belt. Metamorphism in the J. C. Pura belt reaches amphibolite facies grade (Venkata Dasu *et al.* 1991). Mapping of this belt reveals its highly discontinuous and locally disrupted character, which contrasts with the simpler structural pattern of the Kibbanahalli Arm. The Kibbanahalli Arm is made up of a cartographic interdigitation of the Chitradurga belt (Fig. 1), appearing as a synclinorium whose discordance with the material of the J. C. Pura belt has been recently described (Venkata Dasu *et al.* 1991). Monomict conglomerates and quartzites can be observed, mainly at the base of the sequence, as well as basic to intermediate volcanic rocks. There are no felsic intrusions into this material, whose metamorphic grade is generally indicative of the greenschist/amphibolite transition (Venkata Dasu *et al.* 1991).

STRUCTURAL AND STRAIN DATA

The map of foliation trajectories in this area ($\lambda_1\lambda_2$ of the finite strain ellipsoid) shows two large domains with distinct strain patterns (Fig. 3a & b).

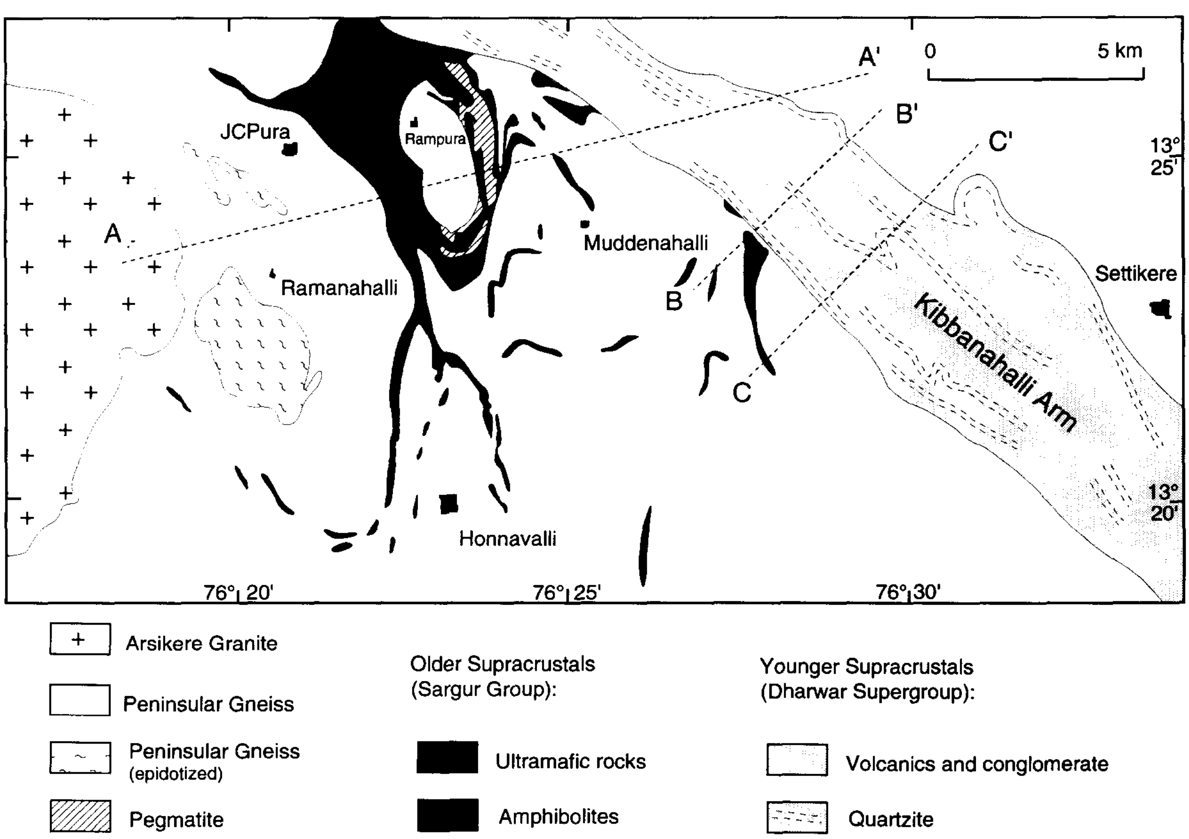

Fig. 2. Geological map of the J. C. Pura area (after Venkata Dasu *et al.* 1991 and this study). Lines AA′, BB′ and CC′ indicate the cross-sections presented on Fig. 8 (location on Fig. 1).

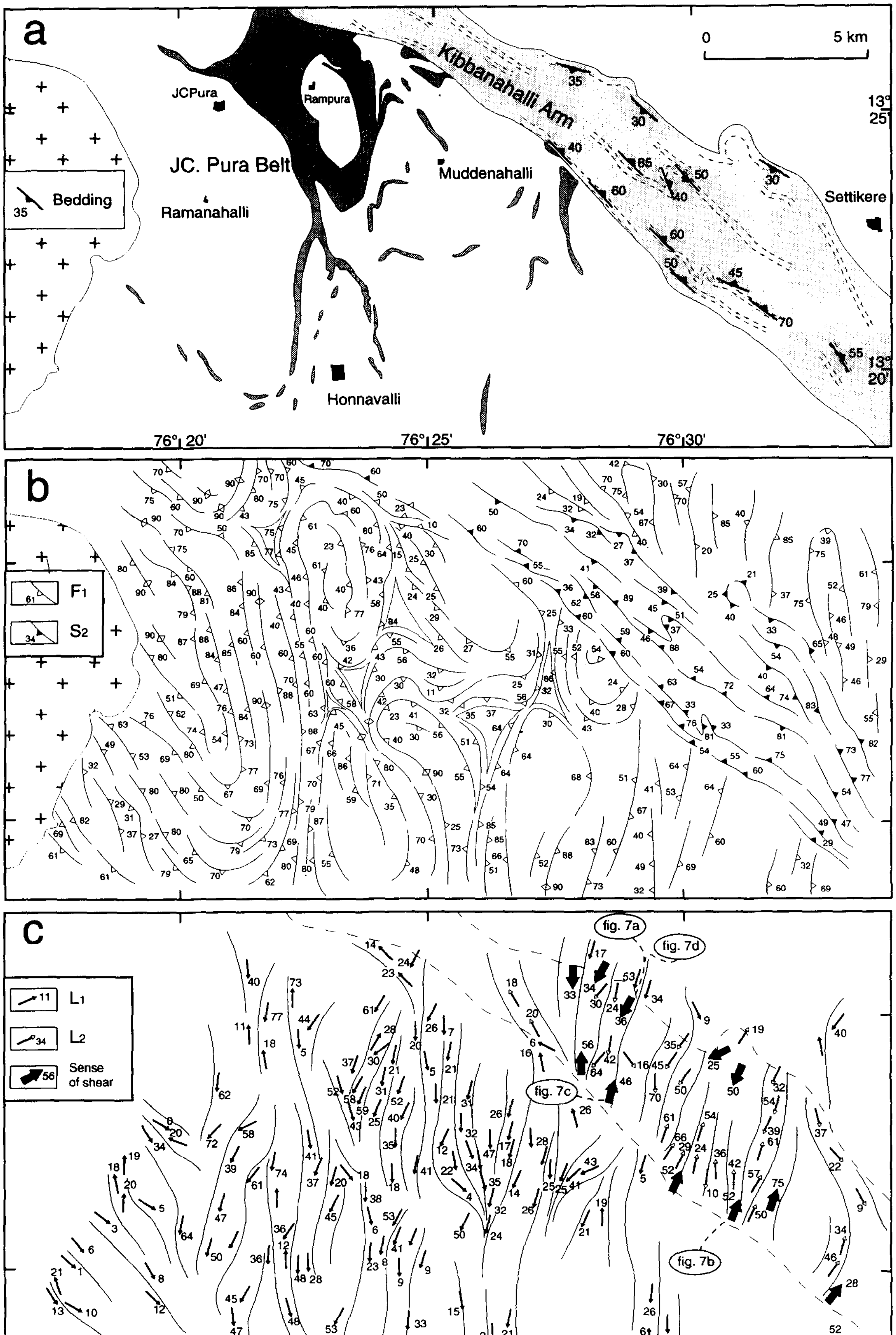

Fig. 3. Strain trajectory maps of the J. P. Pura area (location on Fig. 1). (a) Geological sketch map; (b) foliation trajectories ($\lambda_1\lambda_2$ plane); (c) stretching lineations (λ_1 axes) and shear criteria. The F_1 foliation triple points are shown.

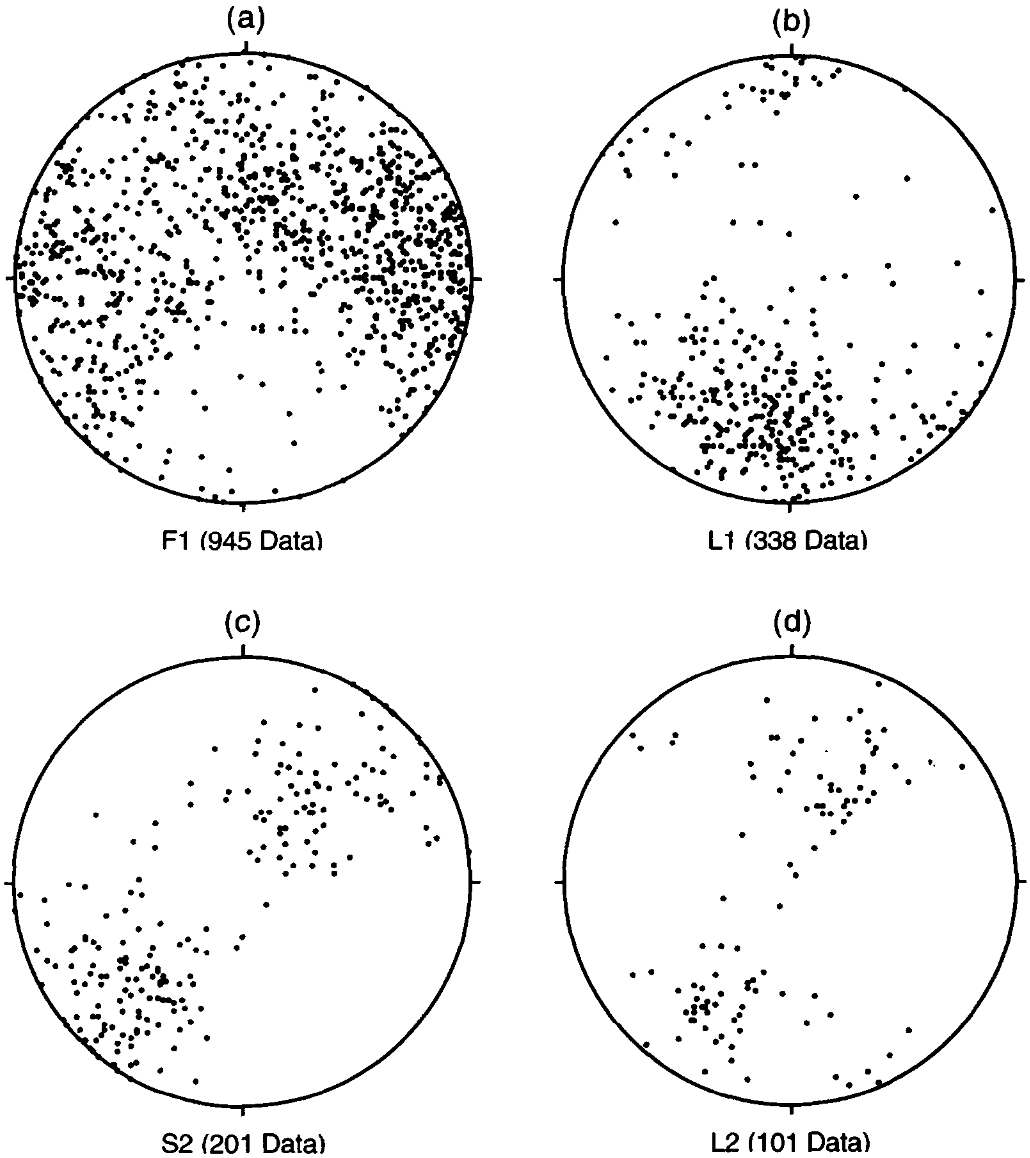

Fig. 4. Equal area stereoplots of strain fabrics. (a) Poles to F_1 foliations outside the Kibbanahalli Arm; (b) L_1 stretching lineations outside the Kibbanahalli Arm; (c) poles to S_2 schistosities inside the Kibannahalli Arm; (d) L_2 stretching lineations inside the Kibbanahalli Arm.

Basement rocks

The basement complex (gneisses and Sargur greenstones), which accounts for three-quarters of the mapped area, is characterized by various foliation trends (Figs. 3b and 4a); the trajectories of the regional foliation (denoted here as F_1) are curved to concentric and define dome-and-basin structures. Triangle zones corresponding to the interference zone between three flattening axes are observed between the concentric trajectories of the domes. These particular points—called triple points—have been attributed to the interference of diapiric migmatitic bodies (Brun *et al.* 1981, Bouhallier *et al.* 1993, 1995). It should also be noted that the contact of the Arsikere Granite clearly cuts across the trajectories described above, which implies that this granite postdates the F_1 basement foliations.

The F_1 foliation in the basement is systematically parallel to the lithological contact between the J. C. Pura belt rocks and the gneisses. The gneisses are generally located in the domes (e.g. Rammanahalli, Honnavali, Rampura, etc.) whereas greenstones are mainly to be found in the basins. To the east of the Kibbanahalli Arm, the foliation strike seems to follow a constant N–S direction, as observed, for instance, in the Tiptur trondjhemite (Fig. 3b).

The mapped outlines of the domes are elliptical rather than circular. Transcurrent shear zones a few km in length locally affect the steeply dipping foliations of the dome borders (notably at the southwestern and eastern edges of the Rammanahalli dome). The lineation generally shows a slight plunge. These shear zones are systematically injected with pegmatites. Analysis of the C/S structures in the pegmatites (Berthé *et al.* 1979a) reveal various shear directions along the faults. It is noteworthy that some of these deformation corridors locally affect the Arsikere Granite.

The L_1 stretching lineation follows a N 180° trend with a slight plunge to the south (Fig. 4b). Lineation trend trajectories systematically converge toward the foliation triple points (Fig. 3c).

In the basement complex, the semi-quantitative analysis of textural fabrics leads to the determination of various types of strain ellipsoid (Flinn 1965). This has helped to establish a cartographic zonation of the three major ellipsoid types (Schwerdtner *et al.* 1976, Schwerdtner & Sutcliffe 1978). The foliation triple points always correspond to zones with L-type shape fabrics, which suggests that the strain ellipsoids are of the constrictional type in the gneissic and migmatitic material. In supracrustal rocks, these constrictional zones are characterized by the presence of superposed structures, i.e. two crosscutting cleavages or folds refolding a cleavage. In the latter case, the axes of these post-schistosity folds are parallel with the finite stretching lineations and the highly variable axial planes. In Fig. 4(b), measurements of the L_1 lineation reflect the mean 3-D geometry of the triple points (lines) (Brun *et al.* 1981, Brun 1983a, Bouhallier *et al.* 1995). LS or S fabrics prevail within the domes.

The Kibbanahalli Arm

The second domain corresponds to the mapped area of the Kibbanahalli Arm and is characterized by much straighter strain trajectories with a constant SE–NW strike, which clearly cut across the previous domain and run parallel with the axial trace of the synform and its cartographic limits (Fig. 3b). The dip of the regional schistosity (denoted here as S_2) is variable (Fig. 4c), but it is shallow at the edge of the Kibbanahalli Arm and becomes steeper towards its core.

Folds found inside the Kibbanahalli Arm are upright with subhorizontal axes. These folds on various scales exhibit the characteristics of post-schistosity folds. Beneath the flanks of the synform, the angular relations observed between stratification and cleavage are incompatible with those expected from a cleavage due to the formation of a synclinal fold. On each side, the cleavage dip is less than that of the stratification (Fig. 5). As one gets nearer the synclinal axis, the upright post-schistosity folds appear and a second cleavage is locally developed. This second cleavage (denoted here as S_3) is of crenulation type and runs parallel to the axial planes of the folds. The late formation of these folds with respect to the shallow-dipping cleavage observed on the outer flanks of belt can be demonstrated by a study of the S_2 cleavage which is refolded in the cores of D_3 folds and by angular relations preserved on their limbs (Fig. 6).

The trajectories of S_2 cleavage in the Kibbanahalli Arm (Fig. 3b & c) cut across the F_1 foliation observed outside the arm of this belt. L_2 lineation follows a constant N15–N25° direction, that is perpendicular to the boundaries of the arm (Fig. 4d). Given the overall strike of the S_2 foliation, the pitch of lineation L_2 is always around 90° (comparison between Fig. 4b & d).

Deformation within the Kibbanahalli Arm is heterogeneous. There are large strain gradients, the most

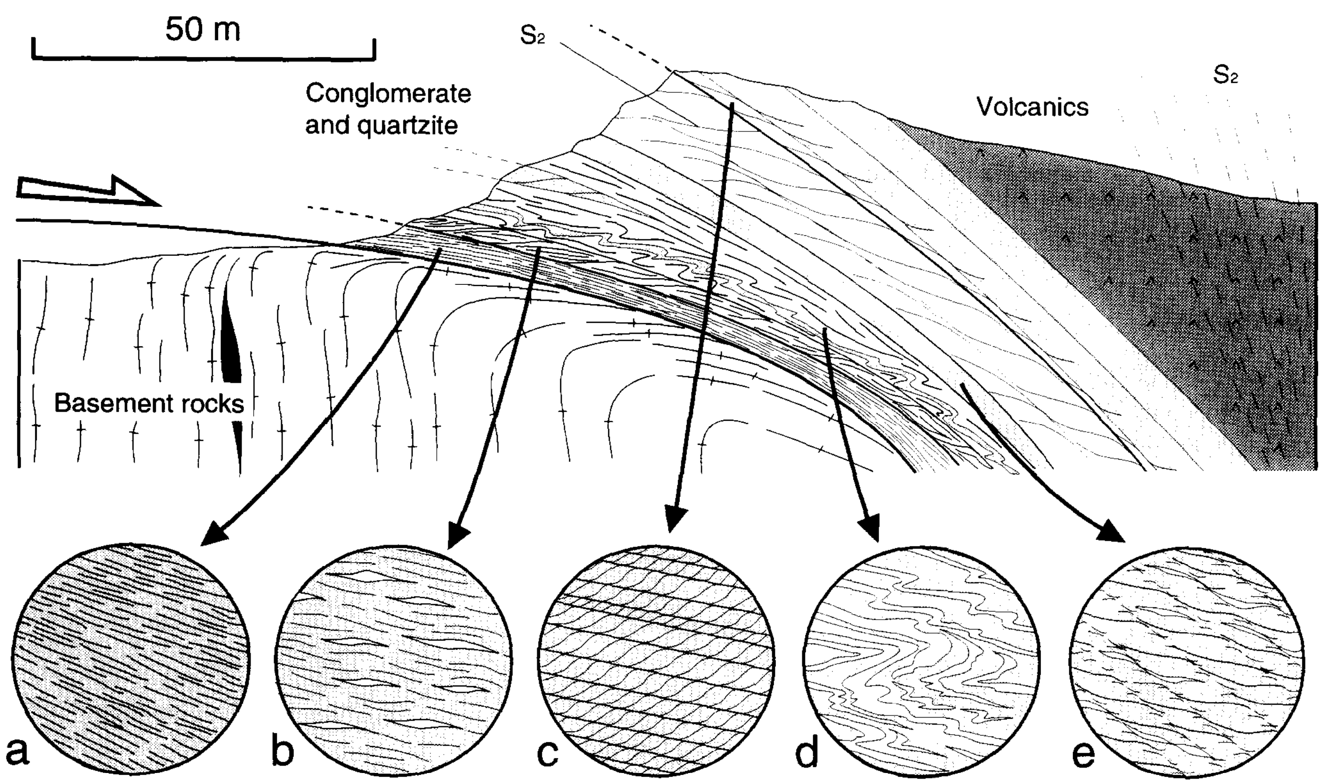

Fig. 5. Synthetic cross-section of the basal contact of the Kibbanahalli Arm showing different types of structure associated with shear deformation. (a) Mylonitic fabrics in quartzites; (b) sigmoidal quartz porphyroclasts in phyllites; (c) discrete shear bands in conglomerates; (d) asymmetric microfolds in banded quartzites; (e) C′ shear bands in phyllites. Diameters of circles are: (a) 2 cm, (b) 6 cm, (c) 20 cm, (d) 40 cm and (e) 30 cm.

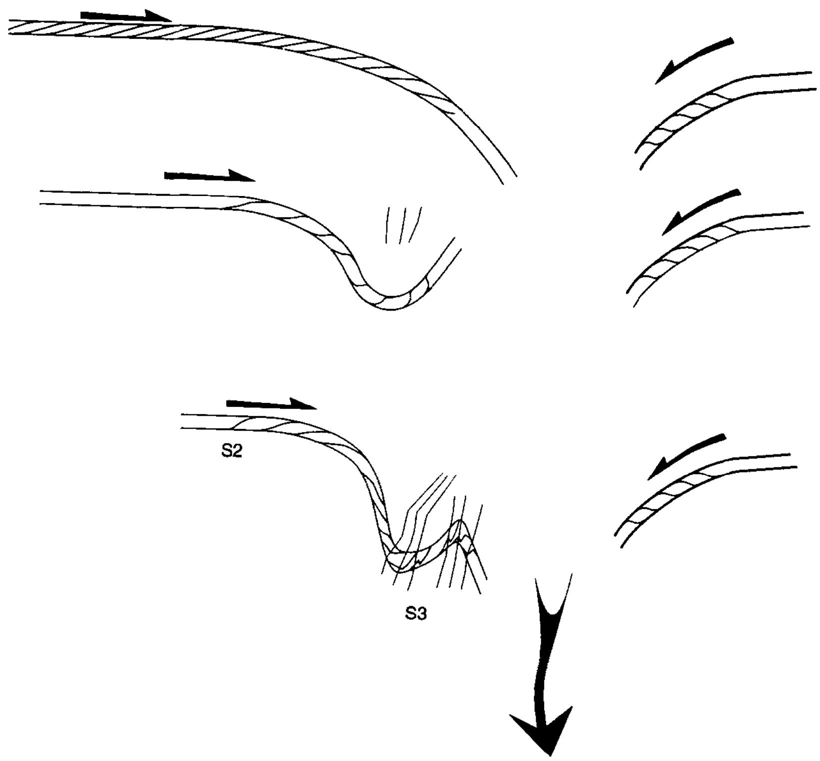

Fig. 6. Sketch cross-section illustrating the progressive folding of the basal shear zone-related cleavage S_2 into F_3 upright folds in the Kibbanahalli greenstone Arm.

important of which are located along the margins of the arm. Indeed, the strain becomes greater near the base of the sequence where major mylonitic zones can be observed. Mylonitization is developed in progressive sequences: the size of quartz pebbles making up the conglomeratic layers diminishes while the matrix becomes increasingly phyllitic and quartzitic beds are transformed into phyllonites. The thickness of these mylonitic zones, which lie parallel to the lithological banding, ranges from one to several metres. S_2 cleavage is clearly linked to the mylonitization sequences, displaying a sigmoidal trace between the mylonitic zones which is characteristic of the sense of shear implied by them (Fig. 5).

In the gneisses and ultrabasic rocks belonging to the substratum of the Kibbanahalli Arm, the F_1 foliation is seen to be affected by D_2, becoming parallel to the contact with the mylonitic zone in the space of a few metres. A transition is observed from a steeply-dipping F_1 foliation strongly oblique to strike in the Kibbanahalli Arm to a foliation—compatible with S_2—that is parallel to the unconformity and concordant with the stratification at the base of the Kibbanahalli Arm. This disruption is also accompanied by a very important strain gradient. The frequency of shear zones decreases noticeably near the top of the basal sedimentary succession, where only a few rare and rather thin examples are to be found, always located on the boundaries of the quartzite beds.

The kinematic analysis of mylonitic zones at the base of the Kibbanahalli Arm is straightforward since indicators of shear direction are numerous and consistent (Fig. 5): apart from the sigmoidal shape of S_2 between shear zones, information is provided by the shape of quartz porphyroclasts (ancient pebbles) within the conglomeratic quartzites. In addition, type-C′ shear bands (Berthé *et al.* 1979b) may be seen, with thickness ranging from a few mms to a few cms, as well as dense networks of parallel bands (Figs. 5b, 7a); these shear bands are slightly oblique to the borders of the mylonitic zones, which are themselves parallel to the lithological layering. They overlap the fabrics defined by the S_2 foliation. Within sequences showing alternations of cm-scale beds of variable competence (such as quartzite/phyllite alternations), asymmetrical synfolial microfolds can be observed at various stages of development (Berthé & Brun 1980) (Figs. 5d, 7b). For all the outcrops studied at the base of the Kibbanahalli Arm, all the shear criteria inferred from the above mentioned structures are compatible. They systematically indicate a movement of the Kibbanahalli series towards the inner part of the belt (Fig. 6). The shear directions are therefore reversed from one side of the arm to the other (Fig. 3c).

INTERPRETATION

Relative timing and structural evolution

Given the discordance between the Kibbanahalli Arm and the F_1 foliation trajectories, the dome-and-basin tectonics of the basement complex must be attributed to an early D_1 episode. The intrusion of the Arsikere pluton into the dome-and-basin structures along with its radiometric age implies that this first deformational episode took place before at least 2.59 ± 0.12 Ga ago. The second event corresponds to the development of Kibbanahalli Arm tectonics during a D_2 deformational episode (associated with S_2 foliation). This deformation took place after the eruption of the Dharwar Supergroup volcanics (which is not yet constrained by zircon ages: 3.0–2.52 Ga), and could be linked to the major phase of cratonization around 2.5 Ga (Drury *et al.* 1984, Chadwick *et al.* 1989, Jayananda & Mahabaleswar 1990, Bouhallier *et al.* 1993). The formation of the transcurrent shear-zones affecting the dome-and-basin structures and the Arsikere Granite could then be coeval with or later than D_2.

The D_1 event

The structure described in the TTG gneisses and J. C. Pura belt taken together bears a close similarity with the one described in a neighbouring region to the south (Bouhallier *et al.* 1993):

(1) Foliation trajectories trace out dome-and-basin structures.

(2) Rock-types systematically occupy similar positions within these structures (the supracrustal rocks being found in the basins and the granite–gneisses in the domes).

(3) Finite strain ellipsoids of constrictional type are found only at the triple points of the foliation.

(4) The main stretching directions (λ_1) converge towards the triple-point junctions. As in the Holenarsipur area (Bouhallier *et al.* 1993), the D_1 strain field can be interpreted as resulting from diapiric gravitational instabilities (Brun *et al.* 1981, Gapais & Brun 1981) between the gneisses and the greenstones (Fig. 8). This strain field may also be interpreted as the result of superimposed large-scale folds (Myers & Watkins 1985).

The D_2 event: décollement and centripetal sinking of the Kibbanahalli trough

This question raises a fundamental point: given the slight deformation and metamorphism associated with the D_2 episode, which is the only one affecting the Kibbanahalli Arm, this deformation can be considered as a characteristic feature of an early stage of the structural development of an intracratonic greenstone belt with respect to its basement.

To summarize the results discussed above:

(1) D_2 took place after the development of dome-and-basin tectonics.

(2) The base of the Kibbanahalli succession is a major décollement surface between the previously deformed crust and the cover rocks (the Dharwar Supergroup).

(3) All the shear direction criteria consistently indicate—on each side of the belt—a displacement of the volcanosedimentary sequence towards the cartographic axis of the belt.

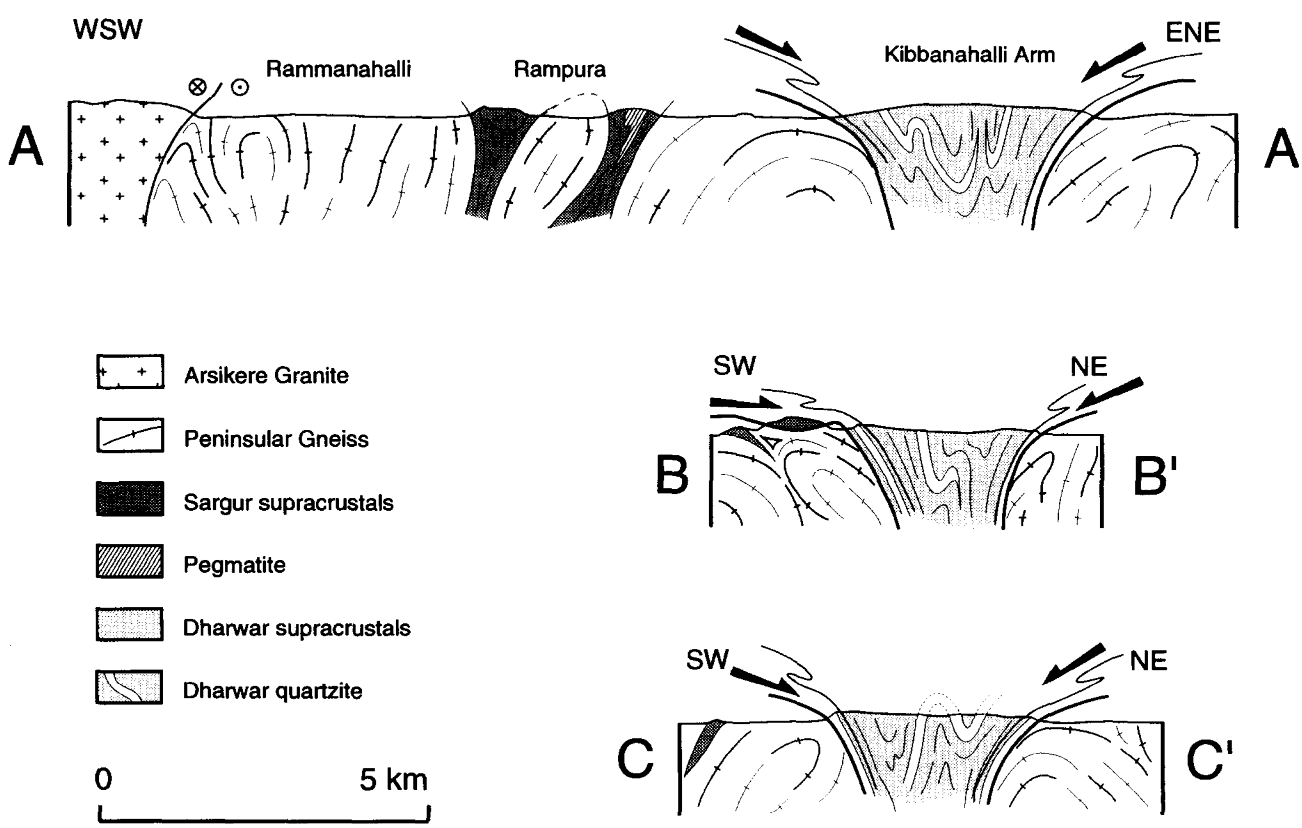

Fig. 8. Interpretative cross-sections through the J. C. Pura area (locations on Fig. 2).

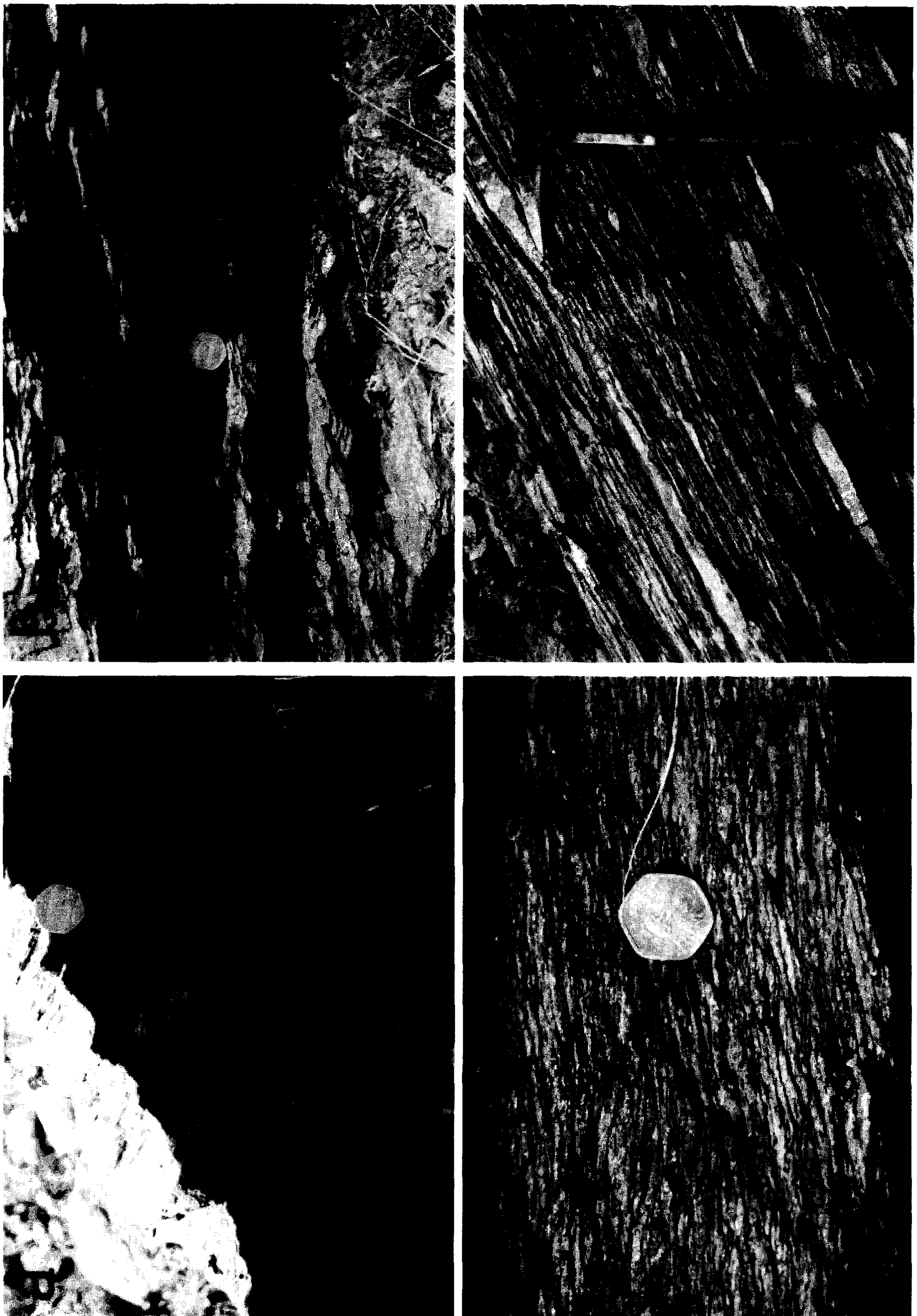

Fig. 7. Macroscopic structures indicating downward sense of shear in the basal shear zone on both limbs of the Kibbanahalli synform. (a) Asymmetrical synfolial folds developed in quartzite–phyllite interbedded sequences; (b) Shear bands in quartzites; (d) C′ shear bands in phyllites (locations on Fig. 3).

(4) Within the Arm, post-schistosity folds (post-S_2) with horizontal axes and vertical axial planes are superposed onto the deformation due to décollement.

It is necessary to consider whether the marked subsidence of the Kibbanahalli Arm sequence due to the effects of centripetal displacements is compatible with the late-stage vertical flattening observed within the belt (Fig. 7). In fact, this type of hypothesis has been experimentally tested. Analogue modelling (Ramberg 1963, pp. 76–84, Dixon & Summers 1983, Talbot *et al.* 1991) has led to an understanding of the spatial and temporal variability and accounts for the coherence between the structures and the internal deformation of subsiding troughs (Fig. 9). Gorman *et al.* (1978) were the first to propose a model of progressive deformation in sinking belts, where vertical flattening occurs necessarily after the initial stages of subsidence in the axial parts of subsident or sagducted sequences. Goodwin & Smith (1980) used the term 'sagduction' to describe this phenomenon. We propose to reconsider this mechanism as being the only one able to explain the progressive deformation observed within the Kibbanahalli Arm.

The new evidence introduced by study of the Kibbanahalli Arm example suggests that sagduction can be facilitated by a décollement at the basement–belt interface. This type of centripetal décollement has been recently described at the base of the Barberton belt (Heubeck & Lowe 1994), where it is considered to have taken place at a very early stage in the evolution of the belt since it is partly contemporaneous with the deposition of the sequence.

STRUCTURAL TESTS AND DISCUSSION

Triple point orientation and dome-and-basin strain pattern

There are noticeable differences between the strain fields described in the basement complex (D_1 domain) and the Holenarsipur domain (Bouhallier *et al.* 1993). These differences concern the geometries of lineation trajectories and triple points. In the Holenarsipur region, the main vertical stretching directions are predominant outside the domains of influence of ductile strike-slip faults—which overlap the dome structures—and the triple points are vertical. In the case studied here, the main directions of λ_1 and the triple points are nearly horizontal (Figs. 3c and 4b). If it is accepted that the D_1 strain field is the result of diapiric geodynamics, as proposed in the case of the Holenarsipur belt, then its features set it apart from the strain fields described in the literature (Brun *et al.* 1981, Bouhallier *et al.* 1993, Jelsma *et al.* 1993). Indeed, contrary to these last mentioned studies, the zones of constrictional strain associated with the triple points of the F_1 foliation generally show gently plunging fabrics. This may be due to three causes:

(1) Some of the triple points described here may be located within the domes; their shallow plunge could be due to an interference between regional horizontal shortening and the diapiric strain field (Bouhallier *et al.* 1995).

(2) For triple points located between the domes, this 'anomaly' may be explained by diapiric movements. The domes are at different stages of development and/or represent various structural levels; the more mature

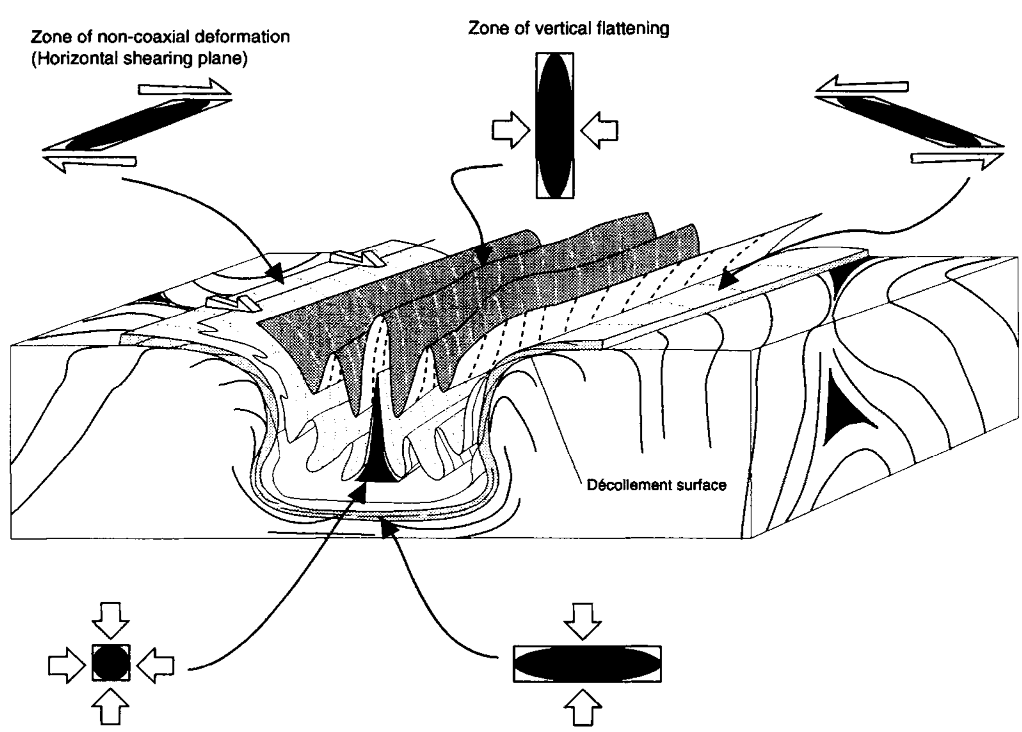

Fig. 9. Schematic block diagram illustrating the spatial variations of structures, strain ellipsoid type and strain regime within a subsiding trough (partly based on Dixon & Summers 1983 and Bouhallier *et al.* 1995). This model is supported by the results of the subsurface structural analysis of the Kibbanahalli Arm presented in this paper.

diapirs may 'cover up' other domes whose ascent has been arrested prematurely (Fig. 3b). The triple point due to constrictional interference between the various diapirs therefore show complex trajectories and may be locally horizontal or shallow-dipping.

(3) The shallow plunge of the triple points may also be the result of the D_2 deformation event. Indeed, the triple points with the shallowest plunge are found near the Kibbanahalli Arm. A post-D_1 horizontal flattening—resulting from D_2 deformation—could explain the reorientation of the main axes of the strain ellipsoid into domes and basins (e.g. less pronounced plunge of the triple points) (Dixon 1975, Schwerdtner *et al.* 1978, Dixon & Summers 1983). Nevertheless, there is no evidence in the basement complex for ductile penetrative deformation postdating the F_1 foliation. The only place where this kind of structure is observed is near the décollement surface, where F_1 is transposed in the mylonite zone, in the space of a few metres.

Tectonics of the Kibbanahalli Arm

At present, because there is no other satisfactory explanation, we consider that the tectonic interpretation of the structural data obtained in the Kibbanahalli Arm can be interpreted in terms of a sagduction model. Indeed, we should point out that:

(1) No folding mechanism can account for the synformal geometry of the Kibbanahalli Arm. The kinematic criteria observed on each side of the arm and the angular stratigraphy/schistosity relations in the basal quartzites are incompatible with a kinematic model for synclinal folds.

(2) The apparent jump in metamorphic grade at the décollement level is a sufficient argument to suggest the applicability of a 'Metamorphic Core Complex' (MCC) model as described in post-orogenic extensional zones (e.g. Brun & Van Den Driessche 1994), and which has already been proposed in the literature on Archaean tectonics (James & Mortensen 1992, Williams & Currie 1993, Kusky 1993). However, such a model cannot be adopted here for the following reasons:

Firstly, the deformation observed cannot be the direct consequence of a thickening process previous to the formation of Archaean crust in this region. Geological observations show the discordance of the formations of the Kibbanahalli belt with respect to a previously eroded basement complex. On the other hand, the deformation observed within the Kibbanahalli Arm clearly results from a early episode which has affected this belt, and no other previous thickening-type deformation can be demonstrated even outside the Kibbanahalli Arm. The metamorphic core complex model cannot account for the post-schistosity folds (post S_2) with vertical axial planes, nor can it explain the S_3 flattening associated with these folds. It should be noted that, in the hanging walls of the detachment faults associated with metamorphic core complexes, the only observed folds have shallow dipping axial planes (Davis 1987). The axes of these folds are parallel to the regional stretching direction associated with the extension (Davis 1975, 1983, Mancktelow & Pavlis 1994), whereas the F_3 folds of the Kibbanahalli Arm show axes that are perpendicular to the stretching direction associated with the décollements.

In an extensional model, the décollement takes place at the dome–cover interface. When the base of the supracrustal sequence is well decoupled, as in this case, the décollement cannot be linked to dome structures located directly in its footwall. In other words, among the many domes found below the Kibbanahalli Arm, none has a geometry that can be directly linked to the geometry of the observed décollement itself. Moreover, the stretching and the shear directions in metamorphic core complexes are relatively constant over large areas (e.g. Gautier & Brun 1994). They also remain compatible and consistent between the footwall and the hanging wall of the detachment fault. This is not the case here, since there is no link between the characteristics of the strain field in the basement and those in the Kibbanahalli Arm; the pattern in the basement is clearly related to an ancient tectonic event.

Finally, in those segments of young mountain chains where post-orogenic extension has been described, the extensional structures are generally asymmetrical and controlled by one or more detachment faults, very few of them being antithetic (e.g. Gautier & Brun 1994). In the KNA, the mylonitic zones observed at the margins are symmetrical and antithetic (with shears converging from one side to the other and a strain intensity at the décollement level similar on both sides), while no detachment fault in the strict sense (Ramsay & Huber 1987, pp. 517–518) has been recognized.

CONCLUSIONS

The present structural study of the Kibbanahalli Arm has made it possible to characterize a late-stage event in the Archaean history of the Dharwar craton. Since this event concerns an upper structural level of the continental protocrust, and because the single-phase tectonic framework studied here is simple, it provides a great deal of information on the relationships between greenstone belts and their underlying granite–gneiss basement. The results of this study can be summarized as follows: the progressive deformation observed within the belt is only compatible with incipient sagduction of greenstones within the underlying basement, which implies it is mainly caused by gravity-dominated processes. Such sagduction is made easier by decoupling at the base of the subsiding material.

From a more regional viewpoint, but still based on structural arguments, this study shows that it is justifiable to subdivide the supracrustal rocks of the Dharwar craton into two diachronous entities. Each greenstone development cycle is associated with distinct deformational episodes. These two episodes are best characterized in the upper structural levels of the Dharwar crust.

In agreement with Bouhallier *et al.* (1993), the present study provides no evidence in favor of tangential tectonics having contributed to crustal thickening in the Dharwar terrain. The simple structure of the Kibbanahalli Arm and the strain field affecting it, as well as the kinematic criteria observed at its borders, are all incompatible with the geometry of a greenstone nappe or thrust slice. In the same craton, diapiric structures have been described at deeper structural levels (Bouhallier *et al.* 1995) which has unequivocally undergone both of the tectonic events clearly identified in this study.

At a time when many authors endeavor to apply uniformitarian models to the evolution of the Archaean terrains, we maintain that vertical movements of the gravitational instability type were predominant in the tectonic development of the Archaean crust of the Dharwar craton (Bouhallier 1995, Choukroune *et al.* 1995). Vertical movements related to body forces have not been described on such a scale in young orogenic belts. Such movements are evidence of the specific mechanical behavior of protocontinents.

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