The Caldas Novas dome, central Brazil: structural evolution and implications for the evolution of the Neoproterozoic Brasilia belt

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Abstract

The Caldas Novas dome (Goiaas state, central Brazil) lies in the southern segment of the Neoproterozoic Brasilia belt (center of the Tocantins Province) between the Goias magmatic arc and the margin of the ancient Sao Francisco plate. The core of the dome comprises rocks of the Meso-Neoproterozoic Paranoa group (passive margin psamitic-pelitic sediments and subgreenschist facies) covered by a nappe of the Neoproterozoic Araxa group (backarc basin pelitic-psamitic sediments and volcanics of greenschist facies, bitotite zone). Hot underground waters that emerge along fractures in the Paranoa quartzite and wells in the Araxa schist have made the Caldas Novas dome an international tourist attraction. A recent detailed structural analysis demonstrates that the dome area was affected by a D_1D_3 Brasiliano cycle progressive deformation in the — 750-600 Ma interval (published U-Pb and Sm-Nd data). During event D_1, a pervasive layer- parallel foliation developed coeval the regional metamorphism. Event D_2 (intense F_2 isoclinal folding) was responsible for the emplacement of the nappe. D_1 and D_2 record a regime of simple shear (top-to-SE relative regional movement) due to a WNW-ESE subhorizontal compression ($\boldsymbol{a}_1$). Event D_3 records a WSW-ENE compression, during which the dome rose as a large-scale F_3 fold, possibly associated with a duplex structure at depth. During the dome's uplift, the layers slid back and down in all directions, giving way to gravity-slide folds and an extensional crenulation cleavage. A set of brittle fractures and quartz veins constitutes the record of a late-stage D_4 event important for understanding the thermal water reservoir.

Keywords: Caldas novas dome; Brasilia belt; Brasiliano cycle; Neoproterozoic; Structural analysis; Thermal water; Tocantins province

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1. Introduction

The Caldas Novas dome (Goiás state, central Brazil) is a structural window in the central part of the Neoproterozoic Tocantins Province, a N-S-trending, double-verging orogen between the Amazon and Sao Francisco cratons, bounded by the Parana and Parnaiba Phanerozoic basins (Fig. 1a). According to a summary in D'el-Rey Silva and Barros Neto (2002), the Araguaia and Brasilia belts reflect the collision of the Sao Francisco and Amazon ancient plates ca. 650-600 Ma ago (Brasiliano cycle), whereas the Paraguay belt records a 550-500 Ma old inversion of a rift opened in the former Amazon plate and filled 600-550 Ma ago. The Brasilia belt, in which there is also evidence of a 800-700 Ma old tectonic event, actually comprises the Goias mag matic arc and internal and external zones (MA, IZ, and EZ; Fig. 1b).

Around the parallel of the federal district (Fig. 1b), the Brasilia belt trends NE in its northern segment and SE in its southern segment. The bend is roughly coincident with the Pirineus shear zone, a nearly E-W- striking zone of high strain that affects rocks of the MA, IZ, and EZ up to the federal district. Farther east, the EZ is in contact with the Meso-Neoproterozoic sedimentary cover of the Sao Francisco craton (Marini et al., 1984; Fuck et al., 1993; 1994). Polyphase Neoproterozoic deformation in the MA, IZ, and EZ is largely characterized by frontal ramps and E-verging thrusts and folds. The easternmost remnants of two major nappes that displaced higher metamorphic grade rocks of the IZ onto the lower metamorphic grade rocks of the EZ are found up to the surroundings of Brasilia (Fig. 1b). The Araxa, Passos, Luminarias, and Socorro-Guaxupe nappes constitute the main tectonic features of the southern end of the belt (Fig. 2).

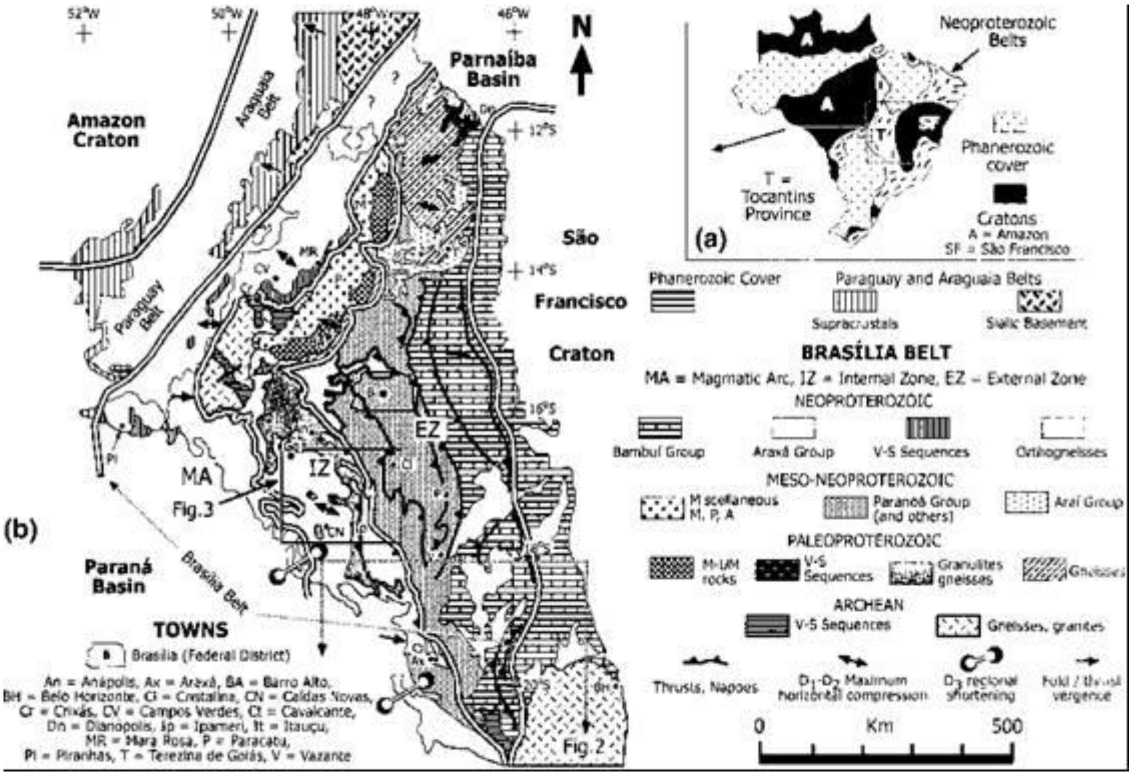

Fig. 1. (a) Simplified map displaying the major lithotectonic units of Brazil and highlighting the Tocantins Province between the Sao Francisco and Amazon cratons. (b) Simplified geological map of the central part of the Province, based on Marini et al. Marini et al. (1984), Fuck et al. (1993, 1994), and Pimentel et al. (1999, 2000). Vectors for maximum horizontal compression (σ_1) are according to a vast inventory of lithostructural data (D'el-Rey Silva and Barros Neto, 2002). Direction for D_3 shortening is based on this paper and Seer and Dardenne (2000).

Although the Caldas Novas dome is well known for its thermal waters, few international publications (e.g. Drake, 1980; Troger et al., 1999) discuss its geology, and the most comprehensive general and recent study of the dome, which remains unpublished, does not address a structural analysis (Campos et al., 2000). Since 1999, a Brasilia University-Technical University of Berlin research program has provided additional information on the regional hydrology. Further data on the geology and water chemistry appear in technical reports sponsored by DNPM (a federal office) and local companies (e.g. UHE-Corumba, Geocaldas, Geocenter).

In this article, we summarize the results of a structural analysis carried out in the Caldas Novas dome area in October-December 2000. We provide a structural database relevant for understanding the regional geology and underground reservoirs of hot water. All data are integrated into a model that accounts for the tectonic evolution of the dome and the physical-chemical characteristics of the hot water reservoirs. Our results, combined with the results of previous and independent structural analyses, geological mapping, and age determination shed light on the critical difference between the two segments of the Brasilia belt, a challenge for any interpretation of the tectonic evolution of Tocantins Province.

2. Tectonic setting

2.1 The magmatic arc (MA), internal zone (IZ,) and external zone (EZ)

The MA, originally characterized by Pimentel and Fuck (1992), consists of tonalitic orthogneiss, granites, and volcanic-sedimentary sequences (Lacerda Filho et al., 1999), but it remains poorly known in the north (Pimentel et al., 2000). The IZ includes, in its northern segment, an Archean-Paleoproterozoic continental block of gneiss, tonalite, and greenstone-type volcanic-sedimentary sequences affected by polycyclic tectonism (Queiroz, 2000); three mafic-ultramafic layered intrusions metamorphosed to the granulite facies (plus metavolcano- sedimentary sequences to the west, not distinguished in Fig. 1b) and narrow tectonic slices of Paleoproterozoic basement rocks to the east; and a strip of metasediments that comprises, from S to N, rocks of the Araxa, Paranoa, and Serra da Mesa groups (Fig. 1b; D'el-Rey Silva and Barros Neto, 2002). In the south, the IZ includes the backarc basin-like Araxa group, which mainly comprises amphibolite-greenschist facies metasediments and basic metavolcanics (e.g. Brod et al., 1991) and a longitudinal belt of Neoproterozoic granulites. The eastern boundary of the IZ is a contractional frontal ramp that dips less than 30° to the west and places high-P rocks in sharp contact with the Araxa nappe, as well as with low greenschist facies and strongly shortened passive margin-like metasediments of the EZ (Fig. 1b).

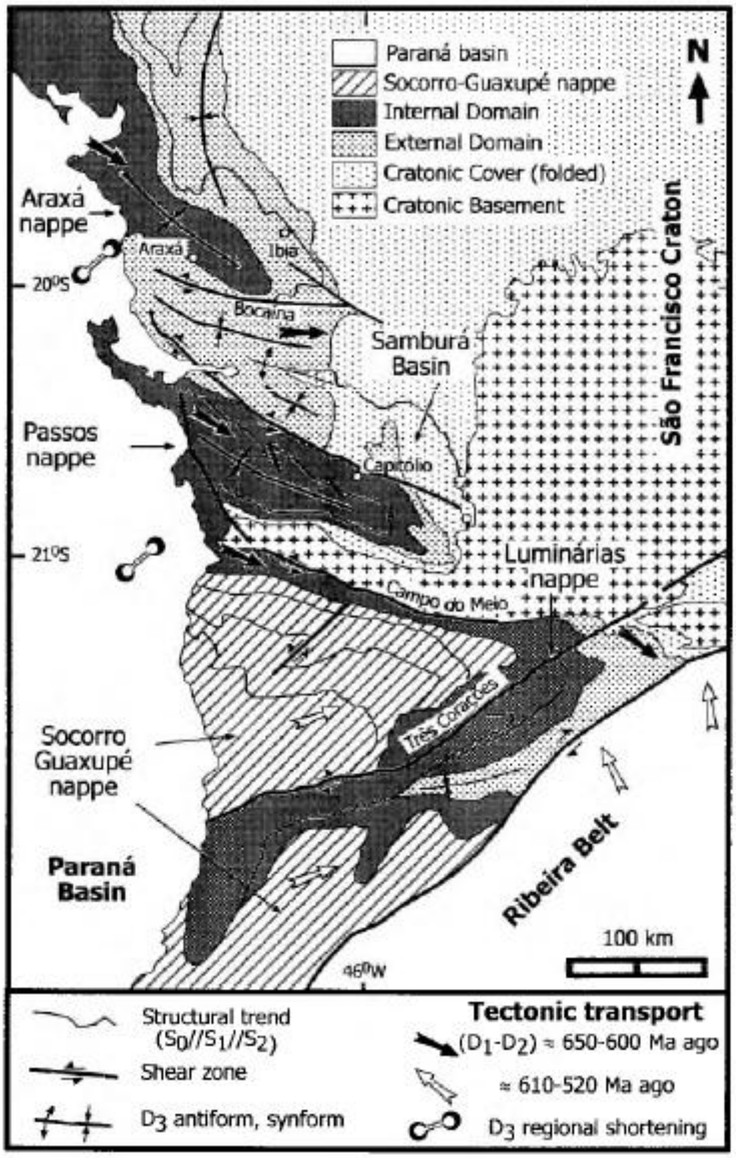

Fig. 2. Simplified geological map of the southern segment of the Brasilia belt (based on Valeriano et al., 2000, with addition of D_3 shortening direction). The Araxa, Passos, and Luminarias nappes displaced rocks of the IZ to the ESE onto the EZ during D_1-D_2 and were lately shortened in the ENE direction (the D_3-related Araxa and Passos synforms and intervening antiforms). Emplacement of the Socorro-Guaxupe nappe system onto the Passos and Luminarias nappes followed in time and occurred in the same overall ENE direction.

The EZ comprises, from N to S, Archean/Paleoproter- ozoic basement rocks and Meso-Neoproterozoic, sub-greenschist to amphibolite facies metasediments and metavolcanics. The younger rocks include the Mesoproterozoic rift basin-like Aral group (metasiliciclastics and metavolcanics), which correlates with the Serra da Mesa group; the passive margin-like Meso-Neoproterozoic

Paranoa and Canastra groups (metasiliciclastic and metacarbonate sediments; calc-schist, metapelite, and quartzite) and the Neoproterozoic Ibia and Vazante groups (respectively, siliciclastics and fine-grained silici- clastics and metacarbonates); and the Bambui group (siliciclastic and carbonate metasediments), which rests on the Paranoa group along the eastern margin of the EZ and stretches over the Sao Francisco craton.

2.2 Surrounding the Caldas Novas dome

The area surrounding the Caldas Novas dome displays north-trending tectonic slices of a frame of Brasiliano cycle frontal ramps and intervening lateral ramps (Fig. 3). Data in Lacerda Filho et al. (1999), Fischel et al. (1998, 1999a-c) and Pimentel et al. (1999, 2000) indicate a general lithostratigraphy consisting of a Paleoprotreozoic crystalline basement (gneiss, orthogneiss, and the Silvania metavolcano-sedimentary sequence); a Meso-Neoproterozoic cover that includes the Paranoa, Canastra, and Ibia groups; and a large set of Neoproterozoic rocks that includes metasediments of the Araxa and Ibia groups, metavolcano-sedimentary sequences termed Rio Veríssimo and Marata, orthogneiss of the Goias MA, granulitic rocks akin to the Anapolis-Itaucu complex that occurs outside the area to the north of Leopoldo de Bulhoes, and the Ipameri, Aragoiania, and Rio Piracanjuba granite suites.

The area also includes elongated bodies of (mylonitic) granitic intrusions syntectonically emplaced in the Araxa supracrustals (Pimentel et al., 2000; not individualized in Fig. 3). They comprise the subvolcanic Marata granites (, 794 Ma old, U-Pb data) derived from the Paleopro- terozoic continental crust, as well as 700-780 Ma old (Rb-Sr data) granites, which display strongly and slightly peraluminous signatures, respectively. The former also occurs in the Marata area, east of Pires do Rio (Fig. 9 in Pimentel et al., 1999), and has an Sm-Nd composition with ^{t}dm model ages in the 1.73-2.5 Ga interval, similar to that of the Araxa metasediments. The latter are Sesmaria- and Tambu-type granites, with main outcrop areas east of Ipameri along the Araxa-Ibia contact (Fig. 10 in Pimentel et al., 1999). These granites derive from Neoproterozoic juvenile material with Sm- Nd model ages of 1.1-1.0 Ga, equivalent to that found in the adjacent Goias MA (Pimentel et al., 2000).

The Caldas Novas dome area is particularly important because it lies in the southern part of a granulite belt that includes (Fig. 1b), three M-Um intrusions (granulitized ca. 780 Ma ago) and the Anapolis-Itaucu complex extending south of Ipameri (east of Caldas Novas). Granulites of this complex derive from supracrustals and granites, are aged ca. 603 ± 31 and 633 ± 28 Ma, according to U-Pb data in garnets and zircons (Fig. 1b; Fischel et al., 1999a,b; Fischel, 2002) and occur tectonically intercalated within lower metamorphic grade rocks of the ca. 750 Ma old Araxa nappe. According to these authors, both high- and low-grade rocks display similar Sm-Nd isotopic signatures. To provide a mechanism to explain the uplift of the granulites of the IZ and their field relationship with the lower-grade metasediments of the nappe and the EZ, D'el-Rey Silva et al. (1996, 1997) and D'el-Rey Silva (2002a) have shown that the contraction of the EZ post-nappe emplacement must have occurred through the mechanism of underthrusting, with the upper crust accompanying the wetward ductile flow of the ancient Sao Francisco plate.

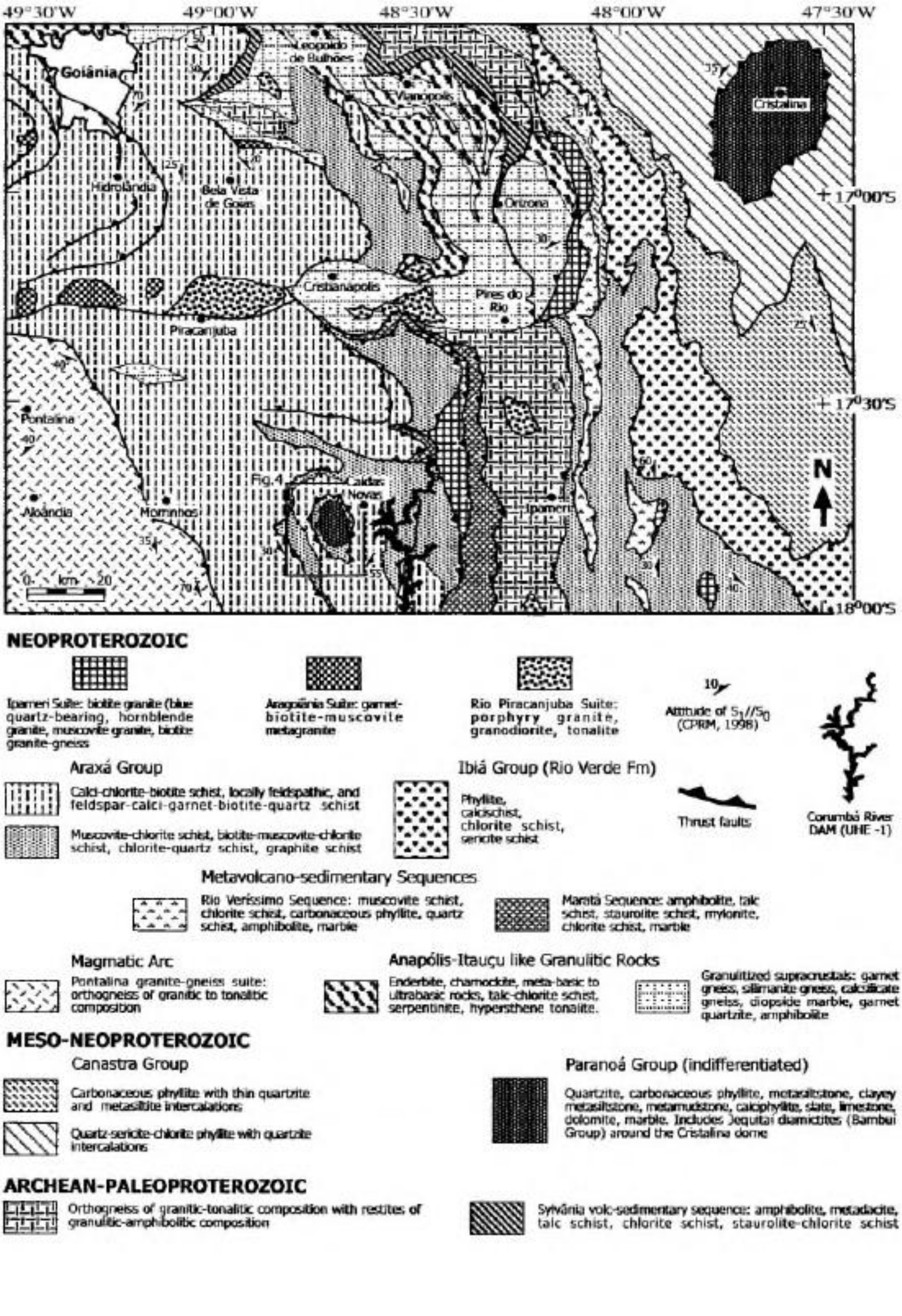

Fig. 3. Simplified geological map of the southern segment of the Brasilia belt in the surroundings of the Caldas Novas dome (inside the study area, Fig. 4) based on Lacerda Filho et al. (1999), Fischel et al. (1998, 1999a-c) and Pimentel et al. (2000). The subdivision and the vertical succession of the Neoproterozoic rocks are informal and do not imply any relative age relationship.

3. Summary geology of the Caldas Novas dome area

3.1 Summary lithostratigraphy

The Caldas Novas dome is a — 20 X 12 km elliptic structure, with its longer axis trending NNW (Fig. 4). The study area includes Proterozoic tectonites of the Paranoa and Araxa groups. The former occupies the core of the dome, whereas the latter surrounds it and spreads in all directions. Undeformed conglomerates of a local body (Fig. 4) have been assigned to the Cretaceous Areado group (Campos et al., 2000).

In the study area, the Paranoa group is a psamo-pelitic sequence comprising four lithostratigraphic units found in normal contact (Campos et al., 2000). The basal unit consists of commonly white, wavy rippled, cross-stratified orthoquartzite that is silicified and crops out along the internal border of the dome but is partially covered by younger laterite (omitted in Fig. 4) in the central part.

Five drill holes along an E-W cross-section prove the continuity of the quartzite, at least at depths of up to 100 m. Moreover, gravity data indicate a negative anomaly (Haraly, 1978, 1980) that indicates that the Paranoa group (lighter material) extends to 1000 m deep underneath the dome. Above, the basal orthoquartzite passes to a unit of brown to red, argillaceous, immature, and commonly laminated quartzite; then to intercalated layers of quartzite and metasiltstone (a metarhythmite unit); and finally to an upper unit of metasiltstone, metamudstone, and argillous metasiltstone.

The Araxa group consists mainly of schists or layered sequences of mica schist, quartzite, siltic schist, metapelite, pelitic metasiltstone, or metasiltstone. The thicknesses of these layers may vary within dm- to m- or even dam-scales. Slightly or unweathered schist commonly exhibits a greyish green color, and the schist layers display a basic mineralogy of chlorite, biotite, muscovite, and quartz combined in variable percentages across the area. Feldspars may occur in the matrix, but no feldspar grain was found in the field, even with the aid of a hand lens. Nevertheless, cm-size flakes of white mica are relatively common (e.g. outcrops 2, 68, 72). In several outcrops closer to the contact with the Paranoa rocks around the dome (e.g. 33, 37, 38, 39, 50, 57, 66, 67, 70), the bottom of the group consists of biotite-quartz schist, biotite-quartzite, and chlorite-biotite-schist that forms a continuous layer resistant to erosion. Small bodies of ultramafic rocks also have been identified regionally within the Araxa group (Drake, 1980), as well as in the SW corner of the study area (omitted in Fig. 4 Campos et al., 2000).

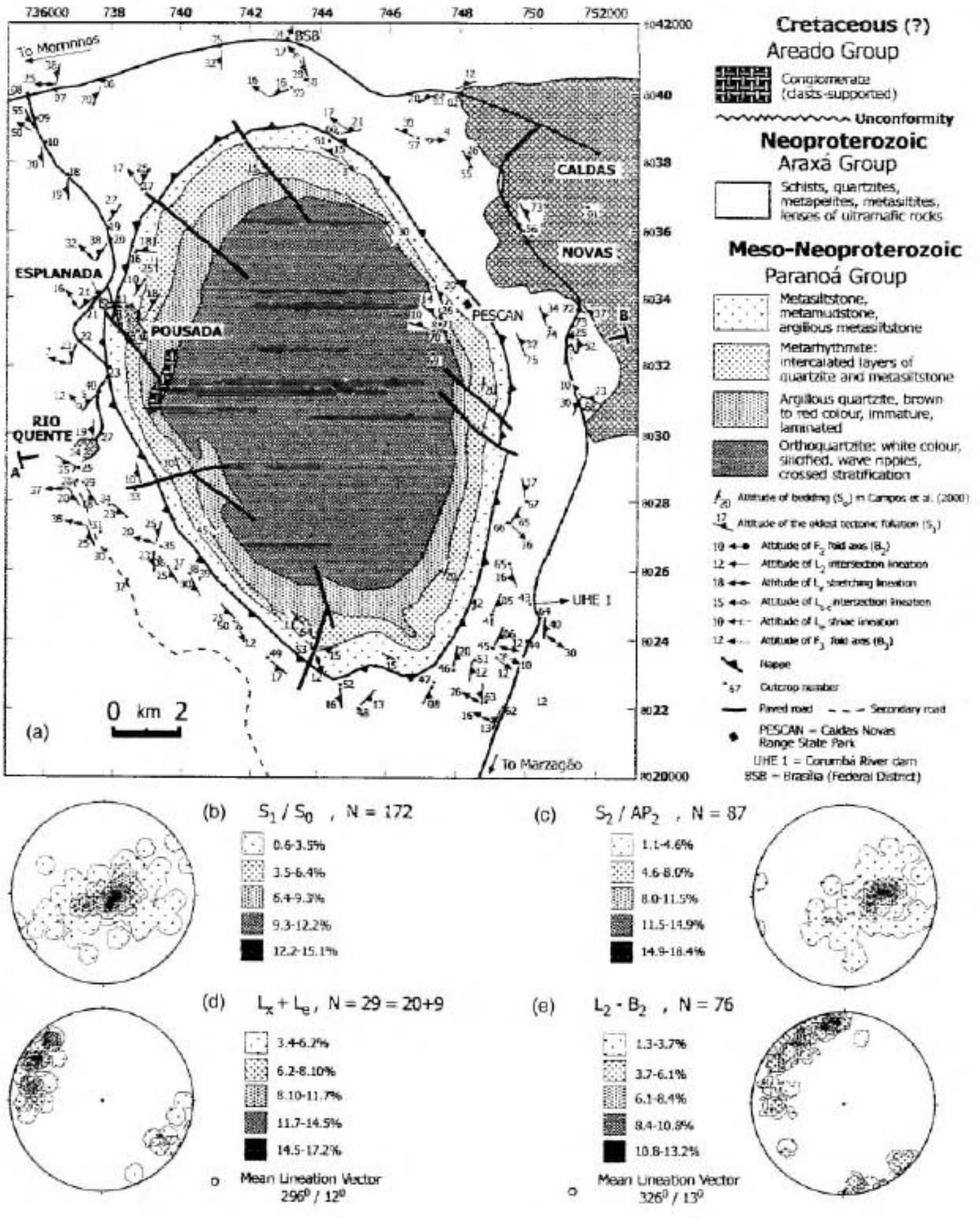

Fig. 4. (a) Simplified geological map of the studied area displaying some of the main D_1-D_2 foliation and lineation data. All overburden has been omitted. Main towns are Caldas Novas, Esplanada, and Rio Quente. The approximate limits of a major tourist park (Pousada) are also shown. Location of the roads is approximate between two successive outcrops,

particularly those far from one another. Line AB stands for the cross section in Fig. 12. Lower hemisphere stereograms (Schmidt-Lambert net) of structural data are in b-e.

3.2 Underground Waters

Thermal waters have been noted in the region of Caldas Novas since the beginning of the eighteenth century (Oriente, 1982). The first site was found along a stream that started in a fracture zone and cross-cut quartzites of the western flank of the dome and passed into Pousada park. From here, the stream turns into the Rio Quente, which flows through Esplanada (Fig. 4) with a rate of 1.4 m^3/s of water as hot as 38 °C but becomes colder to the west. According to Zschocke (2000), there are three kinds of underground water reservoirs surrounding Caldas Novas: (1) the Paranoa aquifer whose thermal waters (> 51 °C) flow from open fractures in the low greenschist metamorphic facies quartzites, (2) the Araxa aquifer whose thermal waters (> 41 °C) have been explored from the Araxa schists; and (3) a porous aquifer consisting of cold waters (24 °C) in the lateritic cover.

Since the early 1970s, Caldas Novas has been a tourist site, benefiting from the activities of the numerous hotels and private clubs established in the city with pools of thermal waters from more than 400 boreholes made in the Araxa schist. Tourism currently also spreads into localities to the west of the dome and has caused a new wave of civil construction and tourism jobs. The intensive exploration of the thermal waters has driven DNPM toward more restrictive control in an attempt to avoid misuse of the underground resources.

3.3 Summary of deformation events and associated structures

The area displays clear evidence of a progressive (Brasiliano cycle) evolution during three ductile to ductile-brittle deformation events (D_1-D_3) that affect the primary layering (S_0), defined by the intercalation of beds of different composition and easily recognizable at different scales in Paranoa and Araxa metasediments. The inventory of structures includes folds, such as F_2 and F_3, three axial-plane foliations (S_1-S_3), fold axes (B_2 and B_3), and lineations such as the intersections of S_2 with S_0 and S_1 and stretching and striae lineation (L_2, L_x, and L_e). The Araxa schists also display two extensional crenulation cleavages (c^0 and d^0). Approximately five sets of brittle fractures (S_4) have been recognized, some of which may host cm- to dm-thick quartz veins.

Events D_1 and D_2 record a progressive simple shear along frontal ramps and imply a relative top-to-ESE displacement along surfaces subparallel to S_0 that gently dip to the NW. The D_1 event developed the pervasive layering-parallel S_1 foliation, but metamorphism is more intense in the Araxa rocks (biotite zone, greenschist facies) than in the Paranoa rocks (subgreenschist facies). The D_2 event is responsible for the placement of higher metamorphic grade Araxa rocks as a nappe on the lower-grade Paranoa group. The Araxa group displays many isoclinal to tight F_2 folds of cm- to dm-scale, whereas Paranoa rocks exhibit F_2 folds that are dm- to m-scale and commonly isoclinal to tight. Further slip resulted in the development of the c^0 extensional crenulation cleavage, which records progressive displacement within Araxa schists with systematic top-down to various directions between NE and SE. The D_3 event developed a major, NNW-SSE-trending, double-plunging F_3 antiform (i.e. the dome itself) that also affected the nappe.

4. Main tectonic structures

4.1 D_1 event of deformation

Event D_1 mostly developed the S_1 foliation and probably the F_1 folds as well. Drake (1980) assumes that such folds occur, though they are hard to find and impossible to measure because they were so severely flattened that no hinge can be recognized. In areas where deformation records a progressive inter- and intralayer flow, as in Caldas Novas, F_1 folds may not develop; therefore, F_1 should be recognized as just those folds that are actually refolded by F_2 folds or cross cut by S_2 foliation.

Foliation S_1 is penetrative and subparallel to the primary bedding. In the Araxa rocks, S_1 is commonly a mylonitic structure that consists of a cm-scale S -C pair in the schists or a mineral foliation in relatively coarser-grained siliciclastic layers. The S-C mylonitic structure also may be identified in some intercalated quartzite layers. Almost everywhere, the S-C asymmetry indicates a top-to-ESE relative movement. The intersection lineation may be safely measured only locally, due to the S and C planes in S_1 ($Li_{S\text{-}C}$); therefore, it has not been treated statistically. The Araxa schist commonly displays cm-thick and dm-long plate-like quartz veins subparallel to the S or C foliations of the pair S-C = S_1. Such plates are useful for identification of F_2 folds, foliations *c!* and d^0, and L_e striae lineation. In the Paranoa rocks, S_1 is more commonly a mineral foliation defined by fine-grained white mica and flattened quartz in quartz-rich layers or very fine-grained mica (sericite) in phyllites or layers with greater clayey components. The stereogram for S_1/S_0 (Fig. 4b) shows not only a clear dispersion along the center of the diagram, but also a girdle that reflects the influence of the F_2 and F_3 folds. Despite all this deformation, the maximum concentration of poles still records gentle dips to the W or WNW.

4.2 D_2 event of deformation

Event D_2 developed the F_2 folds, the axial plane S_2 foliation, the penetrative L_2 lineation, and the B_2 fold axis. The stretching lineation (L_x) and extensional crenulation cleavage (c^0) relate to both D_1 and D_2 events. Foliation S_2 (commonly a mineral foliation or, more locally, a slaty cleavage) is less prominent than S_1 but is also penetrative. Fine-grained flakes of sericite and chlorite (less muscovite) and flattened crystals of quartz define S_2 in Araxa rocks, whereas sericite and flattened quartz do so in Paranoa rocks. The slaty cleavage consists of closely spaced planes of partition and is more common in fine-grained Paranoa rocks, even quartzites. The mineralogy and small grain size of micas along S_2 indicate that D_2 is associated with a low greenschist facies metamorphism.

The F_2 folds are cm- to dm-scale in outcrops of Araxa schists and up to 10 m scale in the Paranoa rocks. They are commonly tight to isoclinal, gently to moderately inclined, or even recumbent, and often evolve into sheath folds (Fig. 5a and b). The latter are key features to decipher the tectonics in the area, in that they keep a constant orientation and clearly define a generally down-dip stretching direction L_x with a strong striae lineation (L_e). Although the F_2 sheaths are observed directly in the mica schists (e.g. Fig. 5b, outcrops 25, 45), they are more commonly seen in Araxa schists that contain large amounts of S_1-parallel plate-like quartz veins. In some localities, the quartz plates are remnants of thin quartzite layers commonly encountered in the Araxa metasediments (e.g. Fig. 5). The plates of quartz also may display a strong striae lineation parallel to the axis of the F_2 sheath, a few centimeters apart (e.g. outcrops 25, 35, 45). This evidence justifies the interpretation that the striae indicate slip along the stretching direction; therefore, L_x and L_e data could be included in a single stereogram (Fig. 4d).

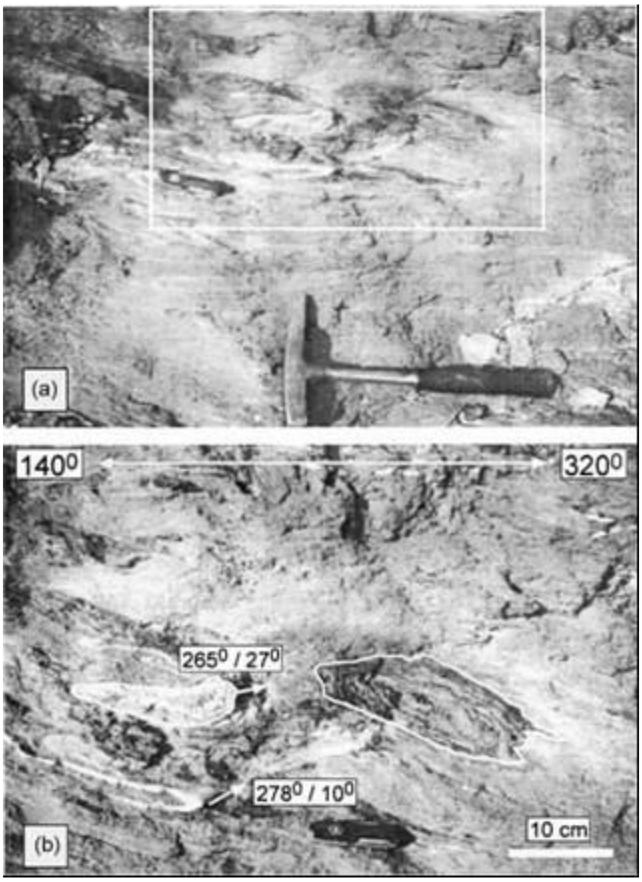

Fig. 5. (a) Isoclinal F_2 folds affecting Araxa siltic schist (outcrop 22, W of the dome, Fig. 4) in which primary layering (S_0) is defined by intercalation of cm-thick layers and is parallel to a penetrative schistosity (S_1). The rectangle area is in the detail (b) and displays two dm-size elliptic structures defining the tubes of F_2 sheaths, the one to the right developed in the mica schist. Axes of F_2 folds (arrows) are parallel to the tube axis (attitudes are shown) and down-dip on the S_1 planes.

The F_2 sheaths are also found in Paranoa rocks at a dm- to m-scale (e.g. outcrops 26, 53,54). The stereogram for S_2 and AP_2 (Fig. 4c) is similar to that for $S_1//S_0$, which indicates that such planes are nearly parallel in the field because the F_2 folds are frequently isoclinal. In areas where this situation is common, the term L_x applies because it is impossible to distinguish whether tectonic slip took place during D_1, D_2, or both events. The distinction is clear where actual F_2 sheath folds are observed, but the striae observed on a plate of quartz likely developed progressively and probably record the net slip from both events.

Both B_2 and L_2 both plunge to a fan of directions that includes three main populations (Fig. 4e). The first is the same for L_x (Fig. 4d) because of the F_2 sheathing process. The second (and smallest) population indicates axes that plunge shallowly to the SW, according to the few F_2 folds preserved in their original trend. The third and largest population indicates plunges of B_2 and L_2 to the N-NNW or S-SSE, which result from the incomplete rotation of B_2 from the SW (original trend) to the direction of maximum stretching (WNW). Thus, the mean lineation vector (326°/ 13°; Fig. 4e) differs slightly from the average L_x (296°/12°; Fig. 4d).

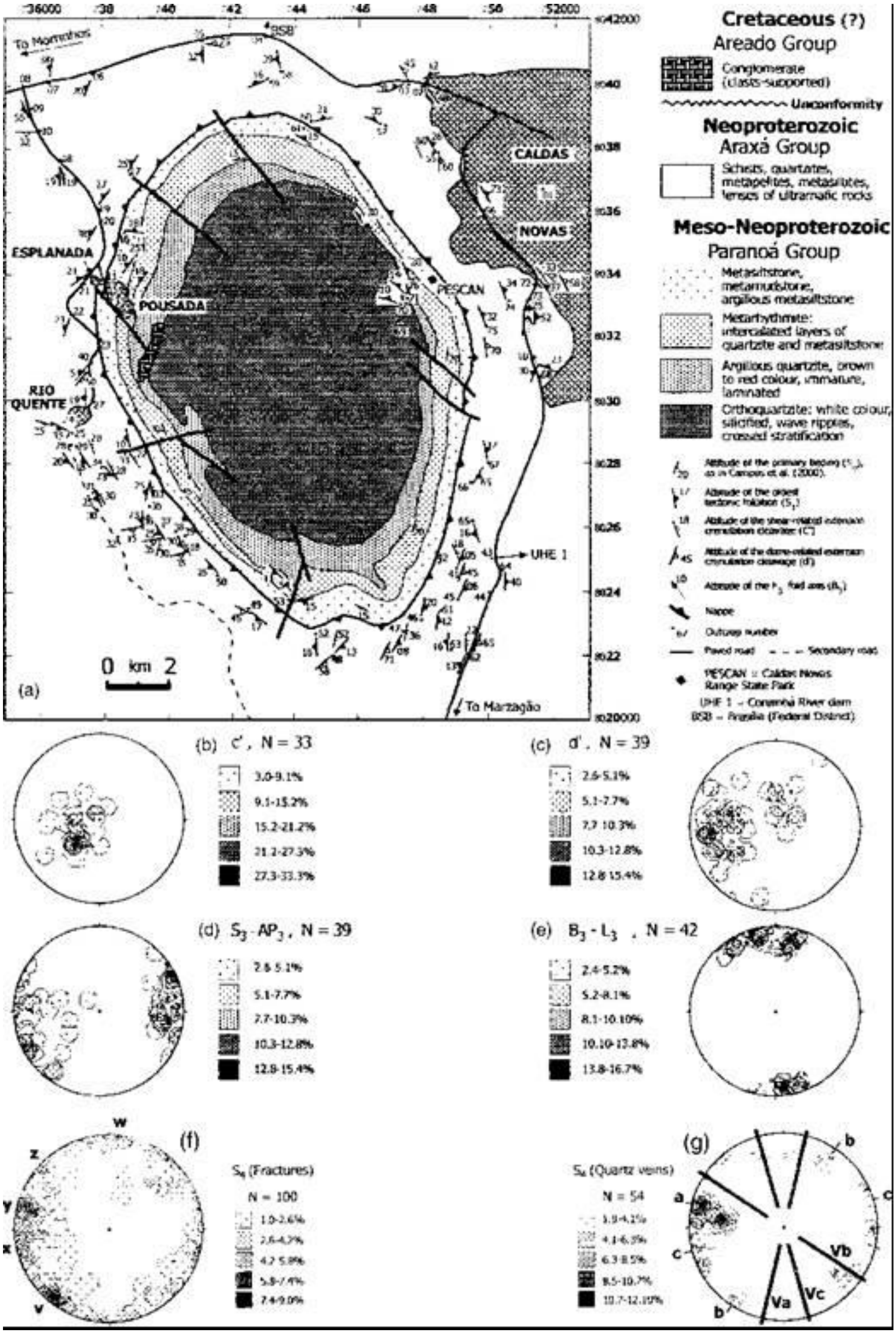

Fig. 6. (a) Same map as in of Fig. 4 displaying the main foliation (S_1/S_0) and the two extensional crenulation cleavages (c0, d0) that affect $S_0//S_1$ in the study area. Lower hemisphere stereograms (Schmidt-Lambert net) of structural data are in b-g.

The S_1 data from Fig. 4a are plotted on a map (Fig. 6a) with the other main foliation in the area (c^0, d^0), whereas other data for brittle fracturing and quartz veining are plotted in Fig. 6b-g. The extensional crenulation cleavage c^0 is common in the Araxa mylonitic schists, both inside and outside the study area, such as in outcrops along the road to Ipameri and to the east of the Corumba River (Fig. 3). Foliation c^0 consists of subparallel, discontinuous, slightly curved surfaces (not rarely anastomosed) that strike at a high angle with the direction of L_x and dip gently to moderately to the east (Fig. 6b). The c^0 planes affect the S-C foliation and make it a sigmoidal-shaped feature, which suggests top- down to easterly directions (Fig. 7). The discontinuous nature of c^0 is a key feature that distinguishes it from d^0 planes. Discontinuity occurs because c^0 surfaces start and die along planes (foliation c) close to one another, which resembles the c^0 structure described by Lister and Snoke (1984) for progressive extension during the evolution of SC mylonites in general. Planes of c^0 formed during D_1 and D_2; they are observed displacing F_2 folds in the Araxa schists (Fig. 7).

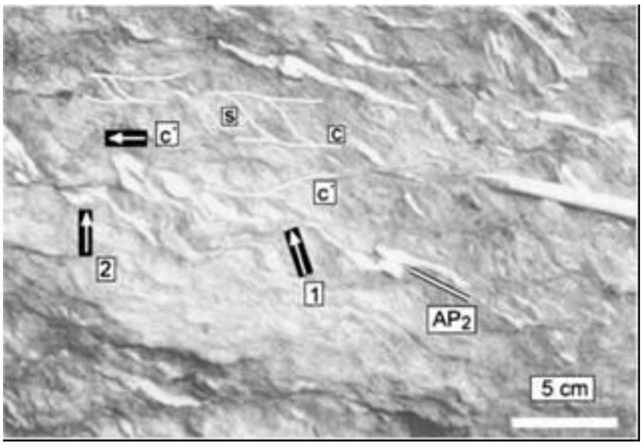

Fig. 7. The syn-D_x/D_2 extensional crenulation cleavage (c^0) affects the Araxa mylonitic schist and is observed everywhere in the area. It affects a typical S-C pair (= S_1 foliation) or even F_2 folds in several dam- to hm- scale outcrops W and E of the dome. Photograph taken in outcrop 35. Arrows 1 and 2 indicate foliation c^0 displacing quartz veins that had been affected by F_2 folds.

4.3 D_3 event of deformation

The structures described in this section (S_3, F_3, and d^0) relate to the uplift of the Caldas Novas dome. This event not only modified the dip of D_1-D_2 structures around the NNW-SSE-trending axis of the dome, but also moved the layers down, mainly to the E and W directions, but also to the NE, NW, SW, and SE.

Foliation S_3 is a spaced or crenulation cleavage, generally marked by very fine-grained sericite, that commonly occurs as the axial planar cleavage of F_3 folds. These folds are open to tight, steeply inclined to upright, and subhorizontal to gently plunging and indicate a WSW-ENE shortening during D_3. This interpretation, justified by the constant and consistent orientation of both S_3//AP_3 and the fold axis B3 (Fig. 6d and e) is positively tested against other data from the regional geology. The F_3 folds display varied style, scale, and vergence due to two basic processes that may be attributed to their formation: a late-D_2 progressive slip along S_0 -S_1 and S_2 planes (continuation of the D_1 -D_2 inter- and intralayer shearing) and gravity sliding.

The F_3 folds formed by inter- and intralayer shearing include (1) open to tight folds of 10 cm scale, upright or steeply inclined, east-verging, with crenulation-style that refold F_2 isoclinal folds (e.g. outcrop 02 for Araxa, 54 for Paranoa rocks); (2) gentle folds of 10 m- or greater scales found far from the dome, such as east of Caldas Novas, along the road leading to Ipameri; and (3) tight, 10 m scale folds that occur in outcrops 68 and 73 (Figs. 4a and 6a) and indicate a km-scale F_3 fold affecting the Araxa schist to the east of the dome, with a fold axis nearly N-S (to the east of PESCAN, Fig. 6). The western limb of the major F_3 fold east of PESCAN is affected by upright, 10 cm scale F_3 crenulations oriented similar to the larger fold, as seen in outcrops 68, 73, and 74. Farther south, between the road to Marzagao and the dome (Fig. 6a), the Araxa layers that dip toward one another (see outcrops 62, 63, and 65 and their relationship to surrounding outcrops) suggest the existence of other km-scale F_3 folds. The same finding is valid south of the dome (outcrops 48 and 52). For all these folds (measured or deduced), the fold axis B_3 plunges to the NNW or SSE (Fig. 6e). The F_3 folds, due to gravity sliding, appear on the western and eastern sides of the dome. Those on the western side include 10-100 m scale, WNW-verging folds (Figs. 8 and 9) that affect D_1 and D_2 structures and display vergence compatible with a top-to-the-WNW shear, in contrast to the sense implied by D_1 -D_3 structures in general. The fold shown in Fig. 8 lies closer to the dome and displays layers that underwent intense brecciation during folding. Those on the eastern side include, among others, m-scale folds found in the Paredao outcrop.

Foliation d^0 is an extensional crenulation cleavage that extended D_1 -D_2 structures, according to mm- to cm-scale displacements, top-down to the south, east, or north (Fig. 6a). Foliation d^0 does not occur to the west or northwest of the Caldas Novas dome (Fig. 6c), other than the specimen that dips to the S in outcrop 10 (Fig. 6a). As will be discussed subsequently, this is a key feature for understanding the origin of d^0 and the dome. Foliations d^0 and c^0 look very similar, which suggests compatibility in the conditions of ductility and metamorphism that prevailed during their evolution. Nevertheless, there are key and subtle differences that distinguish extensional crenulation cleavages as d^0 or c^0 in outcrops. Whereas c^0 surfaces are typically sigmoidal shaped and their trace is only visible for few centimeters, foliation d^0 is defined by more discrete and planar surfaces that are continuous for nearly 1 m or more (Fig. 10). In flat-lying or 3D outcrops where both foliations occur (e.g. 36, 37, 38, 41, 47, 62, 75), their distinctive characteristics are evident, and d^0 is always steeper than c^0.

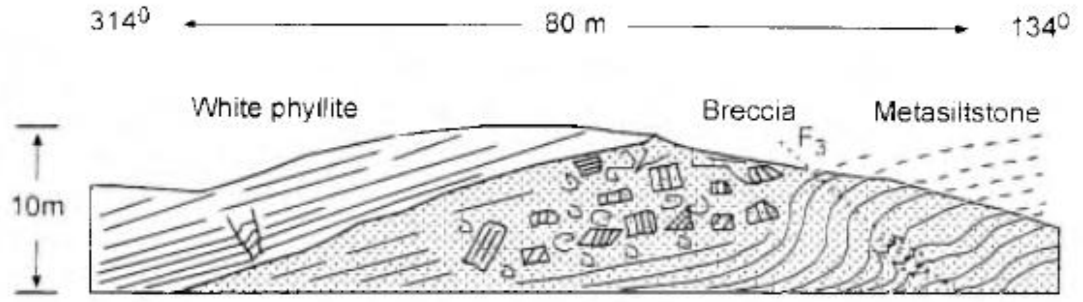

Fig. 8. Sketch displaying information collected during a detailed observation along a nearly 80 m-long, 10 m-high, man-made wall situated behind the tennis courts in Pousada park (outcrop 15). Basal metasiltstone layers topped by white phyllite dip gently to NW and display foliation S_1 (not shown) subparallel to S_0. A steeper mineral foliation S_2 (not shown) cross-cuts $S_1//S_0$ at low angle and dips westward. These rocks and structures are affected by a 10 m-scale and NW-verging F_3 fold in the SE part of the outcrop. The metasandstone layers were disrupted around the hinge of the fold, and the fragments underwent chaotic rotation to form a breccia. This fold indicates a top-to-the-WNW down-dip movement compatible with a m-scale, kink-style F_3 fold seen in the phyllite in the leftmost part of the outcrop.

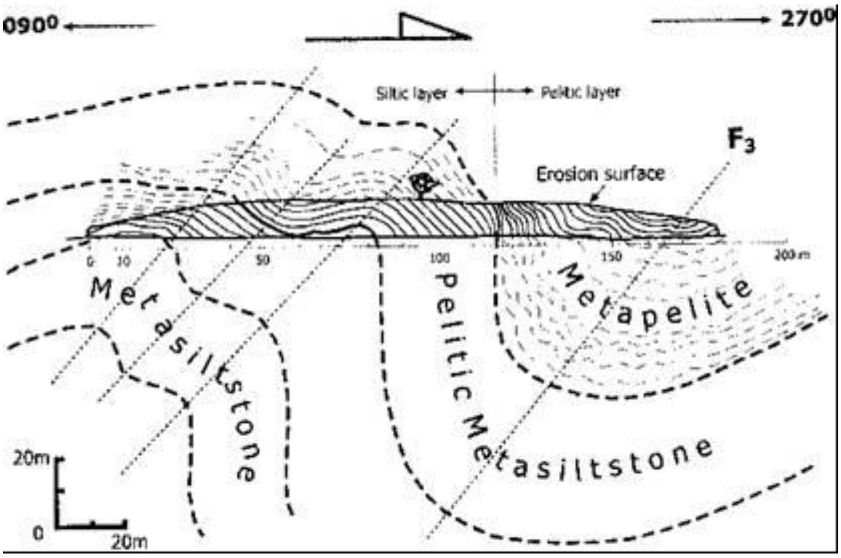

Fig. 9. Sketch of a hm-scale, W-verging F_3 fold observed in outcrop 7, NW of the dome, in the paved road to Morrinhos (Fig. 4). The limits of the road cut are drawn in the center of the figure. A succession of layers of metasiltstone, pelitic metasiltstone, and metapelite define the primary layering (S_0) that is affected by a layer-parallel foliation Si commonly marked by fine-grained white mica (biotite also occurs in the lowermost layer to the left), all affected by the S_2 foliation (mostly fine-grained white mica). The L_{2-1} intersection lineation is down-dip along the planes of S_1/S_0 and subparallel to the axis of several 10 cm-scale F_2 folds. Such planar and linear structures change dip or plunge either E or W along the respective limbs of the major folds, demonstrating that the larger structure is an F_3 fold. Fractures and quartz veins (omitted for simplicity) also occur but are less important. The entire structure indicates a top-to-W shear and, as in the fold in Fig. 8, fits the interpretation for a gravity slide during the Caldas Novas dome's uplift. The metapelite layer is more intensively folded to accommodate more intense shortening in the core of the major synform.

4.4 Brittle deformation: fractures and quartz veins

Fractures/joints and quartz veins that occur throughout the area and cross-cut D_1-D_3 structures in all rock types have been labelled S_4 (Fig. 6f and g) and interpreted as the record of a late event of brittle deformation (D_4). They are commonly sharp and well defined in outcrops of the Paranoa rocks but are more rare in outcrops of the Araxa rocks. The fractures are generally subvertical and statistically distributed in five main sets, v to w. Fractures v (the largest) and fractures *z* ~ *y* (the second largest) strike WNW and NNE, respectively, corresponding to directions subparallel and subperpendicular to the direction of L_x. Fractures x and w strike along intermediary directions between these two extremes and are nearly perpendicular to each other. Data in Fig. 6g define three main populations of quartz veins: Va, Vb, and Vc. The Va veins, which are subperpendicular to L_x, are generally thicker than Vbs, which are subparallel to the direction of L_x, and both are found everywhere. Veins Vc are parallel to the main population of S_3 and AP_3 (Fig. 6d) and more common in Paranoa quartzites of the eastern limb of the dome (outcrops 70 and 71, Fig. 6a). Only two such veins were found in outcrops 47 and 49 of Araxa, close to the SE and SW borders of the dome, respectively. The vein in outcrop 49 is as thick as Va veins (maximum thickness = 25 cm).

5. Tectonic evolution of the Caldas Novas dome

5.1 Structural evolution

The fact that S_1, S_0, and S_2 are subparallel, the ESE vergence of F_2 folds and their isoclinal style, the asymmetry of S-C (= S_1), and the down-dip stretching lineation all indicate a D_1-D_2 progressive event of intra- and interlayer slip that corresponds to a regional rotational deformation (assumed as simple shear) along surfaces dipping at low angle to the WNW—in other words, a typical regime of frontal ramps. Therefore, D_1-D_2 maximum compression *(S_1)* in the area was subhorizontal and took place along the same direction of the mean stretching vector (296°). Nevertheless, though D_3 records a progressive shortening of the area, the attitude of D_3 structures implies that the shortening direction shifted to WSW-ENE.

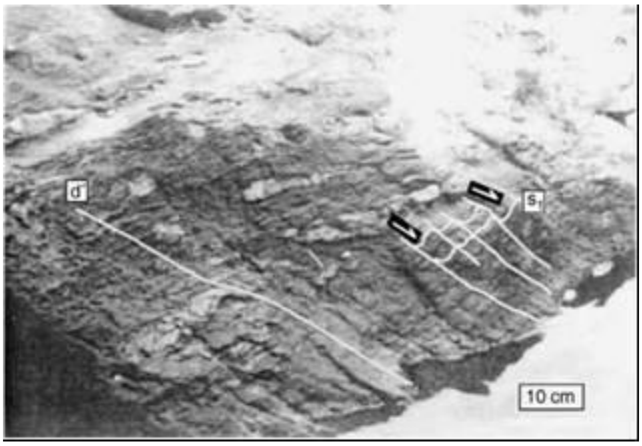

Fig. 10. Photograph taken in outcrop 2 (NE of the dome) to show the d^0 = 053748° extensional crenulation cleavage affecting planes of *S_1* = 340712° in the Araxa schist and implying a top-down to ENE slip. The outcrop is along a creek, a few meters away from the paved road.

The interpretation of WSW-ENE shortening during D_3 is strongly supported because the direction of L_x (D_1-D_2 stretching lineation) is the same on the western and eastern sides of the Caldas Novas dome. If, for example, the dome was formed initially as a consequence of WNW-ESE regional ductile flow and underwent rotation to its present NNW-SSE position (whether due to different intensities of the flow itself or the slip along WNW-ESE-striking lateral ramps north of the dome, as in Fig. 3), these processes would also affect L_x and the trace of $S_{0/1}$ and S_2, so the dome would appear as a rotated porphyroblast-like structure in regional maps (similar to the Itabaiana dome in the Sergipano belt, NE Brazil; D'el-Rey Silva, 1992; 1995a,b; D'el-Rey Silva and McClay, 1995). The complete lack of evidence of such rotation indicates that the final position of the Caldas Novas dome is the same as that in which it was formed.

Relative to D_4, it is notable that veins Vc display orientation and distribution compatible with the uplift of the dome. However, fracture planes labelled v are usually open, subvertical, strike parallel to the subhorizontal regional compression *(S_1),* dip parallel to the subvertical internal compression (S_{1i}), and host veins (Vb) that are subparallel to L_x. In addition, Va veins are hosted along fractures that are subperpendicular to L_x and, by playing the role of T fractures, are expected to develop in the same shearing regime. As shown in Fig. 6g, Va veins are concentrated in two populations of planes with similar strike but different dips (subvertically or steeply) to the ESE, which may be the result of the progressive emplacement and rotation of veins along T fractures. Finally, the parallel relationship between Vc veins and S_3//AP_3 match the direction of extension expected for the limbs of the dome during gravity-induced slides.

5.2 The Caldas Novas dome as a D_3 structure

On the basis of the unequivocal evidence found in the Paredao outcrop, it can be safely stated that the Caldas Novas dome formed during a D_3 deformation. The outcrop is continuous along a 3 m wide stream that flows down to the east in a cascade created by the rounded hinges of east-verging, steeply inclined, < 5 m scale F_3 asymmetric antiforms (Fig. 11a-d). These hinges affect Paranoa quartzite layers that display the S_1//S_0 foliation, as well as several ~ 1m scale F_2 isoclinal folds that refold around the F_3 hinges (Fig.

11e). The isoclinal style of the F_2 folds and the maintenance of their original vergence to the ESE demonstrate that the dome is a major F_3 antiform, not a large F_2 fold. Moreover, the vergence of the < 5 m scale F_3 folds supports their interpretation as gravity sliding structures related to the dome's uprise. If they were parasitic folds in the E-dipping limb of a major antiform, they would verge to the west.

Because of the gravity slide, foliation d^0 developed to extend the matter, whereas several F_3 folds developed to shorten it during the way down. The F_3 folds on the western side of the dome verge west; some may attain a scale of 300 m or more (Fig. 9), whereas others may imply the development of a collapse breccia (Fig. 8). The F_3 folds within the Araxa schists of the eastern border of the dome are hm scale but verge east, and their eastern limbs (closer to the dome) are associated with small-scale intense crenulation that may compensate for the amount of extension implied by d^0 foliation, which is very intense in outcrops of Araxa schists in the same area (e.g. outcrops 47, 41, 75; Fig. 6a).

The uplift of the dome modified the attitude of the D_1 —D_2 anisotropy ($S_0//S_1//S_2$) in such a way that the original dip, mostly to the NW, turned in all directions. However, despite this obvious effect, the strike of the stretching lineation was maintained across the area (Fig. 4a), and the plunge of L_x changed from 296° to the opposite direction (116°; Fig. 4d) with no significant variations to the S or N. A similar situation is observed for L_2 and $L_{S\text{-}C}$. In some outcrops along the western and eastern margins of the dome, the lines are down-dip and parallel to the regional L_x (e.g. 09, 17, and 40 to the W; 45 and 66 to the E). Together, these data indicate that the direction of stretching is coincident with the horizontal axis of rotation of the layers during uplift of the dome. Moreover, in outcrops along the northern (57, 59, 60) and southern (12; Fig. 4a) margins of the dome, L_2 and $L_{S\text{-}C}$ are parallel to L_x but display much lower rake and even become parallel to the strike of $S_0//S_1//S_2$. This situation is expected due to the tilt of $S_0//S_1//S_2$ planes and may be attained even with a small modification of their angle of dip.

5.3 Is the Caldas Novas dome a D_3 structure?

Together with the F_3 folds and foliation S_3, plus the similar conditions of ductility for d^0 and c^0 (which reflects great D_2-D_3 metamorphic compatibility), the heterogeneous distribution of d^0 in the area demonstrates that the dome is a D_3 structure formed in a progressive event following D_2. The fact that d^0 is characteristically absent on the western side of the dome works strongly against other possible interpretations.

Foliation d^0 might be interpreted as a D_2 structure that affected c^0 planes during progressive deformation. However, if this was the case, d^0 would be found everywhere in the area, and the uplift of the dome would affect d^0 in all of its quadrants (as happened with c^0). The dome also might be interpreted as a much younger structure (e.g. Cretaceous) due to a magma intrusion. In this case, the lack of a reasonable argument for d^0 as a syn-D_2 structure would lead to the interpretation that d^0 is a Cretaceous structure. Again, there is no reason for d^0 be absent along the west side of the dome; in addition, the similar conditions of ductility for c^0 and d^0 would require a warming up of the dome area in the Cretaceous. An igneous intrusion beneath the site of the dome could provide heat for such event, but it remains hard to accept that the Araxa rocks could be heated more intensively than the Paranoa rocks that lie closer to the hypothetical intrusion. Even accepting this, we also would have to accept that heat propagated heterogeneously upward and did not affect the Araxa rocks west of the dome. In addition to this negative coincidence, two other points work against the idea that the dome is due to a younger igneous intrusion: (1) Why would such a dome have to be elliptical? (2) Why would its longer axis have to trend exactly NNW, not in any other position? The elliptical shape and the trend of the Caldas Novas dome's longer axis both support D_3 evolution in the Brasiliano cycle and fit within the characteristics of a D_3 regional event existing in the southern segment of the Brasilia fold belt. The last possibility against the Brasiliano cycle D_3 evolution of the dome is even harder to accept, in that it considers d^0 as a Brasiliano D_2 structure and the uplift of the dome as due to a Cretaceous intrusion. If this happened, during event D_2 foliation d^0 did not form in a particular area, and the younger intrusion took place exactly at a point capable to make the area coincident with the west side of the dome.

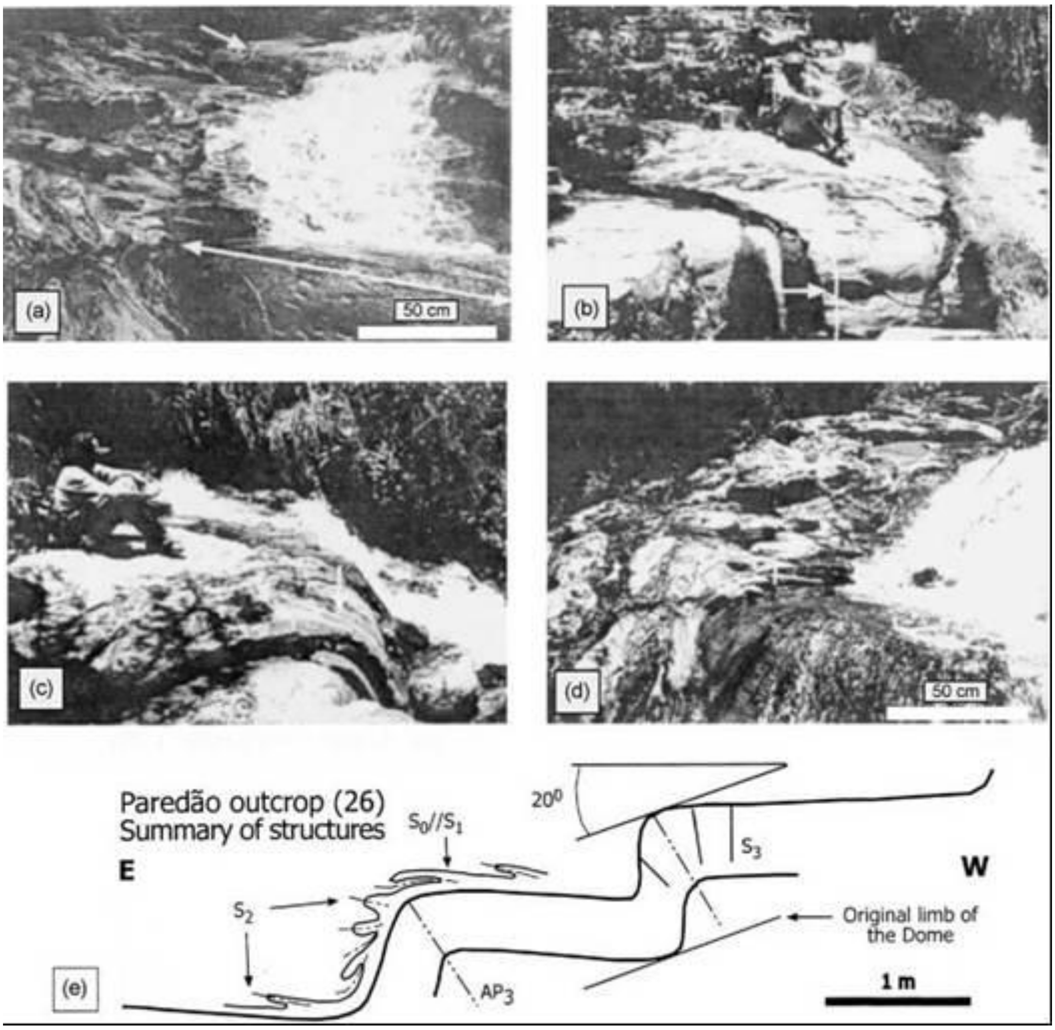

Fig. 11. The Paredao outcrop. (a) Cascade created by the hinge of m-scale F_3 folds. The white line marks the hinge of a major antiform in the foreground, and the arrow in the background points to a white tape laid down in the field around another F_3 hinge. Three other F_3 antiforms occur in between. The scale applies to the foreground. The hinge of the antiform (marked by the white tape) is seen in more detail in (b) and (c). (d) A closer view of the hinge of the main antiform. The F_2 folds systematically bend around the F_3 folds as depicted in (e). The envelope surface for the F_3 folds provides the original dip of the dome's eastern limb (around 20°).

The absence of d^0 in the western limb of the dome can be explained because during D_3, that limb remained in a position similar to that it had during D_2, as well as under compression due to the progressive regional ductile flow that counterbalanced the gravity effect and made it difficult for d^0 to develop there. Simultaneously, as the eastern limb rotated from a dip-to-WNW to a dip-to-E, it entered a zone more protected from the regional flow, so the gravity effect could act freely. Therefore, d^0 is abundant there. The extension to the N and S implied by d^0 fits in and reinforces this interpretation. The isolated occurrence of foliation d^0 in outcrop 10 (NW of the dome, Fig. 6a), which implies it extended to the S, may be attributed to the growth of the rolling hinge of the large F_3 folds seen in outcrop 7 (Fig. 9) located to the N. If this possibility is proven, it lends additional support to the interpretation that the gravity-induced folds are D_3 structures.

5.4 A model for D_3 evolution and underground thermal waters

Any model about the evolution of the Caldas Novas dome should take into account the possibility that the Paranoa quartzite actually extends down nearly 1000 m and should provide an explanation for this abnormal thickness. The 1000 m of quartzites could be due to deposition of voluminous siliciclastics in a particularly deeper depocenter, the site of the future Caldas Novas dome, as is the case for the Itabaiana gneiss dome of the Sergipano belt, where the basal quartzite formation is less than 50 m thick regionally but attains a thickness of 700 m around the dome as a consequence of a kind of half-graben associated with the hanging wall of a system of extensional listric faults (D'el-Rey Silva, 1992, 1995a; D'el-Rey Silva and McClay, 1995). The evolution of such a fault system led to the initial uprise of the dome during the sedimentation and volcanism of the precursor basin, and the incipient Itabaiana dome thus received a crown of carbonates covering the quartzites.

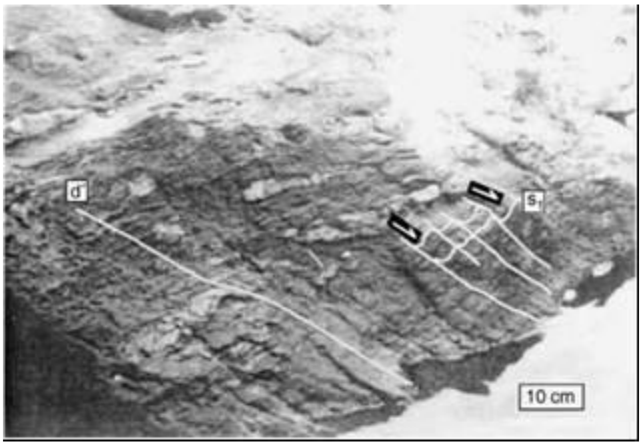

Fig. 12. Photograph taken in outcrop 2 (NE of the dome) to show the d^0 = 053748° extensional crenulation cleavage affecting planes of S_1 = 340712° in the Araxa schist and implying a top-down to ENE slip. The outcrop is along a creek, a few meters away from the paved road.

In contrast, the Caldas Novas dome area displays no lithostratigraphic information that might detail the possible tectonic evolution of the sedimentation in that part of the Brasilia belt; therefore, sedimentary control over the abnormal thickness cannot be demonstrated or ruled out. Nevertheless, a tectonic explanation for the abnormal thickness may be easily envisaged with the aid of out-of-sequence thrusts (Butler, 1987; McClay, 1992) and a D_3 duplex structure or antiformal stack (Boyer and Elliott, 1982) hidden in the core of the F_3 fold that corresponds to the dome (Fig. 12). However, because the study area lacks even a simple control over the regional thickness of the Paranoa group, this model is just one of several plausible ideas that vary according to the regional thickness.

The structural model for the Caldas Novas dome (Fig. 12c) agrees with the physical parameters required for the 50 °C hot springs, such as those shown near Pousada park (Fig. 6). A model for the evolution of the underground reservoir considers that meteoric waters would infiltrate regionally into the Paranoa group and percolate until they reached the bottom of the quartzite layers in the core of the dome, where they would attain the highest temperatures and then migrate quickly upward through a net of brittle structures that could combine planes of S_3 spaced cleavage and D_4 fractures. The model accounts for the nature and inventory of the D_3 and D_4 structures mapped at surface but cannot attain a higher degree of complexity until a very detailed geophysical survey combining, for example, seismic reflection, induced polarization, and hydrologic modelling provides additional information. Researchers working recently in Caldas Novas (e.g. Troger et al., 1999; Campos et al., 2000; Zschocke, 2000) agree that the uprise of thermal waters through different geological units in a complex hydraulic manner, controlled by the structural features observed in the area, leads to reservoirs with different chemical characteristics and temperatures.

It is possible that a younger magmatic rock (e.g. Cretaceous) was placed in the core of the D_3 dome (Fig. 12c), which enhanced the mechanical porosity by reopening at least part of the Brasiliano cycle D_3-D_4 brittle structures and facilitating the underground flow of waters. This hypothetical intrusion also could be the source of at least part of the heat required to warm up the waters. Again, only a very detailed geophysical study could provide reliable information about whether such an intrusion exists or not. The results of recent geochemical and isotope studies (Zschocke, 2000; U. Trogger, pers. comm.) indicate no contribution from a magmatic intrusion for the composition of thermal waters in the Caldas Novas area.

6. Implications for understanding the Brasilia belt and Tocantins province

The structural and tectonic analyses of the Caldas Novas area suggest the following:

1. The D_1-D_3 (D_4) evolution described may be bracketed in the ~ 750-600 Ma interval, according to the age of syntectonic granitic intrusions located to the NW and E of Ipameri, as well as of granulite facies metamorphism near the study area;

2. The D_1 -D_2 tectonics of frontal ramps with S_1 oriented to the WNW-ESE in the Caldas Novas area is the same that has been demonstrated in Ipameri, as well as in other parts of both the southern and northern segments of the belt along the magmatic arc-continental plate margin or closer to the Sao Francisco craton (Fig. 1b; D'el-Rey Silva and Barros Neto, 2002). Thus, despite the bend of the Brasilia belt, there is a strong coincidence in the WNW-ESE direction of the regional stretching lineation and maximum horizontal compression (S_1) for D_1-D_2 in both the northern and southern segments of the belt;

The Caldas Novas dome is a D_3 antiformal structure, implying a WSW-ENE-trending shortening, remarkably coincident with the direction of shortening in the F_3 folds that affect the D_1-D_2 Araxa, Passos and Luminarias nappes in the southern segment of the belt (Fig. 2; Seer, 1999; Seer et al., 1999; Seer and Dardenne, 2000; Valeriano et al., 2000);

3. Such a D_3 event is regionally absent in the northern segment of the belt. The single well-documented example of WSW-ENE D_3 shortening in the northern segment of the belt is a local phenomena and related to the escape of the rocks to the SSE during formation of D_2 sheath folds that control emerald ore shoots in the mining district of Campos Verdes in the MA (Fig. 1b; Barros Neto, 2000; D'el-Rey Silva and Barros Neto, 2002; D'el-Rey Silva, 2002b);

4. The age of D_3 deformation in the southern segment of the belt must be close to 600 Ma (Pimentel et al., 1999 for the Ipameri area; Seer, 1999; Valeriano et al., 2000 for the southernmost part of the belt; Seer (1999) for the structural analysis), which fits the age of opening of the rift-like basin precursor of the Paraguay belt (Alvarenga and Trompette, 1992; Pimentel et al., 1996; Alvarenga et al., 2000); and

5. The direction of D_3 shortening fits with the movement of the Socorro-Guaxupe nappe that evolved in the 620-510 Ma interval and displaced high-P basement rocks onto the F_3 regional folds that affect the Araxa, Passos and Luminárias nappes (Fig. 2; Valeriano et al., 2000).

We have avoided correlating the WSW-ENE direction of shortening (D_3) with a particular stress field because the likelihood that the NNW-SSE trend of the Caldas Novas dome (the F_3 antiform) records the same position in which it formed originally does not suggest a straightforward similarity between the direction of shortening and the direction of maximum compressive stress. In addition, the Caldas Novas dome F_3 antiform did not form in the same NE-SW trend of the nonrotated F_2 folds in the study area, which does not authorize a denial that the WSW-ENE D_3 shortening could be due to the same WNW-ESE subhorizontal stress field (sj) active since the Dj-D_2 events. However, to correlate the WSW-ENE shortening with the WNW-ESE-trending S_1, the three following hypotheses should be considered: (1) D_3 shortening is due to different velocities of the WNW-ESE flow along the planes of S_0/Si/S_2, (2) D_3 shortening was due to S_2, and (3) The rocks in the study area (and in the Araxa, Passos, and Luminarias nappes) underwent D_3 shortening due to WNW-ESE flow that took place in the basement.

Hypothesis 1 seems quite unlikely because the basement crops out as a dorsal between the F_3 synforms that affect the Passos and Luminarias D_1-D_2 nappes (Fig. 2), which indicates that nappe emplacement involved the crystalline basement of the Sao Francisco craton, at least south of the Caldas Novas dome, and the whole vertical column of the DJ-D_2 nappes underwent D_3 shortening. If not involved in the nappes, then the basement marks the hinge of an intervening antiform between two F_3 synforms, or it crops out because of erosion. The first case requires a D_3 detachment within the basement, which is most reasonable because D_3 is a regional event. In the second case, the D_3 detachment lies along the basement-cover interface, and it would have had to jump exactly the narrow site of the basement between two F_3 synforms. Hypothesis 2 also is hard to accept because S_2 alone would have to shorten a thicker pile of rocks after D_1 -D_2. Folds due to S_2 must form simultaneously with folds due to S_1, not afterwards. Hypothesis 3 implies that the D_1-D_2 regional flow continued solely in the basement during D_3, without a detachment between two crustal levels. In addition, the WNW -ESE flow in the crystalline basement would have to be contemporaneous with D_3 in the upper level. To date, ductile flow in the basement during D_1-D_2 does not indicate that the basement was part of the D_1-D_2 nappes. Furthermore, if Hypotheses 1-3 are true, why is a regional WSW-ENE D_3 shortening regionally absent in the northern segment of the belt?

Therefore, most likely D_3 event involved the basement, and the maximum compression responsible for the D_3 structures was not the same as that responsible for D_1-D_2 events. It sounds interesting to interpret another maximum compression (e.g. σ^{1}_{3}) as parallel to the direction of D_3 shortening. Nevertheless, despite its existence, the younger field σ^{1}_{3} does not require that the older one became inactive in the late stage of the evolution of the southern part of the belt. If so, which tectonic scenario could account for coexisting stress fields, and why should the effects of the WSW-ENE field overcome the effects of the WNW-ESE one? Instead of competing with S_1, the younger maximum compression σ^{1}_{3} may have added to the effect of S_2 during D_3 and facilitated the WSW-ENE shortening. Of great relevance is that the southern segment of the Brasilia belt displays clear evidence for a regional event of WSW-ENE shortening that affected both basement and cover and was active for approximately 100 Ma, progressively overprinting D_1-D_2 structures that formed across the belt as a consequence of a WNW-ESE-driven tectonic transport during ~ 750-600 Ma. These two distinct events, their characteristics, and the preceding questions all demand a comprehensive tectonic analysis of the Tocantins Province that is beyond the scope of this paper.

7. Conclusions

The Caldas Novas dome is an elliptic structural window with a ~ 20 km long maximum axis trending NNW, cored by subgreenschist facies rocks of the Meso-Neoproterozoic Paranoa group and mantled by a nappe of greenschist facies (biotite zone) rocks of the Neoproterozoic Araxa group. These rocks underwent three (D_1-D_3) events of mainly ductile progressive deformation in the ca.750-600 Ma interval and indicate typical tectonics of frontal ramps. A set of brittle fractures and veins characterizes the youngest event recorded in the area. Both D_1 and D_2 record an event of regional simple shear along surfaces that dip gently to the NW, with a top-to-SE relative displacement and a maximum subhorizontal compression (S_1) oriented 290-300°/110-120°. The Araxa rocks experienced more severe metamorphism than did the Paranoa rocks during D_1 and were emplaced as a nappe during D_2. Event D_3 implies additional shortening and records a WSW-ENE-driven compression, during which the dome started to rise as a large- scale F_3 fold, possibly associated with a duplex structure at depth. Continuous with the uplift of the dome, the layers slid down, giving way to gravity-slide D_3 structures, such as extensional crenulation cleavage, folds, and tectonic breccia heterogeneously distributed across the area. The underground reservoir likely is fed by meteoric waters that percolate through the Paranoa group, attain temperatures of more than 50 °C at a depth of 1000 m, and travel quickly upward through a mechanical porosity consisting of a net of D_3 and D_4 planar brittle structures. Detailed geophysical work could provide additional information. The D_1-D_2 structural and tectonic evolution of the Caldas Novas dome area is similar to several other areas of the whole Brasilia belt and took place in the ~ 750-620 Ma interval. Moreover, the dome itself demonstrates the importance of the 620-510 Ma old D_3 event of

WSW-ENE shortening that is evident across the southern segment of the Brasilia belt and virtually absent in the northern one.

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