Coeval perpendicular shortenings in the Brasilia belt: Collision of irregular plate margins leading to oroclinal bending in the Neoproterozoic of central Brazil

Luiz J.H. D’el-Rey Silva
Italo L. de Oliveira
Camila B. Pohren
Maria Luiza N. Tanizaki
Rodolfo C. Carneiro
Gabriel L. de F. Fernandes
Priscila E. Aragão

ABSTRACT

The three belts which form the Tocantins province (central Brazil) records Neoproterozoic-EoPaleozoic collisions involving the Amazon and São Francisco paleocontinents and the Paraná continental block. The Brasília belt is a typical orocline bended around the WNW—ESE striking Pirineus Zone of High Strain (PZHS) and is comprised of the NE-trending (northern) and SE-trending (southern) segments. The Brasília dome is an N—S elliptical structural window situated in the center of the belt, at the eastern end of the PZHS. It evidences $D_1$—$D_2$ and $D_{3N}$ shortenings (~ 750—590 Ma) due to ocean closure and Amazon-São Francisco collision following a WNW—ESE path, and demonstrates similar evolution for both segments of the belt. However, in the southern segment, $D_1$—$D_2$ structures are deformed by shortening in the SW-NE direction ($D_{3S}$). New data demonstrating $D_1$—$D_2$ and $D_{3N}$ tectonites deformed by $D_{3S}$ struc- tures in the area close to the dome’s SW margin and SE of the PZHS support understanding the Brasília belt and oroclinal bending as a consequence of the collision of two (Amazon and São Francisco) irregular continental margins leading to separation-rotation of the Paraná block from the Amazon paleocontinent and the Paraná-São Francisco collision.

Keywords: Structural analysis Brasília; belt Tocantins province; Irregular continental; margins Paraná block; Oroclines
1. Introduction

The Brasília belt, a typical orocline, forms together with the Araguaia and Paraguay belts, the Tocantins province, the main orogenic unit stretching between the Amazon and São Francisco cratons, in central Brazil (Fig. 1a; Almeida et al., 1981). The province derives from the collision of the Amazon and São Francisco paleo-continents with participation of the Paraná continental block, and is a remnant of the ~750—510 Ma Brasiliano orogenic cycle responsible for amalgamation of the western Gondwana supercontinent (Trompette, 1994). The Paraná block comprises the lithosphere flooring the Paraná basin (Fig. 1a; Mantovani et al., 2005).

The Brasília belt comprises three fault-bounded longitudinal lithotectonic domains: the Magmatic Arc, the Internal Zone, and the External Zone (Fig. 1b; Fuck et al., 1994, 2006). The Distrito Federal of Brasília (DF) lies in the centre of the External Zone (EZ). Approximately at the latitude of Brasília, the belt becomes divided into the NE-trending northern segment and the SE-trending southern segment (Marini et al., 1984). The axial surface of the orocline is along the Pirineus Zone of High Strain (PZHS), a WNW—ESE striking narrow zone of high strain coincident in the field with an array of magnetic lineaments [~300-km-long; ~10-km-wide; aeromagnetic data in Blum (1995)]. The PZHS rotated the lithostructural grain ($S_0$—$S_1$/S$_2$) in the rocks of both the Internal and External zones, but its effects die out close to the western margin of the Brasília dome, which is the main geological feature inside the DF area (Fig. 2).

For three reasons the evolution of the Brasília dome, oroclinal bending and the PZHS must have been closely related: (1) in despite of orocinal curvature, a vast inventory of structural data (D’el-Rey Silva and Barros Neto, 2002; D’el-Rey Silva et al., 2004, 2008; other references therein) show the persistence of $D_1$—$D_2$ ductile flow in the WNW—ESE direction across the belt, north and south of the PZHS; the direction for $D_3$ shortening is the same for $D_1$—$D_2$ in the northern segment, whereas in the southern segment the $D_3$ folds, foliations and faults which deform $D_1$—$D_2$ tectonites record SW-NE contraction (Fig. 1b); (2) the structures deforming the rocks in the Brasília dome and surroundings (Freitas-Silva and Campos, 1995; D’el-Rey Silva et al., in
preparation) record progressive $D_1-D_3$ ductile flow and contraction in the WNW—ESE direction; and (3) the structural data in the area SW of the Brasília dome (this study) define a Box-junction Zone (Fig. 2) inside which is evident the superposition of structures due to shortening oriented SW—NE, typical of $D_3$ in the southern segment of the belt, upon $D_1-D_3$ structures which imply progressive WNW—ESE shortening typical of both the dome and the northern segment of the Brasília belt.

Fig. 1. a—b: Simplified map (a) emphasizing the lithotectonic units concerning to the evolution of the Tocantins Province: The Araguaia, Brasília and Paraguay belts, the Amazon and São Francisco cratons, and the Paraná and Parnaíba Phanerozoic basins. The simplified map (b) of the central part of the province (based on Marini et al., 1984; Fuck et al., 1994, 2006; Dardenne, 2000) emphasizes the lithotectonic domains and the bend of the Brasília belt. Arrows for $D_1-D_2$ and $D_3$ regional shortening are based on detailed studies (D’el-Rey Silva and Barros Neto, 2002; D’el-Rey Silva et al., 2004, 2008; and references therein). Arrows for tectonic vergence are from plate-scale syntheses (e.g., Marini et al., 1984; Trompette, 1994). Areas J, K, and O are discussed in text.

The data bring to light compelling evidence that two perpendicular shortenings co-existed during $D_3$, in space-and-time, here-in-after referred as $D_3N$ and $D_3S$ respectively for shortenings oriented WNW—ESE and SW—NE.

This study complements the structural analysis recently carried out across the DF area, after which the information collected in other 50 outcrops (Fig. 2) provided information enough to constrain the structural evolution of the
Brasília dome.

Coupled with a larger database which includes published U-Pb age data of syn- and post-tectonic granites throughout the Brasília belt, the new structural data support previous interpretation that oroclinal bending and the PZHS fit very well in events related to the first stage of evolution of the Paraná block and Paraguay belt (D’el- Rey Silva et al., 2004, 2008); place the evolution of the Brasília dome as slightly older than the PZHS; and allow interpreting the Brasilia belt and the Tocantins province due to the collision of irregular continental margins.

2. Summary of the regional geology and tectonic setting

The Brasília belt includes (Fuck et al., 1994, 2006; Dardenne, 2000; or as otherwise referenced): 1 - a continental crust basement comprising Archean and Paleoproterozoic high- and low- grade rocks, as well as metavolcano-sedimentary rocks of an event of intracontinental rift ca. 1.75 Ga ago; 2 - metasedimentary rocks of the Paranoá and Canastra Groups respectively record proximal- and distal-like sequences of a Meso-Neoproterozoic passive continental margin that existed along the São Francisco paleocontinent. They occur in both the External and Internal Zones (Fig. 3; Moreira et al., 2008; D’el-Rey Silva et al., 2004, 2008); and 3 - amphibolite-greenschist facies metamorphites derived from intrusive and volcanic and sedimentary rocks which compose the (intra-oceanic) Magmatic Arc and also the Araxá Group (back-arc basin; Fig. 3). Sm-Nd isotopes point to at least two superposed island arcs and back-arc basins, and for arcs- continent collage ca. ~750 Ma and 645—640 Ma, when actually started the collision involving the Amazon and São Francisco paleocontinents (Fuck et al., 1994, 2006). Metacarbonates and metasiliciclastic rocks (Bambuí and Ibiá Groups) complete the Neoproterozoic record and represent sedimentary input from the arc and the craton.

Granulite facies rocks displaying ~750 and ~645—635 Ma ages of high P-T metamorphism record subduction zone conditions. The older granulites occur in a series of three mafic-ultramafic layered complexes north of the PZHS (a very small part of the southernmost of these complexes, named Barro Alto, is
seen in the NW corner of the map in Fig. 3). The younger are represented by ortho- and para-derived granulites which occur (Della Giustina, 2010) in the Uruaçu Complex (northern segment of the belt) and in the Anápolis-Itauçu Complex (Fig. 3).

\[ D_1 - D_2 \]

D\textsubscript{1}—D\textsubscript{2} nappes of Canastra and Araxá rocks propagated towards ESE, onto the San Franciscan margin, and surround the DF area (Fig. 3; Faria, 1995; Faria et al., 1997). In the southern and southernmost parts of the Brasília belt the D\textsubscript{1}—D\textsubscript{2} nappes propagated lately, \( \sim 640—635 \) Ma ago (Valeriano et al., 2000, 2004) coevally with the younger even of granulitization.

---

Fig. 2. Simplified structural map emphasizing the lithostructural grain \( (S_0 - S_1 - S_2) \); the Brasília dome (outlined by heavy-black lines); and the easternmost part of the Pirineus Zone of High Strain (PZHS). Note the 24 outcrops studied in detail (this paper) and other 50 outcrops described previously. The Box-Junction Zone is explained in text.
It is well-accepted that the Brasilia belt records (Fuck et al., 1994, 2006; Dardenne, 2000) polydeformation of Neoproterozoic age, whereas the southernmost end of the Brasilia belt was completed solely in EoPaleozoic times. In fact, in that part, the high P-T basement rocks of the Socorro-Guaxupé nappes propagated to ENE, 550—510 Ma ago, onto \((D_1 - D_2 - D_3S)\) Neoproterozoic tectonites (Fig. 4; Valeriano et al., 2004; Campos Neto and Caby, 1999). The youngest time interval is basically the same for deformation in the Paraguay belt and also fits the obduction of the ~750 Ma old Quatipuru ophiolites onto a ~ 540 Ma old carbonate platform in the western margin of the Araguaia belt (Paixão et al., 2008).

3. Geological background for the Distrito Federal area

3.1. Lithostratigraphy overview

The DF area encloses metasedimentary tectonites of the Paranoá, Canastra, Araxá and Bambui Groups (Fig. 3; Faria, 1995; Faria et al., 1997; Freitas-Silva and Campos, 1995, 1998). The \(D_3N\)-related Brasilia dome (Freitas-Silva and Campos, 1995; D’el-Rey Silva et al., in preparation) is cored by sub-greenschist facies rocks (Paranoá Group) and mantled by low-greenschist and greenschist facies rocks (Canastra and Araxá nappes, respectively). Rocks of both nappes occur imbricate in the SW surroundings of the DF area (Fig. 3). Low-grade metacarbonate rocks (Bambui Group) spread from the eastern limb of the Brasília dome towards the EZ- São Francisco craton border and beyond.

The **Paranoá Group** is a psamo-pelitic and carbonated regional sequence comprising slate, meta-argillite, metasiltite, quartzite and metacarbonate divided into several units (Faria, 1995) though in regional-scale maps, even around the Brasília dome, solely two or three units have been individualized, however (Moreira et al., 2008).

The **Canastra Group** has been divided into two sequences, one
comprised of Quartz-sericite-chlorite phyllite with quartzite intercalations, another of carbonaceous phyllite with thin intercalations of mica-quartzite and metasiltite (Fig. 3; Moreira et al., 2008). In the DF area, the Canastra Group consists of yellowish-brown to light-grey colored quartz-sericite-chlorite phyllite locally exhibiting intercalated layers of thinly laminated quartzite, local carbonate-bearing schist, and lens of metacarbonate.

The Araxá Group comprises grey-green color muscovite-biotite schist (locally garnet-bearing) with lenses of quartzite and carbonate, and bodies of sheared metamafic igneous rocks.

3.2. Structures and structural evolution of the Brasília dome

After studying the Paranoá rocks in the Brasília National Park (northern part of the DF area; Fig. 3) Freitas-Silva and Campos (1995) concluded that rocks in the Brasília dome record: (1) three phases (D₁—D₃) of overall E—W regional shortening, and a phase of N—S shortening (D₄); (2) local shear zones of small or negligible strike-slip (D₅); (3) fractures and normal faults of minor effect (D₆). Freitas-Silva and Campos (1995,1998) defined a larger F3 synform controlling the Canastra nappes inside the DF area (Fig. 3) and interpreted the Brasília dome as a feature of F3 x F4 folding interference.
Fig. 3. Simplified geological map of part of the central-southern Brasilia belt (based on Moreira et al., 2008) showing the same outcrops (as in Fig. 2) and emphasizing: the border between the Internal and External Zones; and geological
features inside the eastern part of the PZHS (imported from Fig. 1 in Araújo Filho, 2000), and emphasizing: the border between the Internal and External Zones; and geological features inside

![Simplified geological map of the southernmost part of the Brasília belt](image)

Fig. 4. Simplified geological map of the southernmost part of the Brasília belt (adapted from Valeriano et al., 2000; D’el-Rey Silva et al., 2004) emphasizing the directions of shortenings D₁—D₂ and D₃S, as well as the EoPaleozoic nappes.

In reality, structural studies carried out in 100’s of outcrops during the last decade across the External Zone (Fig. 1b; rectangles J, K, and O) plus the structural analysis based on 50 outcrops across the DF area (Fig. 2; Lemos et al., 2008; Silva et al., 2008; D’el-Rey Silva et al., in preparation) produced a large database which demonstrate a far more complete evolution.

The database includes: minor-scale structures, such as folds and/or spaced cleavage (pressure solution or crenulation type) deforming the limbs of some F₂ folds as well as quartz or carbonates veins, so that veins and cleavage parallel each other and co-exist within the of length of 10—20 cm in the same layer, all indicating minor contraction and coeval minor extension due to subordinated N—S movements in direction coincident with the geometric axis of F₂ folds (b tectonic axis); together with the D₁—D₂ features, such minor structures also occur deformed by F₃N folds and axial plane foliation; the limbs, hinges and axial plane foliation of F₃N folds also record coeval contraction-
extension in the direction of the (F₃N) b tectonic axis; some tight F₂ and F₃N folds locally display the axial surface cut-across by sub-horizontal veins and spaced-or-crenulation cleavage, evidence for local contraction-extension in the vertical direction, compatible with coeval minor movements along the respective a tectonic axes; the F₃N folds, usually open-tight and normally strongly-inclined to the west, across the area, are rather isoclinal folds with sub-vertical limbs both affected by minor-scale sub-vertical shear zones subparallel to the gently-plunging hinges, in parts of the western margin of the Brasília dome.

Since then, the structural evolution of the DF area has been explained in terms of progressive ductile deformation according to D₁—D₂—D₃N phases of WNW—ESE regional shortening coupled with nearly N—S movements in consequence of, respectively, D₂- and D₃N-related tectonic escapes (D₂TE and D₃NTE) along the b and a tectonic axes of both the F₂ and F₃N folds, these escapes meaning a kind of transpression deformation of very local/minor importance.

Moreover, the data show that the F₃NTE folds are very tight solely along the northern part of the Brasília dome, indicating the latest N—S horizontal shortening (D₃NTE) strong in that part of the area and vanishing progressively to the south, across the dome, probably requiring the addition of another factor to the simple tectonic escape.

The greater magnitude of the latest N—S shortening in the northern part of the Brasília dome has been attributed to the uplift of crystalline basement rocks in the core of large-scale D₃N hinges in the area of Cavalcante, ~300 km north of Brasília (Fig. 1b). This interpretation is highly recommended because high-resolution satellite images (Moreira et al., 2008) display NNE-SSW continuous hinges of F₃N regional folds plunging to SSW, all across the northern segment of the External Zone, and show these hinges shortened N—S by F₃NTE-like folds solely in the narrow corridor limited to the south and north respectively by the towns of Brasília and Cavalcante, and to the east and west respectively by the areas labeled J and K (Fig. 1b). The data from the two latter areas show the lack of latest N—S shortening of large magnitude.
Most probably the uplift of the crystalline basement in the core of the large-scale F₃N antiforms promoted gravitational slide of part of the mechanically anisotropic stratigraphic section, down the SSW-plunging F₃N hinges still being formed, so the Paranoá layers folded tightly against the barrier created by the still-rising northern margin of the Brasília dome.

4. Structures and structural evolution in the area SW of the Distrito Federal (this study)

4.1. Summary of structures

The detailed description of the 24 outcrops allowed collection of ca. 340 attitudes of tectonic structures (Fig. 5 a—i). Key geological aspects observed in some outcrops are shown in Figs. g—11. The tectonic structures record a polyphase deformation consisting of D₁—D₂ ductile flow and D₃ shortenings. Nevertheless, D₃ structures characterize both shortenings in the WNW—ESE direction (D₃N), typical of the northern segment of the Brasília belt, and in the SW-NE direction (D₃S) which is exclusively found in the southern segment.

Structures in outcrops of the SW limb of the Brasília dome (17, 18,19, 23 and 24; Fig. 2) confirm the D₁—D₂—D₃N evolution of the DF area and that the late N—S shortening associated to D₃N actually vanishes towards the south. Such outcrops do not display any evidence of SW—NE shortening (D₃S) either, therefore establishing a northern limit for the occurrence of D₃S structures. Similarly, a southern limit can be established for the occurrence of F₃N folds, on the basis of the lack of these folds in a series of eleven outcrops (numbered 6—16) west of Luziânia. Foliation S₃N barely occurs in just two of these road-cuts (Figs. 10 and 11), however.

4.2. Phase D]}

This phase is mostly represented by foliation S₁ and rare F1 folds. They deform the sedimentary bedding (S₀) which is easily recognized in the outcrops. In Paranoa siliciclastic rocks, S₀ is defined by the intercalation of compositionally distinct layers. In carbonate rocks, the best indicators are mm- to dm-thick layers of metamarl interleaved with layers of metacarbonate.
Fig. 5. a–i: Statistical distribution for D]—D₃, D₃N and D₃S structures. More details in text.

Foliation Si is sub-parallel to S₀ and varies in intensity and morphology. In rocks of the nappes, Si] is the most prominent planar feature, but the definition of S₀ is possible because these rocks commonly exhibit intercalated strata (quartzite or metacarbonate) parallel to the layers of schist or phyllite. In these rocks, S₁

is commonly marked by micaceous minerals (chlorite, white mica) respectively sized ~ 1 mm and ~ 1 cm. Biotite is also found along Si in the Araxa schist. S₃ commonly displays S—C geometry, with the plane of flow (C) coincident with the bedding. Lens-shaped, mm- to cm- thick bodies of quartz
may be found along C surfaces (e.g. in outcrop 6).

Identification of $S_1$ was not priority in Paranoa rocks because it has been done in several outcrops across the dome (D’el-Rey Silva et al., in preparation; Freitas-Silva and Campos, 1995, 1998). The few examples of F1 folds (seen inside the dome) are isoclinal and intrafolial relative to $S_0$.

Fig. 6. Sketch of the southern road-cut in outcrop 16 (805971 E; 8197733 N; Zone 22S) to show a 100 m-sized isoclinal fold nappe ($F_2$) shortened SW-NE by open to gentle-style $F_3$ folds. Note the Araxa schist structurally below the Canastra phyllite in the overturned limb.

Fig. 7. a-e : Main tectonic structures and structural evolution portrayed in outcrop 24, a ~10 m² map-like exposure (812915 E; 8223513 N; Zone 22S). A ~20 cm-thick layer of quartzite (part of a sequence of Paranoa metarhythmite; not ornamented) is deformed by tight $F_{3N}$ folds (a) which also deform foliation $S_2$ (detail in b). The out-of-scale diagram (c) reconstructs the $D_1$—$D_2$ template before $F_{3N}$ folding, whereas the actual field situation (in d) allows envisaging the outcrop as part of a larger $F_2$ fold refolded by $F_{3N}$ (e).

4.3. Phase $D_2$

$F_2$ folds are tight-isoclinal, asymmetric, gently-moderately inclined, with gently- to moderately-plunging fold axis (B2). They are commonly 10's of centimeters to few meters sized, though the detailed reconstruction in outcrop
16 (Fig. 6) shows that some may reach up to 100’s of meters, with thinned overturned limb typical of fold nappes (e.g. McClay, 1992).

Foliation $S_2$ is very penetrative and also varies morphologically. It is a slaty cleavage in Paranoa metapelite; a finely-spaced cleavage in metasandstone, metacarbonate and quartzite, in certain cases associated with fine-grained white mica; or a very fine-grained mineral foliation (commonly marked by very small grains of white mica) or a crenulation cleavage in the Canastra phyllite and Araxa schist.

The similar statistical distribution of $S_2/\text{AP}_2$ and $S_0/S_1$ (Fig. 5 a—b) evidence how tight, asymmetric and ESE-verging are the $F_2$ folds. It also indicates ductile flow $D_1$ in the same ESE direction of mass (and nappe) propagation implied by the $F_2$ folds and permits envisaging that $S_0$ and $D_1$—$D_2$ structures underwent similarly-oriented shortenings afterwards ($D_3$ phase).

The orientation of the axial plane ($\text{AP}_2$) and axis $B_2$ may vary largely in one outcrop, or from one to another outcrop, in consequence of $F_2$ sheathing and/or refolding. Sheaths of dm- to m-scale are observable in Canastra phyllite of outcrop 17. The sheaths define, together with striations seen on the surface of layers (of quartzite, in particular) or large mica grains, a stretching lineation down-dip on $\text{So}/\text{Si}$ and $S_2$ surfaces ($L_x$; Fig. 5d). The penetrative striation seen on the Canastra quartzite of outcrop 10 (Fig. 11) resembles the stripping lineation described by Sengupta and Ghosh (2007) therefore it may be taken as evidence of intense ductile flow to ESE, during nappe propagation.

Few normal faults were noticed. They display straight and ~5 m-long traces and may be taken as evidence of progressive ductile flow during the propagation of the nappes, because their attitudes and the down-to-WNW (outcrop 17) or down-to-ESE movement (outcrops 12 and 7) of their hanging wall fit well in the finite ellipsoid of $D_2$ strain, in which the fault planes play the role of T-type fractures further rotated and used as planes of local slip during progressive flow. Moreover, the Araxa schist in outcrop 13 clearly exhibits the pair $S—C = S_1$ deformed by extensional crenulation cleavage indicating top-down-to ESE slip, candidates for playing Riedel shears or c’ foliation (e.g.,
Lister and Snoke, 1984) associated to $D_1$ deformation. Nevertheless, because their spacing ($>10$ cm) is much larger than the cm-scale of the $S$—$C$ pair in the outcrop, and because they cut several $C$ (=$S_1$) planes, they rather suggest development during progressive $D_2$ flow (nappe propagation).

Fig. 8. a–d: $D_2$, $D_3N$ and $D_3S$ structures deforming Paranoa metarhythmite in outcrop 21 (811740 E; 822,870 N; Zone 22S). An overall view of the outcrop (a) showing, in the wall beyond the stream, the dam-sized hinge of an up-right $F_{3N}$ fold deforming 10 cm-thick layers of metasiltite intercalated with thinner layers of metapelite. The flat floor (foreground) exhibits the traces of $S_{3N}$ and $S_{3S}$ spaced cleavages. A more detailed look permits (b) observation that the $F_{3N}$ fold deforms the long limb of an ESE-verging and highly-asymmetric $F_2$ fold, the hinge of which is cut by the typical $S_2$ slaty cleavage (see inset). On the opposite side of the stream (c) the $F_{3N}$ hinge keeps evident (background) whereas the outcrop’s flat floor exhibits vertical surfaces of the axial planar foliation $S_{3N}$ (white dots) and the $S_{3S}$ cleavage (black arrows; detail in d) axial planar to $F_{3S}$ folds that shorten both $S_{3N}$ and $B_{3N}$ (fold axis). Fold axes $B_{3S}$ are sub-vertical or gentle, depending whether the $F_{3S}$ hinges deform surfaces of $S_{3N}$ or layers ($S_0$). Because of $F_{3S}$ folding, the surfaces of $S_0$ and $S_2$ both dip to N—NE in the northern part of the outcrop (behind the authors; c) and to S—SW in the southernmost part (see in a). The white arrow in (a) stands for the axis $B_{3S}$.
Fig. 9. a—c: Diagrams illustrating key structural relationships in Canastra rocks of a 50 m-long road-cut, outcrop 4 (813017 E; 8219865 N; Zone 22S). Open to gentle-style F3S folds and the S3 axial planar foliation are evident on the wall of the road-cut (a). F2 isoclinal folds are noticed refolded by F3S (detail in b). The 3-D diagram in (c) summarizes the relationship in the outcrop: F3N folds refold D3—D2 structures and F3S folds refold D1—D2 and D3N. Structures D3N are more difficult to see because they strike oblique at low-angle with the road-cut wall.

Fig. 10. Sketch of F3S folds deforming rocks of the nappes in the southern part of the study area (outcrop 14; 804659 E; 8196079 N; Zone 22S). The folds are outlined by a 1 m-thick packet consisting of dm-thick quartzite layers interleaved with schist (above) and phyllite. Foliation S3N is restricted to the quartzite layers and occurs also shortened by open-style, generally dm-scale F3S folds.

4.4. Structures evidencing WNW—ESE (D3N) shortening

Structures recording WNW—ESE shortening comprise mostly the folds (F3N), their axial planar foliation (S3N) and thrust faults. They are common and evident solely in outcrops closer to the Brasília dome. Moreover, F3N folds are generally tight-open in outcrops 17-21 and 24, nearest the dome, but become
gentle-open, away from the dome.

The $F_{3N}$ folds developed a NEeSW trending and steeply-dipping or sub-vertical axial planar foliation ($S_{3N}$; Fig. 5e). They are generally 1e10 m sized, mostly up-right or steeply-inclined, ESE-verging Metre-scale and gentle $F_{3N}$ folds and cm-dm sized $F_{3N}$ crenulations could be found as far to the south as in outcrop 4 (Fig. 9), but not in outcrops farther south.

Folds $F_{3N}$ refold $F_2$ folds and are refolded by $F_3S$ folds. Coaxial patterns of $F_{3N} \times F_2$ folding interference [type 3 of Ramsay (1967)] are noticed in outcrops 24 and 4 (Figs. 7 and 9). The similarly striking axial planes and axial planar foliations, and folds axes of $F_{3N}$ and $F_2$ (Fig. 5 b,c; e,f) show that $D_{3N}$ and $D_2$ regional shortenings were similarly-oriented WNWeESE, entirely confirming earlier predictions based on the detailed study of the Brasília dome. Examples of $F_{3N}$ refolded by $F_3S$ are common, too (Figs. 8 and 9). Foliation $S_{3N}$ is normally a 1 to 10 cm-spaced cleavage (locally marked by iron oxide) which cuts $S_0/S_1/S_2$ and displays varied degree of anastomose. The larger spacing is found in quartzite or arenaceous metarhythmite, or quartz-rich micaceous rocks of the nappes. In the outcrops to the west of Luziânia (Figs. 2 and 3), shortening $D_{3N}$ is almost absent, apart from three outcrops (14, 12, and 10) where foliation $S_{3N}$ occurs as a local feature shortened by $F_3S$ folds.

In outcrop 14, $S_{3N}$ is characterized by stylolite-type surfaces, anastomose, spaced w1 cm and perpendicular to bedding in a w1 m-thick packet of $F_3S$-folded quartzite layers. The surfaces of $S_{3N}$ generally do not cut across the schistose layers above or below (Fig. 10) indicating strain concentration and pressure solution due to $D_{3N}$ layer-parallel shortening but not strong enough for folding the layers in the southern part of the area. The $S_{3N}$ surfaces cut fold axis $B_3S$ perpendicularly and are $F_3S$-folded, similarly as in outcrop 10 (Fig. 11).

Thrust faults locally seen (outcrops 1 and 17) exhibit short traces in the road-cuts and their attitude is indicative of WNWeESE shortening.
Fig. 11. A 3-D diagram illustrating the main structural relationships found in the southern road-cut which composes outcrop 10 (802260 E; 8193549 N; Zone 22S). A guide layer (10 cm-thick) of quartzite (ornamented within a packet of phyllite) displays a strong and quite evident striation (Le) and is folded by F₃S folds with axis B₃S virtually parallel to Le. Foliation S₃N (a dm-spaced cleavage) is clearly folded by F₃S, too. Note the axis B₃S is sub-vertical or sub-horizontal, according to the dip of the folded surface.

### 4.5. Structures evidencing SW—NE (D₃S) shortening

Structures such as folds (F₃S) and their axial planar foliation (S₃S) constitute the main evidence of shortening in the SW—NE direction (D₃S). In outcrops closer to the Brasília dome (e.g., 1—4, 21 and 22) they are observable upon detailed observation, although in outcrop 21 they are large-scale and quite evident (Fig. 8).

Otherwise, F₃S folds dominate in outcrops away from the dome (e.g. 6—16 west of Luziânia; see Figs. 10 and 11). They are sized from few centimeters up to 100’s meters, trend NW—SE, are mostly upright to steeply-inclined, generally verging to NE (Fig. 5g—h), and normally display an axial planar foliation (S₃S) characterized by surfaces of 1—10 cm-spaced cleavage.

The ratio between the folds height and width (the aspect ratio by Twiss and Moores, 1992) is larger in outcrops 6—16 and the S₃S surfaces become much more evident there than elsewhere, indicating increasing intensity of shortening D₃S towards the southernmost part of the area.

Folds F₃S refold both D₂ and D₃N structures. Coaxial patterns of folding interference were noticed due to the oblique orientation of certain road-cuts
(e.g. in Fig. 9a—b) but may form where fold axes B2 and B3S are sub-parallel. Interference patterns of type 2 (Ramsay, 1967) are spatially reconstructed (Fig. 9c) and are expected to form due to the original orientation of $F_2$, $F_3N$ and $F_{3S}$ fold axes and respective axial surfaces.

Outcrops 2,3,10,11 and 13 fall in a slice of Araxá schist imbricate within Canastra rocks (Fig. 3) therefore indicating the existence of thrust faults. At three different parts of outcrop 2 (a ~ 50 x 50 m large exposure) the schist dips 22° to SW and contains large crystals of muscovite marked by two sets of striations oriented respectively ESE and ENE. The ESE-trending ones are taken as evidence of $D_1$—$D_2$ flow, the others as evidence of $D_{3S}$ flow.

5. Discussion

5.1. Structural evolution

The top-bottom stacking of Araxá schist, Canastra phyllite, and Paranoá (lower grade) metapelite, coupled with the metamorphic mineral assemblages related to $D_1$ and $D_2$; the spaced cleavage associated to $D_{3N}$ and $D_{3S}$; and the structural database in the previous sections, all permit understanding that: 1 - the rocks in the SW surroundings of the DF area underwent a first phase ($D_1$) of inter-and intra-layer slip still in their original sites of formation. M1 metamorphism varied from intermediary- to high-greenschist facies in Araxá rocks (up to garnet zone) to intermediary-greenschist facies in Canastra rocks, and sub- to low-greenschist facies in the Paranoá rocks; 2 - during the second phase of deformation, the nappes of Canastra and Araxá $D_1$-tectonites were successively pushed to ESE, onto autochthonous Paranoá $D_1$-tectonites; 3 - the highly anisotropic $D_1$—$D_2$ stack underwent shortening $D_{3N}$, so that $F_{3N}$ folds refolded $F_2$ co-axially; and 4 - the entire packet ($D_1$—$D_2$—$D_{3N}$ tectonites) was shortened in the SW-NE direction ($D_{3S}$), so that $F_{3S}$ folds trend perpendicularly to $F_{3N}$.

$D_1$ and $D_2$ structures demonstrate a regime of dip-slip shearing along surfaces practically coincident with shallowly-moderately dipping layers, a typical regime of WNW-dipping frontal ramps. $D_{3N}$ and $D_{3S}$ structures sliced the
entire crust (exhumation of granulites) and shortened thicker packets of upper crustal \(D_1-D_2\) tectonites.

The data confirm the Brasília dome as a \(D_3N\)-related structural window and also that late N—S shortening (\(D_3N_{TE}\)) vanished towards the south, across the dome. Actually, outcrops 17—20 and 24 (this study; Fig. 2) do not show evidence of any late shortening deforming \(F_3N\) folds. The succession of folding events (\(F_2, F_3N, F_3S\)) is unequivocal in this study area, and the attitudes of \(F_3N\) and \(F_3S\) parameters (Fig. 5) leave no doubt about the directions of \(D_3N\) and \(D_3S\) shortenings.

The preservation of the WNW—ESE direction of the stretching lineation (Lx) across the Brasília belt, even after orocline bending, was possible because shortening \(D_3S\) was perpendicular to the direction of ductile flow responsible for \(D_1-D_2\) tectonites, consequently the parallelism of fold axis B3S with Lx is common in the field (e.g. outcrops 10 and 14; Fig. 11). These outcrops demonstrate the same relationship (B3S/Lx) deduced previously, but on an indirect basis, around the Caldas Novas dome (Fig. 3), where the L2 stretching lineation remained constant (WNW—ESE) even after the dome’s uplift (D’el-Rey Silva et al., 2004).

5.2. **Crossed \(D_3S \times D_3N\) and the actors playing collision tectonics in the Tocantins province**

\(D_1-D_2\) data in the study area fit very well the direction of \(D_1-D_2\) flow reported elsewhere in the Brasília belt: in the External Zone and Brasília dome (D’el-Rey Silva et al., 2008, in preparation; Freitas-Silva and Campos, 1995); in Araxá rocks inside or close to the eastern part of the PZHS (Fig. 3; Araújo Filho, 2000); and in nappes surrounding the towns of Caldas Novas and Ipameri and further south in the Brasília belt (Figs. 3 and 4; Valeriano et al., 2004) or inside the Magmatic Arc (D’el-Rey Silva and Barros Neto, 2002; and other references therein).

Moreover, the continuity of \(D_3N\) shortening across the Brasília dome, plus the constant direction of the Lx stretching lineation on both sides of the PZHS, altogether shows a WNW—ESE path for collision that would have remained, if
another tectonic agent had not produced SW-NE shortening \( (D_3S) \) in the southern segment of the Brasília belt. The Paraná block is the unique agent capable to explain \( D_3S \) in the Amazon-São Francisco scenery of collision.

### 5.3. Relationships between shortenings \( D_3N \) and \( D_3S \), the Brasília dome, and the PZHS

For understanding that \( D_3S \times D_3N \) structures co-existing just SW of the DF area, the Brasília dome and the PZHS are closely related makes necessary to consider that the PZHS rotates: (A) imbricate slices of Anápolis-Itauçu granulites and Araxá schist existing along the Internal-External Zones border; (B) imbricate slices of Araxá, Canastra and Paranoá rocks to the NW, W and SW border of the DF area (Fig. 3; Araújo Filho, 2000; Moreira et al., 2008; this study); and (C) the granulite belt characteristic of the Internal Zone (Marini et al., 1984; Araújo Filho, 2000).

Also relevant are the ~ 645—635 Ma age of granulitization in rocks of the Anápolis-Itauçu and Uruaçu Complexes (see section 2) and the ~ 620 Ma age of Itapuranga syn-tectonic granite emplaced in the PZHS (U-Pb and geology data by Pimentel et al., 2003).

The above geological data imply: 1 - Canastra and Araxá \( D_1—D_2 \) tectonites already formed a continuous blanket of allochthonous rocks over \( D_1—D_2 \) autochthonous Paranoá tectonites in the areas of the future Internal and External zones, before exhumation of the belt of granulites, i.e. before \( D_3N \) and \( D_3S \) shortenings; and 2 - because the PZHS affect both the Rio Maranhão Thrust (RMT in Fig. 1b; D’el-Rey Silva et al., 2008) then the RMT must have evolved in the ~ 635—620 Ma age interval and the PZHS is most likely aged ~620 Ma.

### 5.4. Age intervals and the co-existence of shortenings \( D_3N \) and \( D_3S \)

The age intervals ~ 750—640 Ma (\( D_1—D_2 \)) and ~ 640—590 Ma (\( D_3N \)) are respectively supported by: U-Pb age data of ~ 750—640 Ma and ~ 635 Ma for diachronous amalgamation of the back-arc basins, north to south along the São Francisco paleocontinent margin; the ~ 645/635—620 Ma interval for granulite
formation-and-exhumation; the link between $D_3N$ regional shortening and the Amazon-São Francisco collision; and the $\sim 590-580$ Ma age of post-tectonic granites in the Brasília belt (Pimentel et al., 1999; Valeriano et al., 2004).

The age interval for deformation $D_3S$ ($\sim 630-590$ Ma) and the $D_3N \times D_3S$ co-existence are constrained by the following U-Pb age data and structural relationships: (1) emplacement of Araxá nappes $\sim 640-635$ Ma ago, in the southern segment of the Brasília belt (Valeriano et al., 2000, 2004); and (2) exhumation of granulites $\sim 630-620$ Ma ago in both the northern and southern segments of the belt, respectively due to $D_3N$ and $D_3S$, because (as already discussed) the PZHS rotates the whole granulite belt and encloses the $\sim 620$ Ma old Itapuranga granite.

Thus, although $D_3S$ overprints $D_3N$ structures, these two shortenings must have been coeval during granulite exhumation because the SW—NE direction of shortening is remarkably the same recorded by regional folds deforming the $D_1-D_2$ tectonites further south-southwest of the DF area, up to Caldas Novas and Ipameri (Fig. 3) and beyond, around Araxá and Capitólio (Fig. 4), and the structural data reported for these areas (D’el-Rey Silva et al., 2004; and references therein) indicate the inexistence of post-$D_1-D_2$ shortening in any other direction different of SW—NE ($D_3S$), in particular the WNW—ESE direction of $D_3N$.

Nevertheless, the EoPaleozoic events in the very southern end of the Brasília belt (Fig. 4) and the obduction of the Quatipuru ophiolite $\sim 540$ Ma ago (Paixão et al., 2008) justifies further discussing whether the lower age for $D_3S$ is actually $590-580$ Ma or $540-510$ Ma (section 5.5).

5.5. Tectonic model

A feasible tectonic model for the Brasília belt and Tocantins Province should consider that the structural evolution of the Brasília dome implies the PZHS developed in an area already shortened by $D_3N$. Moreover, the geology data in the Tocantins Province fit in the expected consequences of the collision of irregular continental margins (see also D’el-Rey Silva et al., 2011) and the $D_1-D_2$ and $D_3N$ structural data here summarized indicate the Amazon and São
Francisco always along a WNW—ESE-trending route of collision (Fig. 12 a—b).

As explained ahead, the crystallization age of ~780—750 Ma for the Quatipuru ophiolite and its obduction ~540 Ma ago onto a carbonate platform (Paixão et al., 2008) make necessary envisaging the very beginning of A-subduction of the Amazon paleo- continent, by the time the magmatic arc was pushed against São Francisco paleocontinent margin, leading to the blanket of nappes on the sites of the future Internal and External Zones, until the area of the future Distrito Federal (Fig. 12b). Furthermore, if nappes propagated ~640—635 Ma ago onto the southern paleomargin (Valeriano et al., 2004) the ocean remained opened south of the locking point 1 (Fig. 12b) strongly suggesting control by the irregular margin of the São Francisco paleocontinent about the Distrito Federal latitude (Pimentel et al., 1996b).

A diachronous opening of the Quatipuru ocean towards the south is physically required, since ~750 Ma ago, should the ocean to the south of locking point 1 continues to close (Fig. 12c). This means: despite the Araguaia belt consists of ~630 Ma old tectonites the Estrondo Group (Paixão et al., 2008) should not be a surprise if metamorphism as old as ~750 Ma come to be futurely known in rocks of that belt.

Should the ocean close entirely and collision continues to the south, the Quatipuru ocean spreading must reach locking point 1. The Amazon paleocontinent situated west of the Quatipuru ocean is then pulled farther into subduction, moving to point 2 for pushing younger Di—D₂ nappes onto the southern paleomargin, coevally with granulite facies metamorphism of supracrustal rocks carried to depth into the subduction zone.
High P-T metamorphism soon before 635 Ma ago, under the São Francisco paleocontinental margin, and the fact that the age interval for deformation D₃S in the Brasília belt (~ 630—590 Ma) overlaps the age interval for sedimentary-volcanic infilling of the Paraguay basin (~ 620—550 Ma; Pimentel et al., 1996a) plus the EoPaleozoic emplacement of nappes of high-PT basement rocks onto the southernmost paleocontinental margin (Fig. 4) coevally with both the obduction of the Quatipuru ophiolite and deformation-metamorphism (~ 550—510 Ma; Valeriano et al., 2004) in the Paraguay belt, altogether implies: A - the evolution of the Paraguay belt, Paraná block and the southern segment of the Brasília belt must have been strongly linked; and B - the southeastern part of the Amazon paleocontinent destined to become the Paraná block was very close of the São Francisco margin, or such block moved very fast into collision, most likely both.
Locking point 2 (Fig. 12c) demanded rifting (Paraguay) controlled by an anti-clockwise rotation in the southeastern part of the Amazon paleocontinent, whereas (to the N) the recently-formed Araguaia block (Fig. 12d) was pushed further into A-subduction, promoting $D_3N$ shortening of the northern São Francisco paleomargin. Shortening $D_3S$ was brought to the tectonic scenery since about ~635 Ma ago (Fig. 12e) when the Paraná block drifted-apart and followed along a SW—NE path for A-subduction beneath the São Francisco paleomargin.

5.6. Further discussing the timing for deformation $D_3S$

Despite of the several structural and geological reasons for shortenings $D_3N$ and $D_3S$ co-exist in the ~635—590/580 Ma age interval (previous section) whether the lower limit of age for shortening $D_3S$ is ~590 or ~540 Ma still demands a more detailed analysis.

In fact, the lower limit becomes ~540 Ma if new geochronological data come to show this as the best age for post-tectonic granites already dated, or for post-tectonic granites not described yet across the Brasília belt. If so, will be not a surprise if ages futurely obtained directly for the deformation ($39Ar/40Ar$ techniques, for example) place in the EoPaleozoic the minimal age for $D_3S$ tectonites. Nevertheless, from the tectonic point of view should not be a surprise if such ages remain >590/580 Ma, either.

Actually, the event of emplacement of the EoPaleozoic nappes to ENE onto the craton margin, although most likely related to further A-subduction of the Paraná block under the São Francisco margin and due to the Amazon-Paraná collision responsible for the Paraguay belt, did not necessarily deform the upper crust further north in the Brasília belt.

This is because, the residual effects of the EoPaleozoic shortening, if any existed there, could have been concentrated mostly in the lower crust of the (overriding) São Francisco plate, facilitated by the higher degree of plasticity of the rocks there, coupled with the difficult for deforming a mechanically stronger upper crust consisting of tectonites with multiply-oriented anisotropies such as: NE-trending axes of $F_2$ folds; ESE-trending axes of $F_2$ sheaths; and NNW-
trending axes of $F_{3S}$ folds. Moreover, additional A-subduction of the Paraná block could have been accommodated by a ramp of underthrusting inside the block itself and far west of the subduction zone, without deforming the rocks in the Brasília orogen, similarly to what has been recently shown in the Indian continent still A-subducting under Asia (Yin et al., 2010).

5.7. Oroclinal bending, PZHS, Brasília dome and box-junction zone

Among the large variety of mechanisms capable of producing an orocline (Sussman and Weil, 2004) there are four which could be thought as applicable to the Brasília belt. The first three are: rotation on the slip direction around a continent corner; rotation of the grain due to collision of an indented margin; and larger magnitude of rotation closer to a major shear zone. The first two fail to explain the same direction of $D_1-D_2-D_3N$ ductile flow on both sides of the PZHS and the delay in the onset of $D_3S$ relative to $D_3N$. The third fails to explain (primarily) why the rotation associated to the PZHS solely affects the Internal and External zones, not the magmatic arc.

The fourth mechanism consists of perpendicular shortenings, such as $D_{3S}$ and $D_{3N}$. Moreover, the tectonic model (Fig. 12) highlights the role of irregular continental margins in the generation of a third continental block capable to produce coeval crossed shortenings in the Brasília belt, emphasizing that such margins may strongly control the geometry-evolution of collision zones in time (such as in Coward and Dietrich, 1989). More recently, Ghiglione and Cristalline (2007) used geophysical and structural evidence supporting the Patagonian orocline as formed in consequence of rotation of a small block. Contrarily to the Paraná-São Francisco scenery of collision, however, these authors reported two differently oriented deformation phases attributed to the small block moving in different directions, at different times, all consistent with a change in the direction of convergence of Farallon-Nazca plate, ca. 27 Ma ago.

Three factors justifies the bending and the PZHS along the interface between the Barro Alto and Anápolis-Itauçu complexes (Fig. 3): (1) - the different rock compositions in these complexes likely implied a zone of crustal weakness; (2) - the Box-Junction Zone and its continuation to the west
represented a zone of relative enhancement in crustal strength, not suitable for bending; and (3) the uniqueness of shortening $D_3N$ or $D_3S$ implied lack of reasons for bending at any other site respectively north of the PZHS or south of the Box-Junction Zone.

Sub-horizontal normal and sub-horizontal shear components of $D_3N$ and $D_3S$ stresses (relative to the PZHS) explain the strong N–S shortening of $D_1$–$D_2$ nappes nearby or inside the eastern PZHS (see spoons by Araújo Filho, 2000; Fig. 3) and rotation of the litho-structural grain along its northern and southern sides. The normal components account also for a non penetrative spaced cleavage ($S_3$ late foliation; stereogram in Fig. 5i) found locally in this study area. D’el-Rey Silva et al. (in preparation) emphasize that both the N–S shortening across and the rotation along the PZHS’ eastern half fit the expected shrinking in the inner arc of the Brasília bend, whereas the series of WNW-trending dykes which characterizes the western half (through the magmatic arc; Moreira et al., 2008) record the extension expected for the outer arc.

During a short time interval (15 million years) the granulitic rocks exhumed from a depth of ~35 km along the Rio Maranhão Thrust (RMT), cut through the blanket of nappes and reached the crustal level currently exposed by the same time of oroclinal bending and formation of the PZHS, ~ 625—620 Ma ago, according to abundant structural and mapping data, plus data on Sm-Nd isotopes relative to metasediments in the hangingwall and footwall of the RMT (area J; Fig. 1b; D’el-Rey Silva et al., 2008). All this means oroclinal bending assisted by a suitably elevated temperature and matches the concept of hot orogen recently postulated for the Brasília belt at that time (Della Giustina, 2010).

The Brasília dome and the $D_3N$ structures must have evolved (Fig. 12d) lately in the ~ 630—625 Ma age interval, coevally with granulite exhumation but before the PZHS’ onset, whereas the Box-Junction Zone (Fig. 2) evolved in the 625—620 Ma age interval, shortly after the dome’s uplift and shortly before orocline bending and the PZHS (Fig. 12e).
6. Conclusions

The Brasília belt is an orocline comprised of N—S longitudinal domains (magmatic arc; Internal and External zones) and divided into a NE-trending northern segment and a SE-trending southern segment. It forms, together with the Araguaia and Paraguay belts, the Tocantins province. This province results from Neoproterozoic-Cambrian collision events involving the Amazon and São Francisco paleocontinents and the Paraná block of central Brazil.

Within the Distrito Federal (DF) area, the Brasilia dome is a structural window demonstrating stratigraphy and structural (D₁—D₂—D₃N events of WNW—ESE shortening) continuity across the External Zone of the Brasília belt. The eastern tip of the Pirineus Zone of High Strain (PZHS) rests few kilometers west of the dome. Abundant structural data have shown deformation of the northern segment due to WNW—ESE shortenings D₁—D₂ and D₃N, and the southern segment due to WNW—ESE D₁—D₂, overprinted by shortening SW—NE (D₃S). Moreover, in the area close to the SW margin of the dome, D₃S structures shorten D₃N structures in a narrow Box-Junction Zone.

The large database allows understanding the Tocantins Province in terms of the collision of irregular continental margins. The Amazon-São Francisco collision produced the Brasilia-Araguaia belt, and resulted in D₁—D₂ and D₃N WNW—ESE shortenings (respectively ~ 750—640 Ma and ~ 640—590 Ma ago) in the Brasilia belt until the southernmost latitude of the Box-junction Zone. The Paraná block rifted/drifted-apart from the Amazon paleocontinent and shortened SW-NE the southern segment of the belt ~ 630—590/580 Ma ago, as indicated by D₃S structures up to the northernmost latitude of the Box-Junction Zone. Shortening D₃ (D₃N + D₃S) accounts for exhumation of granulites in the Brasília belt, for orocinal bending and the PZHS. The Amazon-Paraná collision accounts for both the Paraguay belt and Cambrian nappes in the southernmost Brasília belt.

Acknowledgements

This paper is part of the project Neoproterozoic Tectonic Evolution of the São Francisco Craton and Marginal Belts, coordinated by L.J.H. D’el Rey Silva

References


