



**Universidade de Brasília
Instituto de Geociências
Programa de Pós-graduação em Geologia**

**EVOLUÇÃO TECTÔNICA DAS BACIAS ARAÍ, TRAÍRAS E PARANOÁ
NA FAIXA BRASÍLIA NORTE**

TESE DE DOUTORADO

Nº 139

MARCO ANTONIO CAÇADOR MARTINS-FERREIRA

Orientador: Prof. Dr. JOSÉ ELOI GUIMARÃES CAMPOS

Co-orientador: Prof. Dr. FARID CHEMALE JR

Brasília, 15 de setembro de 2017



**Universidade de Brasília
Instituto de Geociências
Programa de Pós-graduação em Geologia**

EVOLUÇÃO TECTÔNICA DAS BACIAS ARAÍ, TRAIÍRAS E PARANOÁ NA FAIXA BRASÍLIA NORTE

TESE DE DOUTORADO

Nº 139

MARCO ANTONIO CAÇADOR MARTINS-FERREIRA

Banca examinadora:

Prof. Dr. José Eloi Guimarães Campos (Orientador)

Prof. Dr. Carlos José Souza de Alvarenga

Prof. Dr. Fernando Flecha Alkmim

Prof. Dr. Claudio de Morrison Valeriano

Brasília, 15 de Setembro de 2017

"It is the little causes, long continued, which are considered as bringing about the greatest changes of the earth."

James Hutton

AGRADECIMENTOS

Meus sinceros agradecimentos ao Professor Dr. José Eloi Guimarães Campos, pela excelente orientação e ao Professor Dr. Farid Chemale Jr. pela co-orientação.

Aos professores Dr. Fernando Flecha Alkmim, Dr. Claudio de Morrison Valeriano e Dr. Carlos José Souza de Alvarenga, participantes da banca examinadora, pelas coerentes sugestões de correções e aperfeiçoamento da presente tese.

À Universidade de Brasília e ao Instituto de Geociências pelos meios físicos e técnicos disponibilizados para a realização da presente pesquisa e dissertação e a todos os geólogos e professores que tanto me ensinaram e continuam ensinando.

Aos professores Dr. José Oswaldo de Araújo Filho e Dr. Carlos Emanuel de Souza Cruz pela participação da comissão de acompanhamento e pelas discussões.

Ao professor Prof. Dr. Luiz José Homem D'el-Rey Silva pelos ensinamentos em geologia estrutural.

Às professoras Dra. Catarina Labouré Benfica Toledo e Dra. Adalene Moreira Silva, por todo o apoio durante a elaboração da presente tese.

À professora Dra. Mônica Giannoccaro Von Huelsen pelas discussões e contribuições.

Aos alunos de graduação Amanda Mayra Maia de Freitas, Lucas Pozzobon De Bem, Heloisa Carolina de Moraes da Silva, Pedro Henrique Caldas Kimura, Marcelo Pellicano e Thiago Lima pelo acompanhamento nos trabalhos de campo; Brandow Lee Neri e Aline Lacerda pela ajuda com o uso do aplicativo Geosoft.

Agradeço ainda a todos os amigos e familiares que sempre deram suporte.

E finalmente à Lívia Helena e à Nina que foram a fonte de todo o amor, força e determinação, fundamentais para o bom desenvolvimento deste trabalho.

SUMÁRIO

LISTA DE FIGURAS	VIII
LISTA DE TABELAS	XIII
RESUMO.....	XIV
ABSTRACT	XVI
CAPÍTULO 1	1
INTRODUÇÃO	1
1.1 APRESENTAÇÃO.....	1
1.2 JUSTIFICATIVA	3
1.3 OBJETIVOS.....	5
1.4 LOCALIZAÇÃO.....	5
1.5 ESTRUTURAÇÃO DA TESE.....	7
1.6 MATERIAIS E MÉTODOS	8
<i>AMOSTRAGEM</i>	<i>8</i>
<i>Separação mineral e montagem das amostras</i>	<i>9</i>
1.7 GRAVIMETRIA	10
<i>REFERÊNCIAS BIBLIOGRÁFICAS</i>	<i>11</i>
CAPÍTULO 2	15
ARTIGO 1:.....	15
COMPARTIMENTAÇÃO GEOMORFOLÓGICA COMO SUPORTE PARA ESTUDOS DE EVOLUÇÃO GEOTECTÔNICA: APLICAÇÃO À REGIÃO DA CHAPADA DOS VEADEIROS, GO	15
ABSTRACT	16
RESUMO	17
1. INTRODUÇÃO	17
2. ÁREA DE ESTUDO.....	20
3. CONTEXTO GEOLÓGICO REGIONAL.....	22
4. MATERIAIS E MÉTODOS	23
5. RESULTADOS.....	28
5.1. <i>Interpretação dos Lineamentos Morfoestruturais.....</i>	<i>28</i>
5.2. <i>Compartimentação Geomorfológica e Geotectônica</i>	<i>30</i>
5.3. <i>Compartimentos Geomorfológicos.....</i>	<i>31</i>
6. DISCUSSÕES.....	35
6.1. <i>Análise do Mapa de Lineamentos.....</i>	<i>35</i>
6.2. <i>Análise do Mapa de Compartimentos Geomorfológicos</i>	<i>36</i>

7. CONCLUSÕES	39
<i>Agradecimentos</i>	40
<i>Referências Bibliográficas</i>	40
CAPÍTULO 3	43
ARTIGO 2:.....	43
PROTEROZOIC INTRACONTINENTAL BASIN SUCCESSION IN THE SÃO FRANCISCO CRATON WESTERN MARGIN CONSTRAINED BY DETRITAL ZIRCON GEOCHRONOLOGY	43
ABSTRACT	44
1. INTRODUCTION	45
2. GEOLOGICAL SETTING AND SAMPLING.....	46
2.1. <i>Sampling</i>	49
3. MATERIAL AND METHODS.....	52
3.1. <i>Sample Preparation</i>	52
3.2. <i>U-Pb method</i>	52
3.3. <i>Lu-Hf method</i>	53
4. RESULTS AND DISCUSSION.....	53
5. CONCLUSIONS.....	72
<i>Acknowledgements</i>	73
<i>References</i>	73
CAPÍTULO 4	80
ARTIGO 3:.....	80
PALEORIFT STRUCTURE CONSTRAINED BY GRAVITY AND STRATIGRAPHIC DATA, THE STATHERIAN ARAÍ RIFT CASE.....	80
ABSTRACT	81
1. INTRODUCTION	81
2. GEOLOGICAL SETTING.....	84
2.1. <i>Rhyacian Crystalline Basement</i>	87
2.2. <i>Statherian Rift and Calymmian Sag Sequences</i>	87
The Araí Group: Pre-rift and Rift Sequences	88
The Trairas Group: Sag Sequence.....	89
2.3. <i>Stenian to Tonian Intracontinental Sequences</i>	90
2.4. <i>Cryogenian to Ediacaran Intracontinental Sequences</i>	91
3. PREVIOUS GEOPHYSICAL STUDIES	91
4. GRAVIMETRIC DATA ACQUISITION, PREPARATION AND PROCESSING.....	92
5. RESULTS AND DISCUSSION.....	95

5.1. Gravimetric Support to Rift-Related Research	95
5.2. Power Spectrum Results	97
5.3. Euler Deconvolution Results.....	98
5.4. Euler Solutions Interpretation.....	100
5.5. Description of Main Araí Paleorift Structures	103
5.6. Araí Paleorift Tectonic Model	107
5.7. Origins of Anorogenic Magmatism.....	107
5.8. Present-day Local Topography	110
5.9. Global Correlation to Other Statherian Rifts.....	111
5.10. Regional Correlation in the São Francisco Craton.....	112
6. FINAL DISCUSSION AND CONCLUSIONS	117
Acknowledgements.....	118
References	118
CAPÍTULO 5	126
ARTIGO 4:.....	126
PARANOÁ BASIN EVOLUTION CONSTRAINED BY STRATIGRAPHIC AND GRAVIMETRIC DATA AND ITS CORRELATION TO REGIONAL BASINS.....	126
ABSTRACT	127
1. INTRODUCTION	128
2. GEOLOGICAL SETTING.....	130
2.1. Siderian/Rhyacian Crystalline	131
2.2. Statherian	131
2.3. Calymmian.....	131
2.4. Stenian/Tonian	132
2.5. Ediacaran.....	132
2.6. Paranoá Group Geology	132
3. GRAVIMETRIC DATA ACQUISITION, PREPARATION, INVERSION AND MODELING.....	144
4. RESULTS AND DISCUSSION.....	146
4.1. Results Discussion According to Literature Review.....	150
4.2. Regional Correlation	160
4.3. Global Correlation to Supercontinent Cycles	167
5. CONCLUSIONS.....	169
Acknowledgements.....	170
References	170
CAPÍTULO 6	181

ARTIGO 5:	181
CARACTERIZAÇÃO ESTRUTURAL PRELIMINAR DA SALIÊNCIA DO MOQUÉM, SISTEMA DE DOBRAMENTOS E CAVALGAMENTOS DA ZONA EXTERNA DA FAIXA BRASÍLIA NORTE	181
ABSTRACT	182
1. INTRODUÇÃO	182
2. MÉTODOS	184
3. RESULTADOS.....	184
4. DISCUSSÃO	192
4. CONCLUSÕES	194
<i>Agradecimentos</i>	194
<i>Referências Bibliográficas</i>	195
CAPÍTULO 7	197
DISCUSSÕES FINAIS E CONCLUSÕES	197
7.1. EVOLUÇÃO TECTÔNICA DAS BACIAS ARAÍ E TRAÍRAS	198
7.2. A BORDA OESTE DO PALEOCONTINENTE SÃO FRANCISCO	199
7.3. EVOLUÇÃO TECTÔNICA DA BACIA PARANOÁ	200
7.4. INVERSÃO BRASILIANA	202

LISTA DE FIGURAS

CAPÍTULO 1

FIGURA 1 - MAPA DE LOCALIZAÇÃO DA ÁREA DE DETALHAMENTO DO ESTUDO NA FAIXA BRASÍLIA. FB=FAIXA BRASÍLIA. OS DOMÍNIOS NORTE E SUL (FB NORTE E FB SUL) SÃO DELIMITADOS APROXIMADAMENTE PELO LINEAMENTO DOS PIRENEUS À ALTURA DE BRASÍLIA.	6
FIGURA 2 - MAPA DE LOCALIZAÇÃO DA ÁREA DE DETALHAMENTO DO ESTUDO SOBRE IMAGEM SRTM COLORIDA COM MAPA DE POSICIONAMENTO NO ESTADO DE GOIÁS E ARTICULAÇÃO DAS FOLHAS CARTOGRÁFICAS QUE ABRANGE.....	7
FIGURA 3 - DIAGRAMA ESQUEMÁTICO DA ESTRUTURAÇÃO DA PRESENTE TESE.	8
FIGURA 4 - MAPA DA MALHA DE DADOS GRAVIMÉTRICOS DE DETALHE UTILIZADOS. EM MAPA COM LIMITES ESTADUAIS ESTÃO PLOTADAS AS ESTAÇÕES GRAVIMÉTRICAS LEVANTADAS PELO IBGE E A LOCALIZAÇÃO DOS 4 PERFIS DE ULTRA DETALHE ADQUIRIDOS ESPECIFICAMENTE PARA ESTE TRABALHO.	11

CAPÍTULO 2

FIGURA 1 - LOCALIZAÇÃO DA ÁREA NO ESTADO DE GOIÁS, ARTICULAÇÃO DAS FOLHAS CARTOGRÁFICAS E PRINCIPAIS MUNICÍPIOS SOBRE MAPA HIPSOMÉTRICO DA ÁREA DE ESTUDO. EXEMPLO DE MAPA SECUNDÁRIO OU DERIVADO, APLICADO À COMPARTIMENTAÇÃO GEOMORFOLÓGICA.....	21
FIGURA 2 - LOCALIZAÇÃO DA ÁREA DE ESTUDO NO MAPA GEOLÓGICO DA FAIXA DE DOBRAMENTOS E CAVALGAMENTOS BRASÍLIA.....	23
FIGURA 3 - MAPAS (A) HIPSOMÉTRICO, (B) DE REDE DE DRENAGEM, (C) PEDOLÓGICO E (D) GEOLÓGICO DA ÁREA (FONTES: USGS, IBGE, EMBRAPA, CPRM, RESPECTIVAMENTE). MAPAS TEMÁTICOS UTILIZADOS COMO CRITÉRIOS PARA A COMPARTIMENTAÇÃO GEOMORFOLÓGICA.....	26

FIGURA 4 – MAPA DE LINEAMENTOS MORFOESTRUTURAIS SOBRE SRTM 1-ARCO EM ESCALA DE CINZA SOMBREADO.	27
FIGURA 5 – MAPA DAS PRINCIPAIS FEIÇÕES DE RELEVO DA ÁREA DE ESTUDO SOBRE PRODUTO SRTM SOMBREADO.	28
FIGURA 6 – MAPA DE COMPARTIMENTOS GEOMORFOLÓGICOS.	34
CAPÍTULO 3	
FIGURE 1 – GEOLOGIC MAP SHOWING LOCATION OF MAIN INTRACONTINENTAL BASINS IN THE SÃO FRANCISCO CRATON INTERIOR AND MARGINS. THE BLACK LINES INDICATE THE MOST ACCEPTED EXTENT OF THE CRATON AFTER CA. 540 MA AND EXTENT OF THE CONTINENTAL PALEOCONTINENT AT CA. 1.8 GA. DOTTED BLUE LINE INDICATES THE LIMITS OF THE GOIÁS MASSIF ACCORDING TO FUCK ET AL. (1994). I) NORTHERN BRASÍLIA BELT (VEADEIROS SUPERGROUP); II) NORTHERN ESPINHAÇO RANGE; III) DIAMANTINA PLATEAU; (CHAPADA DIAMANTINA); IV) SOUTHERN BRASÍLIA BELT AND V) SOUTHERN ESPINHAÇO RANGE (MODIFIED FROM GUADAGNIN AND CHEMALE, 2015).	48
FIGURE 2 – GEOLOGIC MAP OF THE STUDY AREA. SAMPLING AREA (DETAILED IN FIGURE 3) IS OUTLINED IN BLACK RECTANGLE. THE GOIÁS MASSIF - CAMPINORTE DOMAIN (AFTER CORDEIRO AND OLIVEIRA 2017) IS OUTLINED IN BLACK DASHED LINES (MODIFIED FROM CPRM 2008).	49
FIGURE 3 – DETAIL GEOLOGIC MAP WITH SAMPLE SITES AND NUMBERS INDICATED. GROUP, FORMATION AND MEMBER NAMES ARE ADOPTED FROM TANIZAKI ET AL. (2015) FOR THE ARAÍ GROUP AND CAMPOS ET AL. (2013) FOR THE PARANOÁ GROUP. SEE MAP LOCATION ON FIGURE 2.	51
FIGURE 4 – HISTOGRAM AND RELATIVE PROBABILITY PLOTS FOR SAMPLES TFC 01 TO TFC 12.	65
FIGURE 5 – NORMALIZED AGE PROBABILITY PLOT FOR ALL U-Pb DATA GROUPED BY FORMATION. MAIN AGE INTERVALS ARE IDENTIFIED AS: CA = CALYMMIAN; ST = STATHERIAN, RH = RHYACIAN, NA = NEOARCHEAN, MA = MESOARCHEAN; PA = PALEOARCHEAN. ...	67
FIGURE 6 – U-Pb AGES VERSUS eHf(T) RESULTS PLOT OF DETRITAL ZIRCONS FOR ALL SAMPLES ANALYZED, DISTINGUISHED BY GROUP AND FORMATION. CA = CALYMMIAN; ST = STATHERIAN, RH = RHYACIAN, NA = NEOARCHEAN, MA = MESOARCHEAN; PA = PALEOARCHEAN.	67
FIGURE 7 – CONCORDIA DIAGRAM FOR THE YOUNGEST ZIRCON GRAINS FROM TRAÍRAS FORMATION, ARAÍ GROUP.	68
FIGURE 8 – TECTONO-STRATIGRAPHIC CHART ARAÍ AND TRAÍRAS GROUPS. REGIONAL CORRELATION CHART OF THE PROPOSED VEADEIROS SUPERGROUP WITH THE ESPINHAÇO SUPERGROUP FROM STATHERIAN TO TONIAN PERIODS, SHOWING BASIN TYPE, DEPOSITIONAL ENVIRONMENTS, UNCONFORMITIES, FORMAL LITHOSTRATIGRAPHIC HIERARCHICAL UNITS, FIRST-ORDER SEQUENCES AND MAIN GLOBAL TECTONIC EVENTS (MODIFIED FROM TANIZAKI ET AL. 2015).	69
FIGURE 10 – GEOLOGIC MAP (MODIFIED FROM FIGURE 1) INDICATING A MORE ACCURATE SÃO FRANCISCO PALEOCONTINENT WESTERN LIMIT AT 1.8 GA (RED LINE). THE BLACK LINES INDICATE THE MOST ACCEPTED EXTENT OF THE CRATON AFTER CA. 540 MA AND THE PREVIOUS PALEOCONTINENT LIMIT AT CA. 1.8 GA. DOTTED BLUE LINE INDICATES THE LIMITS OF THE GOIÁS MASSIF ACCORDING TO FUCK ET AL. (1994). I) NORTHERN BRASÍLIA BELT (VEADEIROS SUPERGROUP); II) NORTHERN ESPINHAÇO RANGE; III) DIAMANTINA PLATEAU; (CHAPADA DIAMANTINA); IV) SOUTHERN BRASÍLIA BELT AND V) SOUTHERN ESPINHAÇO RANGE (MODIFIED FROM GUADAGNIN AND CHEMALE, 2015).	72
CAPÍTULO 4	
FIGURE 1 - GEOLOGIC MAP SHOWING THE DISTRIBUTION OF THE PALEO-MESOPROTEROZOIC RIFT-SAG SEQUENCES IN THE SÃO FRANCISCO CRATON INTERIOR AND MARGINS (ORANGE COLOR) FORMED BY THE ESPINHAÇO AND VEADEIROS SUPERGROUPS. THE BLACK LINES INDICATE THE MOST ACCEPTED EXTENT OF THE CRATON AFTER CA. 540 MA AND THE PROBABLE EXTENT OF THE PALEOCONTINENT AT CA. 1.8 GA AS PROPOSED BY MARTINS FERREIRA ET AL. (2017). I) NORTHERN BRASÍLIA BELT; II) SOUTHERN BRASÍLIA BELT; III) NORTHERN ESPINHAÇO RANGE; IV) SOUTHERN ESPINHAÇO RANGE AND V) DIAMANTINA PLATEAU (CHAPADA DIAMANTINA); (MODIFIED FROM GUADAGNIN AND CHEMALE, 2015).	85

FIGURE 2 - STUDY AREA GEOLOGICAL MAP SHOWING THE MAIN CITIES. (MODIFIED FROM CUADROS, 2017, THAT PRODUCED THIS MAP WITH GEOLOGIC DATA COMPILED FROM GEOLOGICAL MAPS SC.23, SD.22, SD.23, SC.22-Z-D, SD.22-X-D, SD.23-V-A, SC.23-Y-C, SD.23-V-C-III, SD.23-V-C-V AND SD.23-V-C-VI BY THE BRAZILIAN GEOLOGICAL SURVEY (CPRM; [HTTP://GEOBANK.CPRM.GOV.BR/](http://geobank.cprm.gov.br/)). MODIFICATIONS MADE TO THE MAP IN THIS STUDY WERE FOCUSED ON THE DIVISION OF PREVIOUSLY UNDIVIDED GEOLOGIC UNITS, INCLUDING THE ANOROGENIC GRANITES (PEDRA BRANCA AND SERRA DA MESA SUITES) AND THE SUPRACRUSTAL RIFT AND SAG SEQUENCES, RESPECTIVELY THE ARAÍ AND TRAÍRAS GROUPS (AFTER MARTINS-FERREIRA ET AL. 2017) USING OUR OWN GEOLOGIC MAPPING DATA AND THE DATA COMPILED FROM THE FINAL GRADUATION DETAIL GEOLOGICAL MAPPING AT UNIVERSITY OF BRASÍLIA IN THE YEARS 1978, 1984, 1998, 1999, 2001, 2005, 2009 AND 2010.. 86

FIGURE 3 - ARAÍ AND TRAÍRAS GROUPS TECTONO-STRATIGRAPHIC CHART WITH AVAILABLE GEOCHRONOLOGICAL DATA AND INTERPRETED BASIN TECTONIC EVENTS AND SETTINGS (MODIFIED FROM TANIZAKI ET AL. 2015)..... 90

FIGURE 4 - 2D CRUST AND MANTLE DENSITY MODEL UNDER THE PORANGATU AND CAVALCANTE DEEP SEISMIC REFRACTION LINES (SOARES ET AL. 2006). THE MODEL WAS OBTAINED BY RAY-TRACING AND RECEIVER FUNCTION MODELING FROM THE CAVALCANTE SEISMIC STATION. THIS STUDY AREA IS OUTLINED IN GREEN, THE CRUST-MANTLE INTERFACE IS OUTLINED IN RED AND THE LOW-DENSITY BODY INTERPRETED BY THE AUTHORS IS MARKED IN BLUE. 92

FIGURE 5 – BOUGUER ANOMALY MAPS PRODUCED WITH THE IBGE DATASET WITH GRAVITY STATIONS SUPERPOSED. DATA WAS GRIDDED A) WITHOUT BLANKING DISTANCE AND B) BLANKING DISTANCE WAS USED WITH 150 M CELL SPACING. THE STATIONS SURVEYED BY THE AUTHORS ARE OUTLINED IN GREY..... 93

FIGURE 6 - DATA PROCESSING FLUXOGRAM. THDR = TOTAL HORIZONTAL GRADIENT; ASI OR TA = ANALYTIC SIGNAL INCLINATION (TILT); BGF = BOUGUER GRAVITY FIELDGRID. 94

FIGURE 7 - RADIALLY AVERAGED POWER SPECTRUM DIAGRAMS FOR A) BOUGUER AND B) VERTICAL DERIVATIVE. 95

FIGURE 8 – FREQUENCY HISTOGRAMS FOR THE EULER DECONVOLUTION SOLUTIONS DATASETS. I = INDEX; T=TIME; W=WINDOW. X AXIS REPRESENTS DEPTH IN KM AND Y AXIS REPRESENTS THE FREQUENCY OF SOLUTIONS PER DEPTH IN NUMBER AND PERCENTAGE..... 98

FIGURE 9 - SELECTED FREQUENCY HISTOGRAMS. COLORED BOXES REPRESENT THE MAIN DEPTH INTERVALS. HORIZONTAL AXIS REPRESENTS DEPTH IN KM. VERTICAL AXIS REPRESENTS THE FREQUENCY OF SOLUTIONS PER DEPTH IN NUMBER AND PERCENTAGE. 99

FIGURE 10 - INTERPRETED EULER FAULT DEPTH INTERVALS A) FREQUENCY HISTOGRAM SHOWING FREQUENCY CURVE AND B) SAME HISTOGRAM PLOTTED OVER SIBSON’S CONCEPTUAL MODEL OF A MAJOR FAULT ZONE (SIBSON 1977). 100

FIGURE 11 - COMBINED EULER SOLUTIONS A) OVER THE BOUGUER MAP AND B) UNDER GEOLOGIC CONTACTS (CPRM 2008). 100

FIGURE 12 - A) EULER LINEAMENT EXTRACTION (RED LINES) PLOTTED OVER EULER SOLUTIONS; B) GEOLOGIC MAP WITH EULER LINEAMENTS (RED LINES) AND THE RESULTING ADJUSTED AND INTERPRETED RIFT STRUCTURES (SOLID BLACK LINES INDICATE INTERPRETED MAIN RIFT FAULTS AND DASHED LINES INDICATE INTERPRETED RIFT TRANSFER FAULTS)..... 102

FIGURE 13 - DERIVATIVES FROM THE BOUGUER IMAGE. A) TOTAL HORIZONTAL GRADIENT - THDR; B) ANALYTIC SIGNAL INCLINATION - TILT FROM BOUGUER IMAGE. 103

FIGURE 14 – GEOLOGIC MAP SHOWING INTERPRETED RESULTS OF MAIN RIFT FAULTS AND TRANSFER FAULTS. I = NOVA ROMA GRABEN; II = ARRAIAS-CAVALCANTE GRABEN; III = BEZERRA HORST; IV = PRATA GRABEN; V = PARANÁ HORST; VI = PARANÁ GRABEN; AM = ÁGUA MORNA FAULT; GA = GOIANA-ABOBOREIRA FAULT. THE LOCATION OF A-A' PROFILE IS INDICATED. 104

FIGURE 15 – A) GRAVITY RESIDUAL MAP FOR UPWARD CONTINUATION AT 1000 M ELEVATION INDICATING THE OUTCROP LOCATION OF ANOROGENIC GRANITES FROM PEDRA BRANCA SUITE AND B) VOXEL 3D FOR THE SAME AREA SHOWING MAXIMUM DEPTH OF ANOMALIES. 110

FIGURE 16 - ANALOGY BETWEEN THE (A) EAST AFRICA RIFT SYSTEM (ADAPTED FROM FONTIJN ET AL. 2010) AND (B) THE SÃO FRANCISCO CRATON PALEOPROTEROZOIC RIFT SYSTEM: Ar) ARAÍ RIFT; Pi) PIRAPORA AULACOGEN; PA) PARAMIRIM AULACOGEM; LV) LOWER

VEADEIROS AND LE) LOWER ESPINHAÇO (ADAPTED FROM REIS ET AL. 2017) THE RED LINE MARKS THE LOCATION OF THE SEIMIC SECTION SHOWN IN FIGURE 19A, THAT CROSSES THE PIRAPORA AULACOGEN.	113
FIGURE 17 – INTERPRETED PIRAPORA AULACOGEN SEISMIC PROFILE (SEE PROFILE LINE LOCATION ON FIGURE 16B). A) FILLED BY UNDIVIDED ARAÍ AND THICK UNDIVIDED PARANOÁ SEQUENCE; B) FILLED BY ARAÍ GROUP RIFT SEDIMENTS FOLLOWED BY SAG SEQUENCE TRAÍRAS GROUP AND A THINNER PARANOÁ SEQUENCE. C) UPPER AND LOWER PARANOÁ SEDIMENTS FILLING THE AULACOGEN, SEPARATED BY AN UNCONFORMITY.	115
CAPÍTULO 5	
FIGURE 1 – SIMPLIFIED GEOLOGIC MAP SHOWING A) THE LOCATION OF THE BRASÍLIA BELT IN CENTRAL BRAZIL RELATIVE TO MAIN CRATONS AND BASINS AND B) PARANOÁ GROUP OUTCROPPING AREA IN THE SÃO FRANCISCO PALEOCONTINENT WESTERN MARGIN.....	131
FIGURE 2 – PARANOÁ GROUP LITHO-STRATIGRAPHIC COLUMN AT ITS TYPE LOCALITY (ALTO PARAÍSO DE GOIÁS / DISTRITO FEDERAL) SHOWING THE RELATED SEQUENCE STRATIGRAPHY CYCLES (MODIFIED FROM CAMPOS ET AL. 2013).....	134
FIGURE 3 – REGIONAL CORRELATION OF THE PARANOÁ GROUP WITH OTHER SEQUENCES IN THE SÃO FRANCISCO PALEOCONTINENT (MODIFIED FROM MARTINS-FERREIRA ET AL. 2017).....	137
FIGURE 4 – LOCAL GEOLOGIC MAP DISPLAYING THE SERRA DO CRISTAL FAULT ACTING AS A BOUNDARY BETWEEN STRATIGRAPHIC AND DEFORMATIONAL COMPARTMENTS.	138
FIGURE 5 - PARANOÁ GROUP LITHOSTRATIGRAPHIC COLUMNS AT BOTH EAST AND WEST SIDES OF SERRA DO CRISTAL FAULT. THE WEST SIDE COLUMN WAS INTENTIONALLY BUILT WITH A GAP IN ORDER TO EMPHASIZE THE ABSENCE OF MOST THE BASAL FORMATIONS. BASEMENT IS COMPOSED OF TRAÍRAS GROUP ROCKS.	141
FIGURE 6 - MORRO VERMELHO CONGLOMERATE A,B) HAND SAMPLES A) OCCURRING APPROXIMATELY 5 KM FROM THE SERRA DO CRISTAL FAULT AND PRESENTS REWORKED ASPECT WITH SUBROUNDED CLASTS; B) OCCURS CA. 200M FROM THE FAULT AND CONTAIN ANGULAR CLASTS OF SLATE, SILTSTONE AND QUARTZ VEIN, IS MATRIX POOR AND CONTAIN OCCASIONAL VENULATED AND BRECCIATED QUARTZITE BOULDERS C) MORRO VERMELHO CONGLOMERATE MEGACLAST (BOULDER) COMPOSED OF INTENSELY VENULATED AND BRECCIATED QUARTZITE BOULDER MEASURING CA. 6x4 M. D) CA. 120M ELEVATION SUPPORTED BY THE MORRO VERMELHO CONGLOMERATE, NOTE UNDERLAIN QUARTZITE AT THE INFERIOR LEFT CORNER.....	142
FIGURE 8 – SÃO MIGUEL CONGLOMERATE A,B) VENULATED PHYLLITE ANGULAR TO SUB-ANGULAR CLASTS; C) SLIGHTLY FOLDED CALCPHYLLITE CLASTS (INSIDE DASHED CONTOUR); D) SUBANGULAR METASILTSTONE MEGACLASTS (INSIDE DASHED CONTOUR); E) ANGULAR UNCONFORMITY AND F) SUBPARALLEL/EROSIVE UNCONFORMITY BETWEEN THE TRAIRAS AND PARANOÁ GROUPS (DASHED BLACK LINE OUTLINES UNCONFORMITY SURFACE AND DASHED WHITE LINE OUTLINES BEDDING).	143
FIGURE 9 - DIGITAL ELEVATION MODEL MAP DISPLAYING THE LOCATION OF THE TERRESTRIAL GRAVIMETRIC PROFILES AND THE SERRA DO CRISTAL FAULT SYSTEM TRACES.	144
FIGURE 10 – CORRECTED GRAVIMETRIC PROFILE DATA SHOWING ALTITUDE (M), BOUGUER (MGAL) AND FREE-AIR (MGAL) DATA. SEE PROFILES LOCATION ON FIGURE 9.	145
FIGURE 11 – GRAVIMETRIC INVERSION (LEFT COLUMN) AND DIRECT GRAVIMETRIC MODELLING (RIGHT COLUMN) RESULTS OF TERRESTRIAL PROFILES 1 TO 4 AND INTERPRETED REACTIVATED BASEMENT FAULTS. SERRA DO CRISTAL FAULT SYSTEM TRACES ARE INDICATED BY BLACK TRIANGLES.	147
FIGURE 12 - REGIONAL SEISMIC PROFILE ACROSS THE EXTERNAL BRASÍLIA BELT BAMBUÍ, VAZANTE AND PARANOÁ GROUPS STRATA UNDERLAIN BY RIFT STRUCTURES (FROM ALVARENGA ET AL. 2012).	148
FIGURE 13 – SCHEMATIC E–W-TRENDING CROSS-SECTION, SHOWING THE FRAMEWORK OF THE LITHOSTRATIGRAPHIC UNITS ACROSS SPITSBERGEN’S LATE PALEOZOIC TROUGH AND HIGHS, NORWAY (SCHEIBNER ET AL. 2015).....	148

FIGURE 14 – BRASÍLIA BELT MAP SHOWING KNOWN RIFT AND AULACOGEN PALEO STRUCTURES AND PARANOÁ PALEOCURRENT DIRECTIONS, MEASURED BY FARIA (1995) ON CROSS-BEDDING STRUCTURES IN THREE DIFFERENT REGIONS, NORTHERN (N=210), CENTRAL (N=194) AND SOUTHERN (N=432) AREAS OF PARANOÁ OCCURRENCE.	149
FIGURE 15 - TYPICAL FATES OF DIFFERENT CLASSES OF SEDIMENTARY BASIN THROUGH TIME, PLOTTED ON LOGARITHMIC SCALE. SMALL ITALICS SHOW BASEMENT TO EACH CLASS. LARGE ITALICS SHOW DEFORMATIONAL OR THERMAL CONSEQUENT FATES OF EACH CLASS (FROM WOODCOCK 2004).	151
FIGURE 16 - MAP OF STRATAL THICKNESS ON THE NORTH AMERICAN CRATON PRESENTED BY BURGESS (2008).	155
FIGURE 17 - TRUE-SCALE ACTUALISTIC ANALOG MODELS FOR SEDIMENTARY BASINS IN INTRAPLATE SETTING (AFTER INGERSOLL 2012).	156
FIGURE 18 - SCHEMATIC CROSS-SECTIONS OVER THE CANADIAN CRATON SHOWING THE DEVELOPMENT OF EPIROGENIC ARCHES IN TIME (SANFORD 1987).	158
FIGURE 19 - INTERPRETED SCHEMATIC SECTION FOR THE JUSCELÂNDIA RIFTING PROPOSED BY MORAES ET AL. (2006).	162
FIGURE 20 – REGIONAL STRATIGRAPHIC CORRELATION CHART OF THE PROTEROZOIC SÃO FRANCISCO CRATON AND MARGINS FIRST-ORDER SEQUENCES SHOWING INTERPRETED BASIN TYPES, TECTONIC SETTINGS AND EVENTS (MODIFIED FROM ALKMIM AND MARTINS-NETO 2012, REIS ET AL. 2017, WITH GLOBAL AGE DATA ACQUIRED FROM LI ET AL. 2008, BOGDANOVA ET AL. 2009, PISAREVSKY ET AL. 2003, ZHAO ET AL. 2003, JOHANSSON ET AL. 2014 AND REGIONAL U-Pb AGE DATA COMPILATION PRESENTED ON TABLE 1, CORRELATION EXPLANATION IN TEXT).....	165
CAPÍTULO 6	
FIGURA 1 - LOCALIZAÇÃO DA SALIÊNCIA DO MOQUÉM NA FAIXA BRASÍLIA (MODIFICADA DE FUCK, 2004).	183
FIGURA 2 - EXEMPLO DE RESULTADO DO MÉTODO DE MONTAGEM FOTOGRÁFICA SEQUENCIAL E VETORIZAÇÃO DE CAMADAS DOBRADAS E FALHAS DE CAVALGAMENTO (LINHAS PONTILHADAS) DA SEÇÃO 1, USADA NA MONTAGEM DO PERFIL CENTRAL (LOCALIZAÇÃO DO PERFIL NA FIGURA 5).	184
FIGURA 3 - MAPA ESTRUTURAL SIMPLIFICADO SOBRE MODELO DIGITAL DE ELEVAÇÃO DA ÁREA DE ESTUDO. EM AZUL TRACEJADO ENCONTRA-SE DESTACADA A SALIÊNCIA DO MOQUÉM. EM TRAÇO BRANCO, OS LIMITES ENTRE OS SETORES NORTE, CENTRAL E SUL DA SALIÊNCIA.	186
FIGURA 4 - PERFIL FOTO-MAPEADO DO TREM DE DOBRAS DO DOMÍNIO MOQUÉM CENTRAL. O PERFIL POSSUI DIREÇÃO E-W, EXTENSÃO DE 35 KM E EXAGERO VERTICAL DE 2 VEZES (LOCALIZAÇÃO DO PERFIL NA FIGURA 5).	187
FIGURA 5 - SALIÊNCIA DO MOQUÉM INDICADA SOBRE MODELO DIGITAL DE ELEVAÇÃO. SETAS PRETAS INDICAM A DIREÇÃO DE TRANSPORTE TECTÔNICO OU STRESS HORIZONTAL MÁXIMO (SHMAX). A LOCALIZAÇÃO DO PERFIL CENTRAL ESTÁ INDICADA PELA LINHA BRANCA CONTÍNUA.	187
FIGURA 6 – A) SLICKENLINES DESENVOLVIDAS SOBRE SUPERFÍCIE SILICOSA (SLICKENSIDE) PARALELA A $S_{0/1}$. B) LINEAÇÃO DE ESTIRAMENTO MINERAL (Lx) DE ESCALA MÉTRICA FORMANDO FITAS DE CALCITA. $Lx = 15/160$. AS FITAS DE CALCITA SÃO PARALELAS A S_3 -TRANSCORRENTE E ENCONTRAM-SE CRENLADAS POR CLIVAGEM S_3 -NORMAL. C) DETALHE DA FOLIAÇÃO S_3 -NORMAL CRENLANDO S_0 . VISTA SOBRE O PLANO DE S_2 . D) BOUDINAGEM DE CAMADA DE QUANTZITO INTERCALADA EM METASSILTITOS DO GRUPO ARAÍ. A DIREÇÃO DE BOUDINAGEM É PERPENDICULAR A S_3 -NORMAL= $300^\circ/75^\circ$. E) LENTE CARBONÁTICA DEFORMADA NOS CALCIFILITOS DA FM. TRAÍRAS DO GRUPO ARAÍ A OESTE DA FALHA DA SERRA DO CRISTAL. RAZÃO AXIAL NO PLANO XY DO ELIPSOIDE DE DEFORMAÇÃO (S_3 -TRANSCORRENTE): $X/Y=4,68$. F) VEIOS ESCALONADOS ROTACIONADOS INDICANDO CINEMÁTICA SINISTRAL E REGIME SEMI-RÚPTIL.	189
FIGURA 7 - AFLORAMENTO DA UNIDADE CALCIFILITO, TOPO DO GRUPO ARAÍ, AFETADO PELO CORREDOR DE TRANSCORRÊNCIA DO SUBSISTEMA TOCANTINZINHO QUE GERA A FOLIAÇÃO LOCAL S_3 -TRANSCORRENTE, QUE POR SUA VEZ É CRENLADA PELA FOLIAÇÃO REGIONAL TARDIA S_3 -NORMAL. A INTERSECÇÃO ENTRE ESSAS DUAS FOLIAÇÕES GERA LINEAÇÃO DE INTERSECÇÃO (Li) BASTANTE EXPRESSIVA EM CAMPO.	190

FIGURA 8 - PROJEÇÃO ESTEREOGRÁFICA (HEMISFÉRIO INFERIOR) DAS FOLIAÇÕES MAPEADAS NA PORÇÃO CENTRO-NORTE DO DOMÍNIO MOQUÉM.	190
FIGURA 9 - MAPA ESTRUTURAL SIMPLIFICADO SOBRE MAPA RESIDUAL DA ANOMALIA BOUGUER. ESTRUTURAS REPLICADAS DA FIGURA 3.	191
FIGURA 10 - MODELOS IDEALIZADOS PARA EXPLICAR CURVATURAS OROGÊNICAS ILUSTRANDO A EVOLUÇÃO DE FALHAS INDIVIDUAIS EM CADA CENÁRIO. LINHAS TRACEJADAS REPRESENTAM MATERIAL ANTERIOR À DEFORMAÇÃO; LINHAS CHEIAS REPRESENTAM O MESMO MATERIAL APÓS AVANÇO; SETAS CHEIAS INDICAM A DIREÇÃO DE TRANSPORTE; SETAS VAZIAS INDICAM A DIREÇÃO DE PROPAGAÇÃO DE FALHA; SETAS COM PONTA DUPLA INDICAM DIREÇÕES DE EXTENSÃO TANGENCIAL DA SALIÊNCIA (WHISNER <i>ET AL.</i> 2014).	192

LISTA DE TABELAS

CAPÍTULO 1

TABELA 1 – LISTA DE AMOSTRAS COLETADAS E INFORMAÇÕES DE LOCALIZAÇÃO, ESTRATIGRAFIA, TIPO DE ROCHA E NÚMERO DE GRÃOS DE ZIRCÃO ANALISADOS.	9
---	---

CAPÍTULO 2

TABELA 1 – RELAÇÃO ENTRE ESTRUTURAS GEOLÓGICAS E OS TIPOS DE LINEAMENTOS EXTRAÍDOS.	30
TABELA 2 – CARACTERÍSTICAS PRINCIPAIS DOS COMPARTIMENTOS GEOMORFOLÓGICOS.	37

CAPÍTULO 3

TABLE 1 - ANALYZED SAMPLES AND RESPECTIVE ROCKTYPES, FORMATIONS AND UTM Z23S COORDINATES.	50
TABLE 2 - SUMMARY OF U-Pb ZIRCON DATA FOR STUDIED SAMPLES OBTAINED BY LA-SF-ICP-MS METHOD.	54

CAPÍTULO 5

TABLE 1 – AGES (U-Pb) AND BASIN CLASSIFICATION DATA COMPILATION USED FOR THE REGIONAL CORRELATION CHART PRESENTED ON FIGURE 20. MAJOR COMPILATION SOURCE IS DERIVED FROM HEILBRON <i>ET AL.</i> (2017), PIMENTEL (2016) AND FERREIRA FILHO <i>ET AL.</i> (2010).	166
--	-----

RESUMO

A evolução tectônica de bacias proterozóicas desenvolvidas na margem oeste do Cráton São Francisco é investigada por meio de estudo multidisciplinar envolvendo mapeamento geológico e estrutural de detalhe, levantamentos gravimétricos e geocronologia pelos sistemas U-Pb e Lu-Hf em zircão detrítico. Estudos de reconhecimento preliminar da área foram realizados por meio da análise de mapas de relevo, drenagem, solos e geologia, resultando no desenvolvimento de método específico de análise de lineamentos e compartimentação geomorfológica como suporte para os estudos de evolução tectônica. Novos dados de idades U-Pb de 1.54 Ga. obtidas nos zircões mais jovens da Formação Traíras, aliados ao mapeamento de discordância erosiva entre esta e a Formação Arraias, permitiram estabelecer um lapso temporal entre essas formações de, no mínimo 228 Ma. Interpretada como uma nova sequência, propõe-se que a Formação Traíras seja elevada à categoria de grupo e seus membros à categoria de formações, ficando o Grupo Araí, restrito à bacia do tipo rifte Estateriana e o Grupo Traíras restrito à bacia do tipo sag Calimiana. A distinção dessas duas fases extensionais, somadas à fase Esteniana que gerou a sequência Paranoá, permitiu correlação mais precisa com as fases extensionais registradas na margem leste do Cráton, representadas pelo Supergrupo Espinhaço, o que levou à proposição do Supergrupo Veadeiros, que engloba os três pulsos extensionais registrados na margem oeste, representados pelas sequências de primeira ordem Araí (Veadeiros Inferior), Traíras (Veadeiros Médio) e Paranoá (Veadeiros Superior). Valores $\epsilon_{\text{Hf}}(t)$ positivos obtidos em zircões Riacionos sugerem proveniência do Maciço de Goiás. Este indício, aliado ao fato da Suíte Serra da Mesa intrudir tanto o Maciço de Goiás quanto o embasamento da sequência Traíras, indica que o maciço já havia sido acrescido ao Cráton São Francisco antes do período Calimiano. Um novo limite para o paleocontinente São Francisco a partir do período Estateriano é proposto. As principais feições como falhas, grabens, horsts e centro vulcânico do rifte intracontinental Araí, foram mapeadas com o auxílio de estudos gravimétricos. Também com o auxílio da gravimetria, foram analisadas as profundidades dos corpos plutônicos anorogênicos e das falhas geradas durante o rifteamento. Os resultados permitiram caracterizar o paleorifte Araí, na região estudada, como do tipo passivo, estreito a divergente, composto por três segmentos principais, que produziu magmas anorogênicos tipo rapakivi ainda hoje alojados na crosta desde a superfície até no mínimo 19 km de profundidade. O paleorifte Araí é parte de um sistema de riftes estaterianos que contornam o Cráton São Francisco. Variações abruptas na sequência estratigráfica do Grupo Paranoá foram investigados com o auxílio de perfis gravimétricos terrestres. Constatou-se que o rifte Araí tem continuidade sob a cobertura sedimentar do Grupo Paranoá e que a instalação da sequência Paranoá foi controlada por reativações das estruturas do rifte, gerando altos epirogênicos que atuaram como barreiras de

isolamento de parte da bacia. A sequência Paranoá foi caracterizada como bacia do tipo Margem Cratônica interligada a bacias intracratônicas e possivelmente de margem passiva em fases de nível eustático elevado. A inversão das bacias estudadas se deu ao final do Neoproterozóico, durante o Ciclo Orogenético Brasileiro. Na área estudada um dos produtos dessa inversão é a Saliência do Moquém, feição regional localizada em zona de antepaís, formada por sistema de dobras e empurrões apresentando forma geral arqueada com concavidade voltada para o orógeno. A feição foi estudada por meio de mapeamento geológico-estrutural de detalhe nos seus diferentes domínios (norte, central e sul) e caracterizada como produto de descolamento horizontal raso tardio no Ciclo Brasileiro. Questões como a influência da arquitetura do embasamento na propagação do empurrão e no padrão deformacional gerado foram investigadas com o auxílio de dados gravimétricos residuais. De acordo com o estudo, os modelos de propagação de saliências que mais se adequam à Saliência do Moquém são o mecanismo de Transporte Divergente com controle de Barreiras formadas por rampas laterais ou oblíquas. Concluiu-se que, durante o desenvolvimento da Saliência do Moquém, o embasamento já se encontrava alto a norte e a Falha da Serra do Cristal já representava uma descontinuidade significativa. Essas duas feições exerceram importante controle na evolução da saliência como rampas laterais e o relevo do embasamento ocorre diretamente associado ao desenvolvimento de braqui-anticlinais e braqui-sinclinais. Propõem-se que a tensão diferencial pontual responsável pelo transporte divergente que gerou a saliência pode ter sido produzida pela exumação do corpo denso do complexo máfico-ultramáfico de Niquelândia. As conclusões do estudo indicam que a arquitetura do embasamento, marcada por falhas e blocos do paleorifte Araí e bacias subsequentes, exerceu papel fundamental no controle da deformação da Faixa Brasília durante o Ciclo Orogenético Brasileiro, de modo que estudos de análise estrutural relativos a essa fase devem considerar as anisotropias herdadas como controles importantes da deformação.

Palavras-chave: Grupo Traíras, Supergrupo Veadeiros, paleocontinente São Francisco, paleorifte Araí, granitos rapakivi, Grupo Paranoá, Saliência do Moquém.

ABSTRACT

The tectonic evolution of Proterozoic basins developed on the São Francisco Craton western margin is investigated by means of a multidisciplinary study involving geological and structural detailed mapping, gravimetric surveys, U-Pb and Lu-Hf geochronology in detrital zircon. Preliminary reconnaissance studies of the area were carried out through the analysis of relief, drainage, soil and geological maps, resulting in the development of a specific method of lineament analysis and geomorphological compartmentation as support for tectonic evolution studies and is presented in the first article of this thesis. U-Pb ages ca. 1.54 Ga., obtained in the younger zircons of the Traíras Formation, together with the mapping of an erosional unconformity between this and the Arraías Formation, allowed to establish a temporal gap between these formations of at least 228 Ma. According to the new data, it is proposed to raise the Traíras Formation to group status and its members to formation status, being the Araí Group restricted to the Statherian rift basin and the Traíras Group restricted to the Calymmian sag basin. The distinction of these two extensional pulses, together with the Stenian pulse that generated the Paranoá basin, allowed a more precise correlation with the extensional pulses recorded on the eastern margin of the Craton, represented by the Espinhaço Supergroup, which led to the proposition of the Veadeiros Supergroup, encompassing the three pulses and represented by the first-order sequences Araí (Lower Veadeiros), Traíras (Middle Veadeiros) and Paranoá (Upper Veadeiros). Positive $\epsilon_{\text{Hf}}(t)$ values obtained in Rhyacian zircons suggest sediment provenance from the Goiás Massif. That new evidence, coupled with the fact that the Serra da Mesa Suite intrudes both the Goiás Massif and the basement of the Traíras sequence, indicates that the massif had already been accreted to the São Francisco Craton before the Calymmian period. A new limit for the São Francisco paleocontinent after the Statherian period is proposed. The main features such as faults, grabens, horsts and volcanic center of the Araí intracontinental rift were mapped with the aid of gravimetric studies. Also with the aid of gravimetry, the depths of anorogenic plutonic bodies and rift faults were analyzed. The results allowed to characterize the Araí paleorift, in the studied region, as a passive, narrow to divergent type, composed of three main segments, that produced anorogenic rapakivi-type magmas still hosted in the crust from the surface to at least 19 km deep. The Araí paleorift is part of a system of Statherian rifts that surround the São Francisco Craton. Abrupt stratigraphic variations in the Paranoá Group are investigated with the aid of terrestrial gravimetric profiles. It is verified that the Araí rift continues under the Paranoá Group sedimentary cover and that the installation of the Paranoá sequence was controlled by reactivation of the rift structures, generating epirogenic highs that acted as barriers, isolating parts of the basin. The Paranoá sequence is characterized as Cratonic Margin type connected to intracratonic basins and possibly connected to a passive

margin basin in phases of high eustatic level. The inversion of the studied basins occurred at the end of the Neoproterozoic era, during the Brasiliano Orogeny. In the studied area one of the products of this inversion is Moquém Salient, a regional scale foreland feature, formed by a concave-to-the-orogen thrust and fold system. The feature was studied through detailed geological and structural mapping in its different domains (north, central and south) and was characterized as a late Brasiliano thrust. The influence of basement architecture over thrust propagation and deformational pattern was investigated with the aid of residual gravimetric data. According to the study, the thrust propagation models that best fit the Moquém Salient evolution are the Divergent Transport mechanism with controls by Lateral or Oblique Ramp Barriers. The analysis led to the conclusion that during the development of the Moquém Salient, the basement was already high to the north and the Serra do Cristal Fault already represented a significant discontinuity. These two features exerted important control in the Salient evolution as lateral ramps and the basement relief was shown to be directly associated to the development of brachi-anticlines and brachi-synclines. It was proposed that the localized differential stress, responsible for the divergent transport that generated the Moquém Salient may have been produced by the exhumation of the Niquelândia mafic-ultramafic complex dense body. The conclusions of this study indicate that basement architecture, marked by faults and blocks developed during the evolution of the Araí paleorift and subsequent basins, played a fundamental role in controlling the deformation during the Brasiliano Orogeny, so that structural analysis studies related to this phase, should consider these anisotropies as important controls for deformation.

Keywords: Traíras Group, Veadeiros Supergroup, São Francisco paleocontinent, Araí paleorift, rapakivi granites, Paranoá Group, Moquém Salient.

CAPÍTULO 1

INTRODUÇÃO

1.1 Apresentação

A evolução tectônica da Faixa Brasília tem sido abordada desde a década de 1960 até os dias de hoje por diversos autores, Almeida (1968), Hasui & Almeida (1970), Marini *et al.* (1981, 1984), Fuck & Marini (1981) Costa (1984), Fuck *et al.* (1989, 1993, 1994, 2001), Fonseca (1996), Fonseca *et al.* (1995, 1997), Dardenne (1978, 2000), Pimentel *et al.* (2004, 2011), Valeriano *et al.* (2004, 2008), D'el-Rey Silva *et al.* (2011), Uhlein *et al.* (2012), Pimentel (2016).

Em clássico trabalho sobre a evolução geotectônica da Faixa Brasília, Marini *et al.* (1981) relatam que as características descritas por Milanosky (1978) para os riftes intracontinentais são encontradas, em quase sua totalidade, na Faixa Brasília: "*grandes falhas retilíneas subparalelas e escalonadas; mergulho das falhas entre 60 e 65°, podendo o rejeito de cada falha exceder a 1.000m; alternância de vários blocos lineares e longos separados por falhas lineares subparalelas; falhas subparalelas às estruturas tectônicas do substrato, a maior parte sendo herança de antigas zonas de fraqueza*".

Os autores mencionam ainda que na ocasião da fase compressiva Neoproterozoica da Faixa Brasília, é provável que grande parte das estruturas que vinham atuando como falhas normais tenham invertido o sentido de deslocamento de seus blocos e passado a atuar então como falhas reversas ou mesmo de empurrão.

Com os dados estratigráficos e geofísicos atualmente disponíveis é possível abordar em maior detalhe o contexto tectono-estratigráfico dos ciclos distensivos e compressivos e sua interação, o que pode revelar detalhes inusitados a respeito do cenário evolutivo das bacias Proterozoicas do Cráton São Francisco e suas margens, bem como os principais elementos condicionantes da tectônica Brasileira na região.

Ao longo de sua história evolutiva, os continentes sofrem repetidos eventos de reativação de estruturas pré-existentes, reequilíbrio isostático, retrabalhamento de terrenos, geração de nova trama estrutural, imposição de paragêneses metamórficas e geração de corpos magmáticos. Tal comportamento se deve principalmente ao fato de que as crostas continentais quartzo-feldspáticas não podem ser subduzidas em função de sua alta flutuabilidade e fragilidade em relação à crosta oceânica e ao manto litosférico subjacente. Como consequência, os continentes

derivam eternamente à mercê da tectônica de placas, o que lhes confere uma alta suscetibilidade a novos eventos tectono-termiais, mesmo quando venham a ocorrer após centenas de milhares de anos de quiescência tectônica (Sutton & Watson 1986). Esse fato faz com que novas bacias sejam geradas sobre bacias pré-existentes, muitas vezes com um lapso de tempo de centenas de milhares de anos entre elas, mas controladas pelos mesmos mecanismos e zonas de fraquezas.

Diversos fatores controlam o *locus* de atividades tectônicas, todavia apenas dois fatores parecem ser de primeira ordem: (1) variações espaciais e temporais no estado termal da litosfera; e (2) a presença de defeitos mecânicos ou fraquezas como falhas, zonas de cisalhamento e fronteiras composicionais (Holdsworth et al. 2001).

O presente estudo é focado na dinâmica de sucessão das bacias intracratônicas do Proterozoico na margem oeste do Cráton São Francisco, com ênfase nas bacias Araí (Tanizaki et al. 2015) e Paranoá (Faria 1995).

A questão das dimensões, localização e orientação do sistema de riftes Proterozoicos da Faixa Brasília foi pouco abordada até o presente. Reis & Alkmim (2015) alegam que isso ocorre em todo o Cráton São Francisco e suas margens e citam o exemplo do aulacógeno Pirapora, identificado através de seções sísmicas e aerogeofísica, mas ainda pouco estudado.

A extensão, duração, preenchimento e inversão desses sistemas rifte são ainda pouco compreendidos. Alguns exemplos como o aulacógeno Paramirim (Alkmim & Martins-Neto 2012) ou o rifte do Espinhaço (Danderfer Filho et al. 2015; Alkmim et al. 1993) tem sido alvo de estudos mais detalhados e, portanto, são mais bem compreendidos nesses aspectos. Todavia existem outros aspectos ainda pouco explorados, como por exemplo a temporalidade e o controle exercido por falhas profundas, geradas durante o rifteamento, na instalação e evolução de novas bacias ou em seu processo inversão.

São poucos os estudos que tratam da dinâmica de abertura da bacia Paranoá. A maior parte dos trabalhos trata de detalhamento estratigráfico e procura descrever paleoambientes, geralmente classificando-a, tradicionalmente, como bacia de margem passiva.

Campos *et al.* (2013) propõem processos e controles deposicionais em diferentes tratos de sistemas e ambientes marinhos baseados nos dados estratigráficos levantados em campo e caracterizam a bacia como de margem passiva, sob regime de subsidência flexural, com influência de processos marinhos em regimes ora retrogradacionais, ora progradação, resultando em um preenchimento sedimentar com ampla continuidade lateral. Todavia, não avançam nos estudos em relação à evolução tectônica da bacia.

Valeriano *et al.* (2004) relacionam a abertura da sequência Paranoá com a fragmentação do

supercontinente Rodínia no final do Mesoproterozoico e a caracterizam como do tipo rifte que evoluiu para margem passiva na margem oeste do Cráton São Francisco.

Guimarães (1997) concluiu que os arcóseos e quartzitos da parte superior da sequência Paranoá são quimicamente semelhantes a sedimentos clásticos de margem passiva.

Alvarenga et al. (2012) interpretam os ambientes de sedimentação como variando de marés, com exposições esporádicas subaéreas, a ambientes de plataformas intermediárias, com fácies frequentes relacionadas a tempestades. Além disso demonstram que o Grupo Paranoá mostra refletores internos horizontais em seções sísmicas, típicos de bacias do tipo *sag*.

Guadagnin & Chemale (2015) afirmam, também com base em perfis sísmicos, que o Grupo Paranoá foi depositado sobre o Cráton São Francisco como uma extensão da sequência Espinhaço Superior, mais especificamente relacionado com o Grupo Conselheiro Mata, interpretando-o como uma bacia do tipo *rift-sag* na margem oeste do Cráton São Francisco e como uma bacia intracratônica quando depositada sobre este.

Reis & Alkmim (2015) também relacionam o Grupo Paranoá à sequência Espinhaço Superior, que chamam de sequência Paranoá-Espinhaço II, e a caracterizam como produto de preenchimento de bacia do tipo *rift-sag* que gradua lateralmente para depósitos de margem passiva, desenvolvida ao longo da margem oeste do cráton São Francisco. Fazendo uso de perfis sísmicos, afirmam ainda que a sequência se estende por vastas áreas da bacia e aumenta em espessura para oeste, preenchendo o aulacógeno Pirapora (Alkmim & Martins-Neto, 2001), onde pode ter espessura de até 7 km e aparentemente cobre outra sequência sedimentar precambriana desconhecida, possivelmente relacionada aos sedimentos do Grupo Araí de sua fase rifte e/ou *sag*.

As idades de abertura e inversão da sequência Paranoá são ainda um tanto imprecisas. Estudos dos estromatólitos das formações Córrego do Barreiro e Ribeirão Piçarrão indicam idade deposicional estimada entre 1.0 e 1.3 Ga (Dardenne *et al.* 1971, Cloud & Dardenne 1973, Dardenne 1978, Faria 1995) ou 1.170 e 950 Ma (Fairchild *et al.* 1996). A partir de estudos isotópicos U-Pb e Lu-Hf em zircões detríticos com sobrecrecimento diagenético de xenotima, Matteini *et al.* (2012) concluem que a idade mínima de deposição do Grupo Paranoá de 1.042 Ma e idade máxima de 1.542 Ma, respectivamente idades de cristais de xenotima diagenética e de população de zircões mais jovens encontrados na base do Grupo.

1.2 Justificativa

O entendimento dos processos que conduzem à formação e evolução de bacias

sedimentares é elemento-chave na elucidação da evolução tectônica da litosfera continental (Ziegler 1994, Cloetingh *et al.* 2015).

O presente projeto de pesquisa objetiva, de modo geral, contribuir para o detalhamento do entendimento da evolução tectono-estratigráfica da margem oeste do Craton São Francisco a partir da investigação da evolução das sequências Araí e Paranoá na Faixa Brasília Norte. Para tanto, pretende-se localizar no tempo e no espaço a fase rifte do Grupo Araí e investigar a relação temporal e tectônica desta fase com as fases extensionais subsequentes, caracterizadas pelas bacias representadas pelas sequências sedimentares da Formação Traíras e do Grupo Paranoá. Por fim, pretende-se identificar a influência de estruturas pré-brasillianas em termos de reativações e retrabalhamento durante o evento Brasileiro.

O projeto foi idealizado após o mapeamento da Folha Alto Paraíso de Goiás (SD.23-Y-A-I), em escala 1:100.000 no ano de 2011 no âmbito do programa PRONAGEO. A interpretação dos dados de campo resultou no levantamento de questões fundamentais de cunho evolutivo sobre a área mapeada até então não abordadas por outros autores.

Os dados gerados e ou modelados especificamente para o presente estudo são provenientes das seguintes técnicas:

- Gravimetria terrestre;
- Geocronologia U-Pb e isótopos de Hf e
- Mapeamento geológico e estrutural de detalhe.

A fim de identificar o estado da arte do conhecimento da geologia local, foi realizado levantamento bibliográfico detalhado dos dados geológicos pré-existentes que, além de artigos científicos, incluem os relatórios e mapas de trabalhos de formatura, relatórios de pesquisa, dissertações de mestrado e teses de doutorado elaboradas por professores e estudantes de diversas instituições de ensino superior, além dos relatórios de mapeamento feitos pela CPRM, entre outros.

As principais bases cartográficas e notas explicativas relativas à região incluem, entre outros, os trabalhos de:

- Baeta *et al.* (1977): Projeto Manganês contendo mapa geológico em escala 1:100.000;
- Botelho *et al.* (1998): cartografia com estudantes da Universidade de Brasília no Projeto Cavalcante;
- Martins (1999): Mapeamento do Parque Nacional da Chapada dos Veadeiros e adjacências com dados compilados em Dissertação de Mestrado com cartografia geológica em escala 1:100.000;
- Dardenne *et al.* (1999): Relatório submetido à FAP-DF no âmbito do projeto Geologia do Distrito Federal e Entorno, que incluiu seções geológicas na Folha Alto Paraíso de Goiás

- na escala de 1:100.000;
- Campos (2009): Mapeamento da bacia do Rio Tocantinzinho no âmbito do Estudo Integrado de Bacia Hidrográfica EIBH.
 - Campos *et al.* (2011, inédito): Folha Alto Paraíso de Goiás 1:100.000 (PRONAGEO /CPRM).

1.3 Objetivos

O presente estudo tem como objetivo responder as seguintes questões:

- Qual a localização dos principais elementos (falhas, *grabens*, *horsts*, centros vulcânicos) do rifte Araí?
- Qual a relação temporal da fase rifte (Formação Arraias) com a atualmente acreditada fase sag (Formação Traíras) do grupo Araí?
- Existe a correlação entre a Formação Traíras do Grupo Araí e demais sequências que ocorrem na Faixa Brasília?
- Qual a influência das estruturas da fase rifte na fase sag da Formação Traíras?
- Qual a influência das estruturas da fase rifte na instalação e evolução da sequência Paranoá?
- Até que ponto as estruturas pré-Brasilianas influenciaram a dinâmica da deformação brasileira?

1.4 Localização

O estudo foi conduzido em três escalas, de modo que a área de estudo abrange, para fins de correlação estratigráfica regional, todo o Cráton São Francisco e suas margens, com ênfase nas margens leste e oeste. Para fins de análise de distribuição dos eixos do rifte Araí e superposição de bacias, a área de estudo abrange a Faixa Brasília Norte e a porção oeste do Cráton São Francisco (Figura 1), para fins de detalhamento envolvendo coleta de amostras e mapeamento geológico-estrutural em áreas-chave, abrange a região conhecida como Chapada dos Veadeiros, localizada na porção central da zona externa da Faixa Brasília Norte (Figura 2).

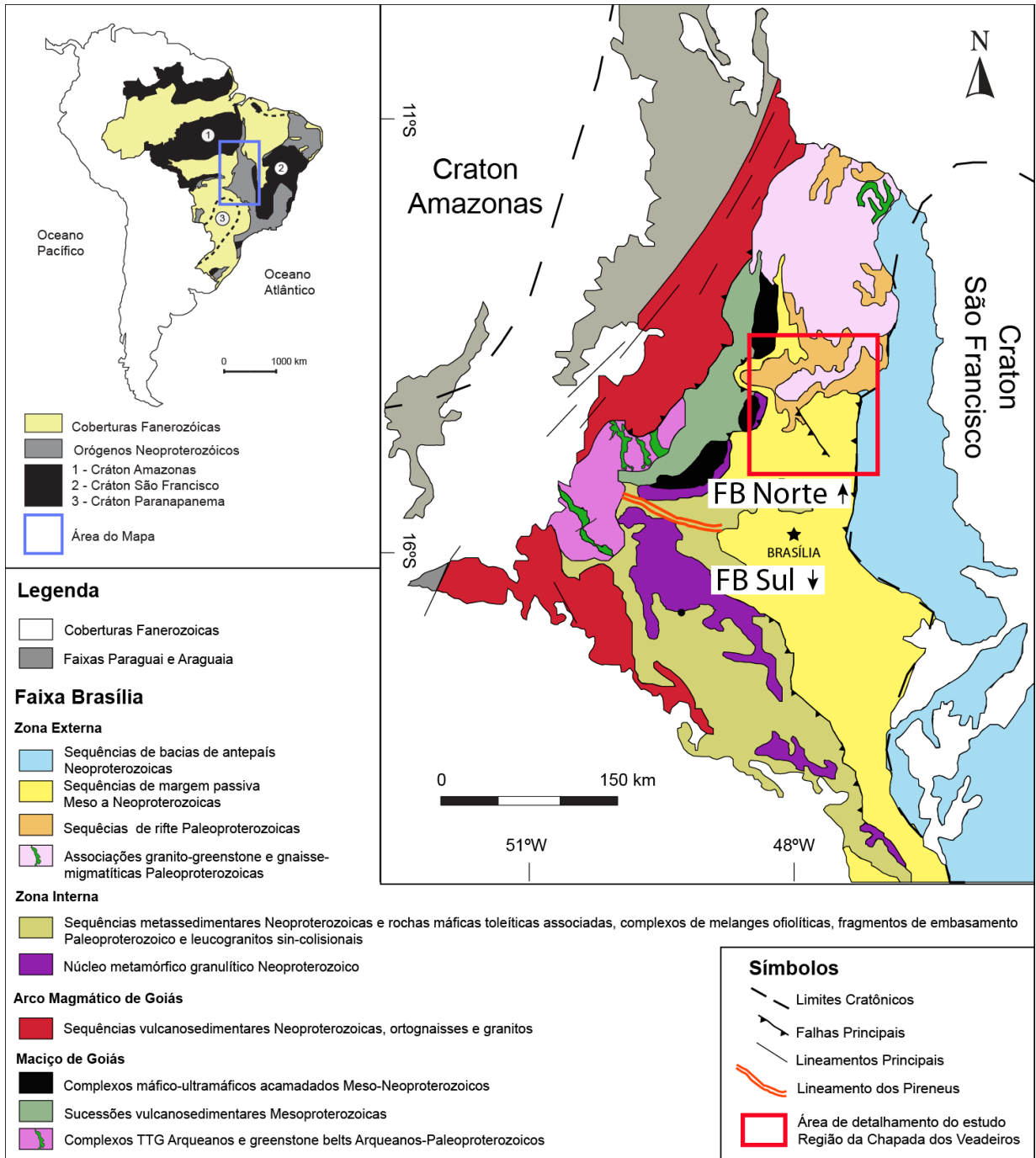


Figura 1 - Mapa de localização da área de detalhamento do estudo na Faixa Brasília. FB=Faixa Brasília. Os domínios norte e sul (FB Norte e FB Sul) são delimitados aproximadamente pelo lineamento dos Pireneus à altura de Brasília.

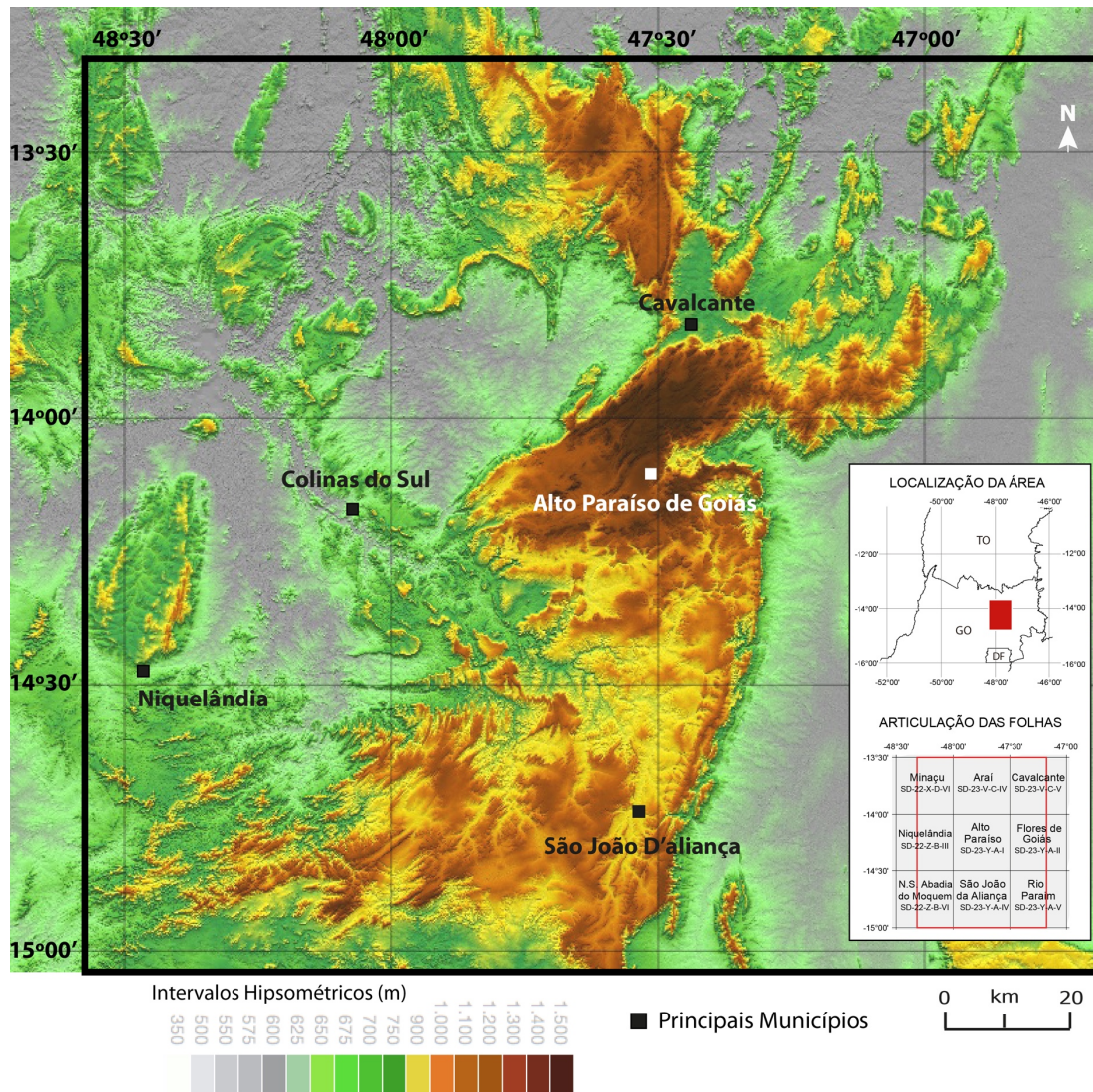


Figura 2 - Mapa de localização da área de detalhamento do estudo sobre imagem SRTM colorida com mapa de posicionamento no estado de Goiás e articulação das folhas cartográficas que abrange.

1.5 Estruturação da Tese

A presente tese de doutorado é estruturada na forma de 5 artigos científicos. Cada artigo representa um capítulo da tese e todos estão inseridos entre o capítulo de introdução e o de discussões finais e conclusões.

O capítulo de introdução apresenta os problemas, objetivos e a justificativa do estudo, a localização da área, a estruturação da tese, os métodos utilizados na pesquisa, além do referencial teórico básico.

Cada um dos cinco artigos foi elaborado durante o desenvolvimento da tese, aproximadamente na ordem em que se são apresentados e abordam temas específicos do estudo, sendo que todos os temas possuem a finalidade comum de resolver os problemas propostos.

O capítulo de discussões finais e conclusões procura integrar os temas apresentados individualmente com o intuito de elaborar hipóteses que englobem as conclusões de todos os artigos conjuntamente. A Figura 3 ilustra a estruturação da tese e apresenta o status de publicação dos artigos que dela fazem parte.

Afim de facilitar a leitura, as referências bibliográficas referentes às citações de cada capítulo são listadas sempre ao seu final.

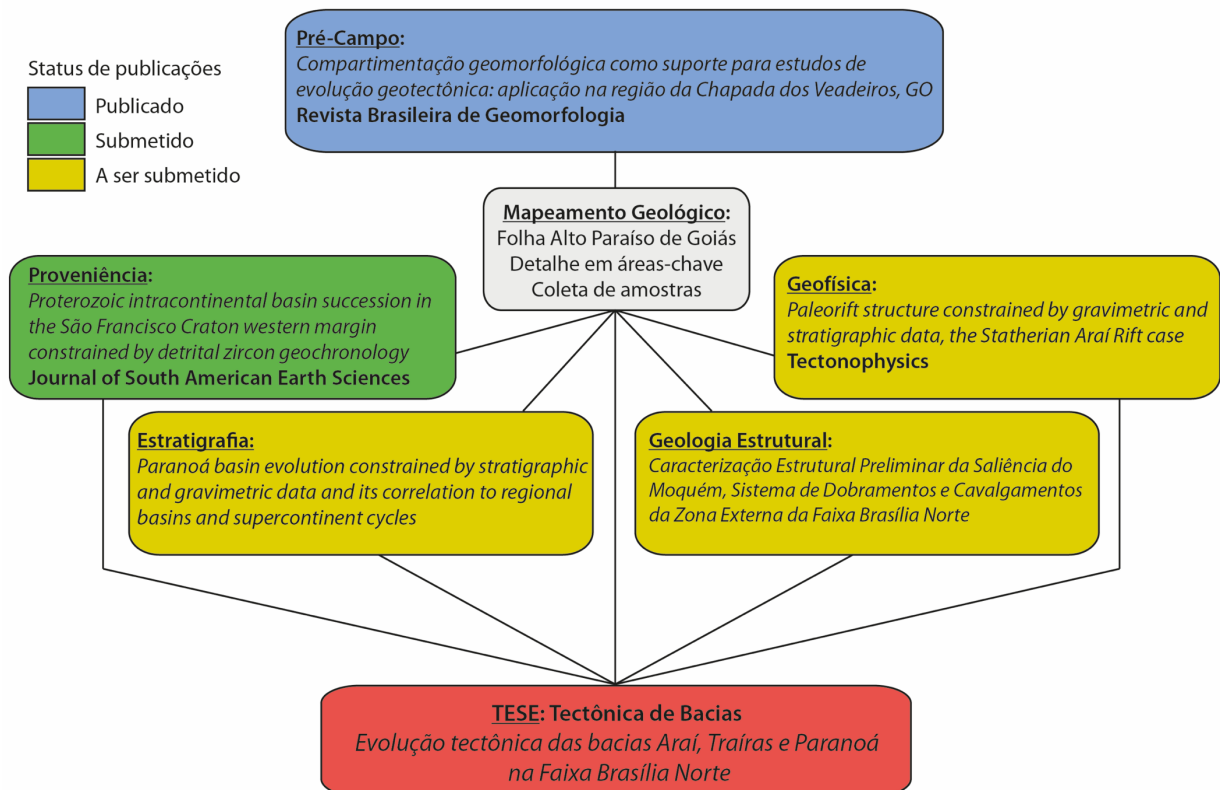


Figura 3 - Diagrama esquemático da estruturação da presente tese.

1.6 Materiais e Métodos

Amostragem

Foram coletadas 12 amostras de rochas metassedimentares dos grupos Araí e Paranoá ao longo de 2 perfis na região do Povoado de São Jorge, área que abrange a zona de contato entre estes grupos e está localizada dentro da Folha Alto Paraíso de Goiás (SD.23-Y-A-I), recentemente mapeada em escala 1:100.000. Um recorte da Folha Alto Paraíso de Goiás mostrando a localização exata de cada amostra é exibido na Figura 2 do Capítulo 3.

O objetivo da amostragem foi coletar grãos de zircão para análises pelos métodos U-Pb e Lu-Hf para estudos de proveniência de sedimentos. Apenas de 11, das 12 amostras coletadas, foi possível extrair zircões. A Tabela 1 lista as amostras coletadas fornecendo as coordenadas

geográficas, tipo de rocha e número de grãos de zircão analisados em cada amostra.

Das 11 amostras que continham zircão foram analisados um total de 265 grãos em 397 *spots*, dos quais, 247 *spots* geraram dados considerados consistentes, tendo sido o restante eliminado por valores de concordância maiores que 105 ou menores que 95 e por suspeita de dano por excesso de U, comum em amostras muito antigas. As amostras 8, 9 e 11 não foram analisadas por insuficiência de grãos.

Tabela 1 – Lista de amostras coletadas e informações de localização, estratigrafia, tipo de rocha e número de grãos de zircão analisados.

Amostra	X	Y	Grupo	Formação - Membro	Rocha	Zircões
TFC_01	181543	8428261	Paranoá	Ribeirão São Miguel	Conglomerado	44
TFC_02	182596	8427837	Araí	Traíras – Boqueirão	Grauvaca	13
TFC_03	184895	8421872	Araí	Traíras – Rosário	Quartzito	55
TFC_04	188062	8425939	Araí	Traíras – Rio Preto	Quartzito	10
TFC_05	193811	8430754	Araí	Traíras – Rosário	Quartzito	60
TFC_06	197858	8431200	Araí	Traíras – Rosário	Quartzito	11
TFC_07	198998	8430014	Paranoá	Ribeirão São Miguel	Conglomerado	20
TFC_08	188620	8408704	Paranoá	Serra da Meia Noite	Quartzito	3
TFC_09	189652	8409467	Paranoá	Ribeirão Piçarrão	Siltito	1
TFC_10	193205	8411983	Paranoá	Ribeirão São Miguel	Conglomerado	23
TFC_11	200709	8414800	Paranoá	Ribeirão Piçarrão	Siltito	0
TFC_12	203264	8419879	Paranoá	Serra do Paranã	Quartzito	29

Separação mineral e montagem das amostras

O processo de separação mineral seguiu os padrões sequenciais comuns: cominuição por britador, peneiramento, concentração de pesados por bateia e separação magnética por meio do equipamento Frantz® *Magnetic Separator* e, quando necessário, separação por densidade por meio de líquido denso (Bromofórmio). As amostras de conglomerado foram preparadas de forma a separar clastos e matriz e apenas o material proveniente da matriz foi usado no processo de separação de grãos.

Os grãos de zircão da fração não magnética foram coletados manualmente e montados em epóxi com disposição aleatória, sem ordenação por morfologia, tamanho ou cor dos grãos. Após a montagem, a amostra foi lixada em três etapas: lixa 1200# (manual); lixa 2400# (4 minutos a 70 rpm na politriz); e por último a lixa 4000# (6 minutos a 60 rpm na politriz). Posteriormente, a amostra foi polida com pasta de diamante de granulometria $\frac{1}{4}$ μm (10 minutos a 60 rpm na politriz).

Cada amostra montada foi imageada por meio de microscópio eletrônico de varredura

(MEV) pelos métodos *backscattered electrons* (BSE) e *cathode-luminescence* (CL). As imagens foram usadas para identificar os grãos individualmente durante as análises.

As análises U-Pb e Lu-Hf foram realizadas via LA-MC-ICP-MS no Laboratório de Geocronologia da Universidade Federal de Ouro Preto (UFOP). Informações técnicas a respeito dos equipamentos, procedimentos de análise e correções encontram-se descritos em detalhe na seção de métodos do capítulo 3.

1.7 Gravimetria

A geofísica possui papel importante na presente pesquisa, todavia é usada como ferramenta periférica, de forma que não caracteriza o tema central da tese. A ferramenta geofísica selecionada para o estudo do tema proposto foi a gravimetria. Dados magnetométricos não foram usados, pois os levantamentos aeromagnetométricos disponíveis não abrangem toda a região estudada.

Foram usados dados de gravimetria terrestre em duas escalas: detalhe e ultra-detalhe.

Os dados de gravimetria terrestre de detalhe são provenientes da rede gravimétrica do IBGE e abrangem uma região mais ampla que a área de estudo, cobrindo parte dos estados de Goiás, Tocantins, Minas Gerais e o Distrito Federal (Figura 4). A aquisição dos dados foi realizada ao longo de estradas com espaçamento aproximado entre pontos de aquisição de 2.000 m. O detalhamento sobre as correções, tratamentos e métodos aplicados ao dado gravimétrico de detalhe encontram-se descritos na seção de métodos do Capítulo 4.

O levantamento gravimétrico terrestre de ultra-detalhe (levantamento próprio) foi realizado ao longo de quatro perfis de direção aproximada NE-SW cortando a Falha da Serra do Cristal. Os perfis possuem entre 30 e 40 km de extensão e o espaçamento entre os pontos de aquisição é de aproximadamente 500 metros em linha reta. Maiores detalhes sobre os equipamentos e procedimentos de aquisição dos dados gravimétricos terrestres, correções aplicadas aos dados, procedimentos de inversão e modelagem encontram-se descritos na seção de métodos do Capítulo 5.

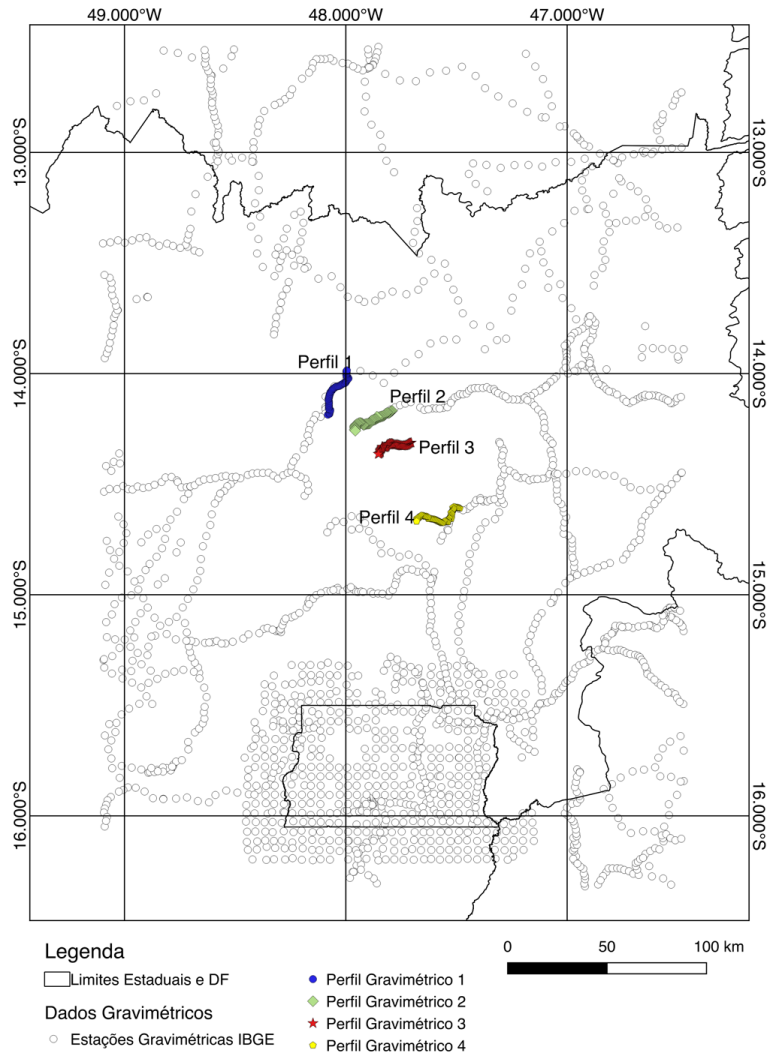


Figura 4 - Mapa da malha de dados gravimétricos de detalhe utilizados. Em mapa com limites estaduais estão plotadas as estações gravimétricas levantadas pelo IBGE e a localização dos 4 perfis de ultra detalhe adquiridos especificamente para este trabalho.

Referências Bibliográficas

- Alkmim, F.F., Martins-Neto, M.A., 2012. Proterozoic first-order sedimentary sequences of the São Francisco craton, eastern Brazil. *Marine and Petroleum Geology* 33, 127–139.
- Alkmim, F.F., Martins-Neto, M.A. 2001. A bacia intracratônica do São Francisco: arcabouço estrutural e cenários evolutivos. In: Pinto, C. P., Martins-Neto, M. A. (eds.) 2001. *Bacia do São Francisco, Geologia e Recursos Naturais*, Belo Horizonte: S.B.G.-MG, 2001. 349 p.9-30.
- Alkmim F.F., Brito Neves B.B., Alves J.A.C. 1993. Arcabouço tectônico do Cráton do São Francisco: uma revisão. In: J.M.L. Dominguez & A. Misi (eds). *O Cráton do São Francisco*. SBG/SGM/CNPq, Salvador, p. 45-62.
- Almeida F.F.M. de. 1968. Evolução tectônica do Centro-Oeste brasileiro no Proterozóico Superior. *An. Acad. Bras. Cienc.* 40 (supl.): 225-295.
- Alvarenga, C.J.S., Dardenne, M.A., Vieira, L.C., Martinho, C.T., Guimarães, E.M., Santos, R.V., Santana, R.O. 2012. Estratigrafia da Borda Ocidental da Bacia do São Francisco. *Boletim de*

- Geociências da Petrobrás 20, 145–164.
- Baeta, J.D.A.; Martins E.G., Leite E.A, Ramos, J.B.; SÁ, J.A.G.; Ribeiro, M.B.; Pires, P.R.O. 1977. Projeto Manganês no Centro-Sul de Goiás. Relatório Final, DNPM/CPRM. Escala 1:50.000. 273p.
- Botelho N.F. 1998. Cartografia do Projeto Cavalcante. Projeto Final de Graduação do Curso de Geologia. Universidade de Brasília. Instituto de Geociências.
- Campos, J.E.G, Dardenne M.A.; Martins-Ferreira, M.A.C. 2011. (inédito): Nota explicativa. Folha Alto Paraíso de Goiás 1:100.000 (PRONAGEO /CPRM).
- Campos, J.E.G. 2009. Estudo Integrado da Bacia Hidrográfica do Rio das Almas - EIBH. 28-49.
- Campos, J.E.G.; Dardenne, M.A.; Martins-Ferreira, M.A.C.; Freitas-Silva, F.H. 2013 Geologia do Grupo Paranoá na porção externa da Faixa Brasília. Revista Brasileira de Geociências. V.43(3): Páginas 461 a 476.
- Cloetingh, S.; Ziegler, P.A.; Beekman, F.; Burov, E.B.; Garcia-Castellanos, D.; Matenco, L. 2015. Schubert, Gerald (ed.), Treatise on Geophysics (Second Edition), pp. 513 - 592 Reference Module in Earth Systems and Environmental Sciences, pp. 513 – 592.
- Cloud, P.; Dardenne, M.A. 1973. Proterozoic age of the Bambuí Group in Brazil. Geol. Soc. Am. Bull., 84:1673-1676.
- Costa J.B.S., 1984. Aspectos litoestruturais e evolução crustal da região centro-norte de Goiás. Belém, IG-UFPA, Tese de Dout., 210p.
- Danderfer Filho, A., Lana, C.D.C., Júnior, H.N., & Costa, A.F.D.O. 2015. Constraints on the Statherian evolution of the intraplate rifting in a Paleo-Mesoproterozoic paleocontinent: New stratigraphic and geochronology record from the eastern São Francisco craton. Gondwana Research, 28(2), 668-688.
- Dardenne M.A. 1978. Zonação tectônica na borda ocidental do Cráton São Francisco. In: Congresso Brasileiro De Geologia, Recife, 30, Anais, p. 299-308.
- Dardenne M.A. 2000. The Brasília Fold Belt. In: U.G. Cordani, E.J. Milani, A. Thomaz Filho & D.A. Campos. 2000. Tectonic evolution of South America. 31 st International Geological Congress, Rio de Janeiro, SBG. p. 231-263.
- Dardenne M.A., Campos J.E.G., Alvarenga C.J.S. de, Martins F.A.L., Botelho N F. 1999. A sequência sedimentar do Grupo Araí na região da Chapada dos Veadeiros, Goiás. In: Simpósio de Geologia do Centro Oeste, 7, e Simpósio de Geologia de Minas Gerais, 10, Brasília., Atas, p. 100.
- Dardenne M.A.; Mello,S.M.G.; Moeri,E. 1971. Os estromatólitos do Grupo Bambuí: classificação, importância estratigráfica e metalogênica. Cong. Brás. Geologia,25, São Paulo, SBG, Bol.Res.Com. Esp., N° 1, p.88
- D'el-Rey Silva L.J.H, Oliveira I.L., Pohren, C.B., Tamizaki, M.L.N., Carneiro, R.C., Fernandes G.L.F., Aragão, P.E. 2011. Coeval perpendicular shortenings in the Brasilia belt: collision of irregular plate margins leading to oroclinal bending in the neoproterozoic of Central Brazil. Journal South Am. Earth Sciences 32: 1-13.
- Fairchild, T.R.; Schopf, J.W.; Shen-Miller, J.; Guimarães, E.M.; Edwards, M.D.; Langstein, A.; Li, X.; Pabst, M.; Melo-Filho, L.S. 1996. Recent discoveries of Proterozoic microfossils in south-central Brazil. Precambrian Research, 80: 125- 152.
- Faria A. De 1995. Estratigrafia e sistemas deposicionais do Grupo Paranoá nas áreas de Cristalina, Distrito Federal e São João da Aliança-Alto Paraíso de Goiás. Tese de doutorado, UnB - IG, 199 p.
- Fonseca M.A. 1996. Estilos estruturais e arcabouço tectônico do Segmento Setentrional da Faixa Brasília. Tese de doutorado, UnB-IG, 165p.

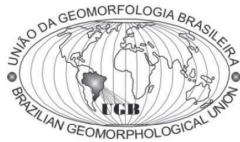
- Fonseca M.A., Uhlein A., Dardenne M.A. 1997. Evidências estruturais de evolução diacrônica entre os segmentos norte e sul da Faixa Brasília durante a orogênese Brasileira. In: SBG, Simpósio de Geologia do Centro-Oeste, 6, Atas, 93-95.
- Fonseca M.A., Dardenne M.A., Uhlein A. 1995. Faixa Brasília Setor Setentrional: Estilos Estruturais e Arcabouço Tectônico. Revista Brasileira de Geociências, V. 25(4), Páginas 267 a 278.
- Fuck, R.A., Dantas E.L., Pimentel, M.M. 2001. Nd isotopes, U-Pb single grain and SHRIMP zircon ages from basement rocks of the Tocantins Province. In: Symposium on Isotope geology, 3, 2001, Pucón, Chile. Proceedings... Chile, 2001. p.311-313
- Fuck R.A.; Pimentel, M.M.; D'el-Rey Silva L.J. 1994. Compartimentação Tectônica na porção oriental da Província Tocantins. In: Congresso Brasileiro de Geologia, 38, Camboriú (SC), Anais, p.215-216.
- Fuck R.A., Marini J.O., Dardenne M.A. & Figueiredo A.N. 1988. Coberturas metassedimentares do Proterozóico Médio: os grupos Araí e Paranoá na região de Niquelândia - Colinas, Goiás. Revista Brasileira de Geociências. 18(1):54-62.
- Fuck, R.A. & Marini, O.J. - 1981 - O Grupo Araxá e Unidades Homotaxiais. In: Simp. Sobre o Craton do São Francisco e suas Faixas Marginais. Salvador, 1981. Anais... Salvador, Sbg/Ba. P. 118-130.
- Guadagnin, F.; Chemale, F. 2015. Detrital zircon record of the Paleoproterozoic to Mesoproterozoic cratonic basins in the São Francisco Craton, Journal of South American Earth Sciences, Volume 60, July 2015, Pages 104-116.
- Guadagnin, F., Chemale Jr., F., Magalhães, J., Santana, A., Dussin, I., Takehara, L., 2015. Age constraints on crystal-tuff from the Espinhaço supergroup - insight into the Paleoproterozoic to Mesoproterozoic intracratonic basin cycles of the Congo-São Francisco Craton. Gondwana Research 27, 363-376.
- Hasui, Y.; Almeida, F.F.M. de. 1970. Geocronologia do Centro-Oeste brasileiro. BoI. Soc. Bras. Geol. 19:5-26, São Paulo.
- Marini, J.O.; Fuck, R.A.; Danni, J.C.M.; Dardenne, M.A.; Loguercio, S.O.C.; Ramalho, R. 1984. As faixas de dobramentos Brasília, Uruaçu e Paraguai-Araguaia e o Maciço Mediano de Goiás. In: Schobbenhaus, C.; Diógenes, A.C.; Derze, G.R. Asmos, M.E (Coords.). Geologia do Brasil. Brasília. DNPM. P. 251-303.
- Marini O.J., Fuck R.A., Danni J.C., Dardenne M.A. 1981. A evolução geotectônica da Faixa Brasília e do seu embasamento. In: Simpósio Sobre o Cráton do São Francisco e suas Faixas Marginais. Salvador, SBG, Anais, p.100-115.
- Martins F.A.L. 1999. Análise faciológica e estratigráfica do Paleo-Mesoproterozoico: Sequência Araí no Parque Nacional da Chapada dos Veadeiros, Goiás. Dissertação de mestrado, Instituto de Geociências, Universidade de Brasília, p. 137.
- Martins-Neto M. A., Alkmim F. F. 2001. Estratigrafia e Evolução Tectônica das Bacias Neoproterozóicas do Paleocontinente São Francisco e suas Margens: Registro da Quebra de Rodínia e Colagem de Gondwana. In: Pinto C. P., Martins- Neto M. A., (Eds.), Bacia do São Francisco: Geologia e Recursos Naturais. Belo Horizonte (SBG/ MG), p.31-54.
- Matteini M., Dantas E.L., Pimentel M.M., Alvarenga C.J.S. & Dardenne M.A. 2012. U-Pb and Hf isotope study on detrital zircons from the Paranoá Group, Brasília Belt Brazil: constraints on depositional age at Mesoproterozoic-Neoproterozoic transition and tectono-magmatic events in the São Francisco Craton:168-181.
- Milanovsky, E.E. 1978. Some problems of rifting development in the Earth's history. Tectonics and Geophysics of Continental Rifts: Volume Two of the Proceedings of the NATO Advanced Study

- Institute Paleorift Systems with Emphasis on the Permian Oslo Rift, held in Oslo, Norway, July 27–August 5, 1977, 37, 385.
- Pimentel M.M. 2016. The tectonic evolution of the Neoproterozoic Brasília Belt, central Brazil: a geochronological and isotopic approach. *Brazilian Journal of Geology*, 46(Suppl 1): 67-82. DOI: 10.1590/2317-4889201620150004
- Pimentel M.M., Rodrigues J.B., Dellagiustina M.E.S., Junges S.L., Matteini M., Armstrong R. 2011. The tectonic evolution of the Neoproterozoic Brasília Belt, central Brazil, based on SHRIMP and LA-ICPMS UePb sedimentary provenance data: A review. *Journal of South American Earth-Sciences*, 31: 345-357.
- Pimentel, M.M.; Jost, H.; Fuck, R.A. 2004. O Embasamento da Faixa Brasília e o arco magmático de Goiás. In: Mantesso-Neto et al (eds) *Geologia do Continente Sul-Americano. Evolução da obra de Fernando Flávio Marques de Almeida*. Becca, São Paulo, pp 356–381
- Reis, H.L.S.; Alkmim, F.F. 2015 Anatomy of a basin-controlled foreland fold-thrust belt curve: The Três Marias salient, São Francisco basin, Brazil, *Marine and Petroleum Geology*, Volume 66, Part 4, September 2015, Pages 711-731.
- Sutton, J., Watson, J.V. 1986. Architecture of the continental lithosphere. *Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences*, 5-12.
- Tanizaki, M.L.N.; Campos, J.E.G; Dardenne, M.A. 2015. Estratigrafia do Grupo Araí: registro de rifteamento paleoproterozoico no Brasil Central. *Brazilian Journal of Geology*, 45(1): 95-108.
- Uhlein, A., Fonseca, M.A., Serr, H.J., Dardenne, M.A. 2012. Tectônica da Faixa de Dobramentos Brasília – setores setentrional e meridional. *Geonomos* 20, n.2: 1-14.
- Valeriano C.M., Dardenne M.A., Fonseca M.A., Simões L.S.A., Seer H.J. 2004. A evolução tectônica da Faixa Brasília. In: Mantesso-Neto V., Bartorelli A., Carneiro C. D.R., Brito-Neves B.B (Eds.) *Geologia do Continente Sul-Americano: Evolução da Obra de Fernando Flávio Marques De Almeida*, São Paulo, 575-592. Ed. Beca.
- Valeriano C.M., Pimentel M.M., Heilbron M., Almeida J.C.H., Trouw R.A.J. 2008. Tectonic evolution of the Brasília belt, central Brazil, and early assembly of Gondwana. In: Pankhurst R.J., Trouw R.A.J., BritoNeves B.B., De Wit M.J. (Eds.), *West Gondwana: Pre-cenozoic Correlations across the South Atlantic Region*. Geological Society, London, Special Publications, 294: 197-210.

CAPÍTULO 2

Artigo 1:

Compartimentação geomorfológica como suporte para estudos de evolução geotectônica: aplicação à região da Chapada dos Veadeiros, GO

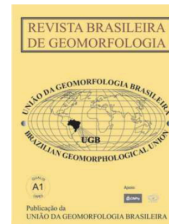


www.ugb.org.br
ISSN 2236-5664

Revista Brasileira de Geomorfologia

v. 18, nº 3 (2017)

<http://dx.doi.org/10.20502/rbg.v18i3.1119>



Compartimentação geomorfológica como suporte para estudos de evolução geotectônica: aplicação à região da Chapada dos Veadeiros, GO

Marco Antonio Caçador Martins-Ferreira¹; José Eloi Guimarães Campos¹

1. Instituto de Geociências, Universidade de Brasília, Campus Universitário Darcy Ribeiro, Asa Norte, 70910900, Brasília, DF, Brasil.

Abstract

Geomorphological analysis results aided by geological information are used as parameters for the geomorphological compartmentation of an orogenic region as a methodological proposal for preliminary analysis to be performed prior to tectonic evolution studies. The method is applied to the Chapada dos Veadeiros region, estate of Goiás, Brazil, that underwent the Neoproterozoic Brasiliano Orogeny. The tectonic relief control is sufficiently important in the area as to allow tectonic features to be the main criteria used in geomorphological analysis. The main lineaments were extracted and categorized according to the type of structure responsible for their generation. The morphostructural lineament map was analyzed according to their cutting relationships. Besides the geotectonic criteria, geomorphological compartmentation also considered hypsometry, relief patterns and shapes, drainage patterns and density and the prevailing soil classes. The study was conducted at a 1:250.000 scale and resulted in the distinction of 16 geomorphological compartments divided into 4 categories: dissected plateaus, valleys, valleys with residuals and escarpments. The compartments were named according to important local geographic features as hills, rivers or cities and were analyzed for their internal characteristics as well as their boundaries to the adjacent compartments. When possible, relative estimates were inferred for the temporal relationship between structures, strain intensity, transport extent and direction, the nature of shear zones, among other structural aspects. The method was successful in delivering a better understanding of the possible structural domains and the relative temporality of events or tectonic regime phases that affected the area.

Keywords: Geomorphological compartmentation, Tectonic relief control, Chapada dos Veadeiros.

Resumo

Produtos de análises geomorfológicas aliados a informações geológicas são usados como parâmetros para a compartimentação geomorfológica de região orogênica como proposta metodológica de análise prévia aos estudos de evolução tectônica regional. O método é aplicado na região da Chapada dos Veadeiros, GO que foi submetida à orogenia Brasileira no Neoproterozoico. O controle geotectônico do relevo na região estudada é suficientemente marcante a ponto de permitir que este seja um dos principais critérios de análise geomorfológica na área. Os lineamentos principais foram extraídos e classificados de acordo com o tipo de estrutura geradora (tabela 1). O mapa de lineamentos morfoestruturais foi analisado de acordo com suas relações de corte. Para a proposição da compartimentação geomorfológica da região, além do critério geotectônico, foram considerados a hipsometria, o padrão de relevo, formas regionais de relevo, padrão e densidade da drenagem e classes de solos predominantes. O estudo foi realizado em escala 1:250.000 e resultou na distinção de 16 compartimentos geomorfológicos enquadrados nas categorias: planaltos dissecados, vales, vales com residuais e escarpas. Os compartimentos foram nomeados de acordo com feições geográficas importantes, como rios, serras ou cidades e foram analisados de acordo com suas características internas bem como seus limites com cada compartimento adjacente. Quando possível, foram estimadas a relação temporal entre as estruturas, a intensidade da deformação, a extensão relativa e direção do transporte tectônico, a natureza das zonas de cisalhamento e demais aspectos estruturais. O método foi bem-sucedido no sentido de proporcionar o avanço do conhecimento quanto aos possíveis domínios estruturais e a temporalidade relativa dos eventos ou das diferentes fases de regimes tectônicos que afetaram a área.

Palavras-Chave: Compartimentação geomorfológica, Controle geotectônico do relevo, Chapada dos Veadeiros

1. Introdução

A análise geomorfológica é ferramenta comum nos meios não acadêmicos e tem sido praticada com finalidades diversas, com destaque para: subsídio ao planejamento de uso agrícola ou urbano; a gestão de recursos naturais de uma região; a avaliação do potencial hídrico para o abastecimento: geração de energia hidrelétrica; projetos de irrigação e outros. O mapa de compartimentação geomorfológica é um plano de informação praticamente obrigatório em

qualquer modalidade de estudo ambiental.

Todavia, grande parte dos estudos geomorfológicos de caráter científico possuem caráter puramente descritivo, sem o intuito de gerar dados aplicáveis às demais áreas de pesquisa das geociências. Segundo Saadi (2013), a maior parte das publicações científicas em geomorfologia deixa a impressão de que se trata apenas de estudo do aspecto da superfície terrestre e, eventualmente, da sua evolução no tempo geológico, como exercício meramente acadêmico, porém muitas vezes sem projeções prático-aplicativas relevantes para o desenvolvimento de outras áreas da ciência.

É consenso que os diversos fatores que influenciam na modelagem do relevo de uma região e podem ser classificados em duas categorias: exógenos e endógenos. Os fatores exógenos são comandados pelo clima (chuva, vento, gelo, temperatura, vegetação, etc.), já os endógenos pela estrutura geológica (dobras, falhas, juntas, foliações, acamamento, etc.) pelo tipo de rocha e pelo ambiente tectônico (subsidência, soerguimento, existência de processos orogênicos ou epirogênicos, formação de domos ou bacias, dentre outros).

Os estudos de compartimentação geomorfológica são uma vertente da análise geomorfológica e buscam individualizar áreas com características comuns baseados em critérios físicos superficiais, elegidos de acordo com a escala de trabalho e os objetivos propostos. Os principais parâmetros adotados pelos estudos de compartimentação geomorfológica incluem dados de hipsometria, declividade, padrão de relevo, forma de relevo, densidade e forma da drenagem, tipos de solos e extração de lineamentos (Souza Filho 1999; Cordeiro *et al.* 2010; Gomes & Vital 2010; Costa & Falcão 2011; Tavares *et al.* 2014; Marent & Valadão 2015).

A extração de lineamentos, todavia, é quase sempre usada sem atenção para a distinção dos tipos de lineamentos extraídos, no que se refere aos diferentes tipos de estruturas geológicas que os formam. É comum que os lineamentos sejam categorizados apenas por sua direção e não por aspectos mais específicos como comprimento, curvatura, padrão de repetição, etc., propriedades que podem revelar o tipo de estrutura geológica geradora dos lineamentos mapeados. A ausência de distinção de características de lineamentos acarreta perda de informações extremamente relevantes, bem como o empobrecimento das possibilidades de aplicação do estudo.

Em trabalho de pesquisa bibliográfica específico sobre a definição dos termos lineamento e lineação em geociências, O'leary *et al.*, (1976) concluem que a definição mais precisa do conceito de lineamento é "*uma feição linear mapeável [na escala de trabalho], simples ou composta, cujas partes estão alinhadas de forma reta ou curvilínea, que seja expressivamente*

distinta de feições adjacentes e que possivelmente reflita algum fenômeno em sub-superfície". Esta definição é adotada no presente trabalho e desenvolvida ao ponto de permitir a distinção entre os diferentes tipos de lineamentos e suas relações genéticas com as estruturas geológicas.

Trabalhos de mapeamento geológico-estrutural permitem inferir que os lineamentos compostos de segmentos curtos e retilíneos formando padrão repetitivo com maior ou menor grau de regularidade, são comumente o reflexo de estruturas como planos de acamamento, fraturas e foliações e geralmente não imprimem feições de relevo significativas, podendo ser melhor identificadas em escalas de detalhe. Já os lineamentos longos e contínuos, ora curvilíneos ora retilíneos, geralmente representam feições regionais como contatos litológicos, eixos ou flancos de dobras e principalmente falhas em regimes tanto compressivos quanto distensivos e transcorrentes e frequentemente são marcadas por feições de relevo óbvias na paisagem.

As estruturas regionais de qualquer regime tectônico exercem influência marcante no padrão de dissecação do relevo e muito frequentemente separam áreas com características geológicas, hidrológicas, pedológicas e conseqüentemente geomorfológicas distintas. De acordo com essa perspectiva, os lineamentos contínuos podem auxiliar no traçado de divisores entre compartimentos geomorfológicos.

Especificamente com relação aos controles geotectônicos das feições geomorfológicas, embora exista uma correlação direta, poucos são os trabalhos que versam diretamente sobre a influência das estruturas geológicas na evolução do modelado terrestre.

Gilbert (1877) em trabalho pioneiro sobre a evolução do relevo pelo equilíbrio dinâmico menciona o papel da resistência e estruturação tectônica das rochas no controle da dissecação do relevo. Para Gilbert as formas de relevo e os depósitos superficiais possuem uma íntima relação com a estrutura geológica. Davis (1899) e Penck (1924), apesar de antagonistas, elaboram modelos que relacionam taxa de soerguimento com taxa de denudação, atribuindo diretamente aos eventos tectônicos o papel de agentes controladores da morfogênese. Já Hack (1960) considera o relevo como o produto de uma competição entre a resistência dos materiais crustais e o potencial das forças de denudação.

Tricart (1968), Morisawa (1975) e Ollier (1981) retomam esses princípios clássicos e os levam adiante em seus estudos. Tricart (1968) entende que a superfície terrestre se estende até a litosfera e sua forma é resultado da interação entre forças endógenas e exógenas, sendo o papel da geomorfologia localizar o estado atual das coisas no quadro do tempo dos mecanismos das forças antagônicas que modelam o relevo. Morisawa (1975) considera a tectônica como o agente fundamental no controle da morfogênese. Ollier (1981) afirma que a compreensão das formas de

relevo de todas as ordens de grandeza passa pelo entendimento da nova geologia e da tectônica. Christofolletti (1980) define a Geomorfologia como a ciência que estuda as formas do relevo, bem com sua gênese, características morfológicas, bases litológicas e os processos modeladores e controladores.

Conforme relatado por Saadi (2013), os resultados obtidos em estudos nos estados de Minas Gerais (Saadi 1991), Ceará (Saadi & Torquato 1992) e Rio Grande do Norte (Bezerra *et al.* 1993), indicam que a compartimentação do relevo é, praticamente em todas as escalas, resultado de controles tectônicos exercidos pelos diversos tipos de descontinuidades crustais. Clark & Small (1982) inovam ao incluir aspectos geológicos específicos como responsáveis pelo controle da ação dos agentes exógenos na evolução do relevo.

Apesar do amplo reconhecimento do papel da tectônica para a morfogênese, são poucos os trabalhos de geomorfologia com aplicação direta para a análise tectônica. Estes geralmente abordam aspectos morfoestruturais e morfogenéticos aplicados à neotectônica (Sousa Filho *et al.* 1999; Marques Neto e Perez Filho 2013; Tavares *et al.* 2014; Alves e Rosseti 2015) e não propõem metodologia voltada para a análise tectônica de eventos prévios. Ab'saber (1998) pondera que “...para o estudo paleogeográfico dos diversos lapsos de tempo geológico que respondem pela gênese e evolução das terras brasileiras, torna-se imprescindível a aplicação de métodos específicos de trabalho, utilizados segundo um rumo e uma combinação de técnicas inteiramente distintas para cada um dos casos que se tenha em vista”.

Com base nas premissas anteriormente apresentadas, o objetivo do presente estudo é propor um método de compartimentação geomorfológica que possa contribuir, como estudo preliminar, para a análise e interpretação da evolução tectônica de regiões submetidas a eventos deformacionais. Dessa forma, o estudo propõe a análise do relevo no espaço geográfico como um dos subsídios para a análise tectônica em escala regional (menor que 1:250.000), escala na qual o método tem sido aplicado pelos autores com sucesso.

2. Área de Estudo

A área escolhida para o desenvolvimento e aplicação do método (Figura 1) está localizada na região da Chapada dos Veadeiros, situada 200 km a norte da capital Brasília, em região que sofreu processo orogenético no Neoproterozoico (1.000 a 540 Ma.). Dardenne & Campos (2001) sugerem que a evolução geomorfológica da região está intimamente associada ao substrato rochoso e à estruturação tectônica. A área apresenta feições estruturais marcantes que atuam como fortes controles de relevo. Além disso, o contraste litológico entre as rochas

expostas do embasamento, das bacias sedimentares e das rochas intrusivas, também induz à erosão diferencial.

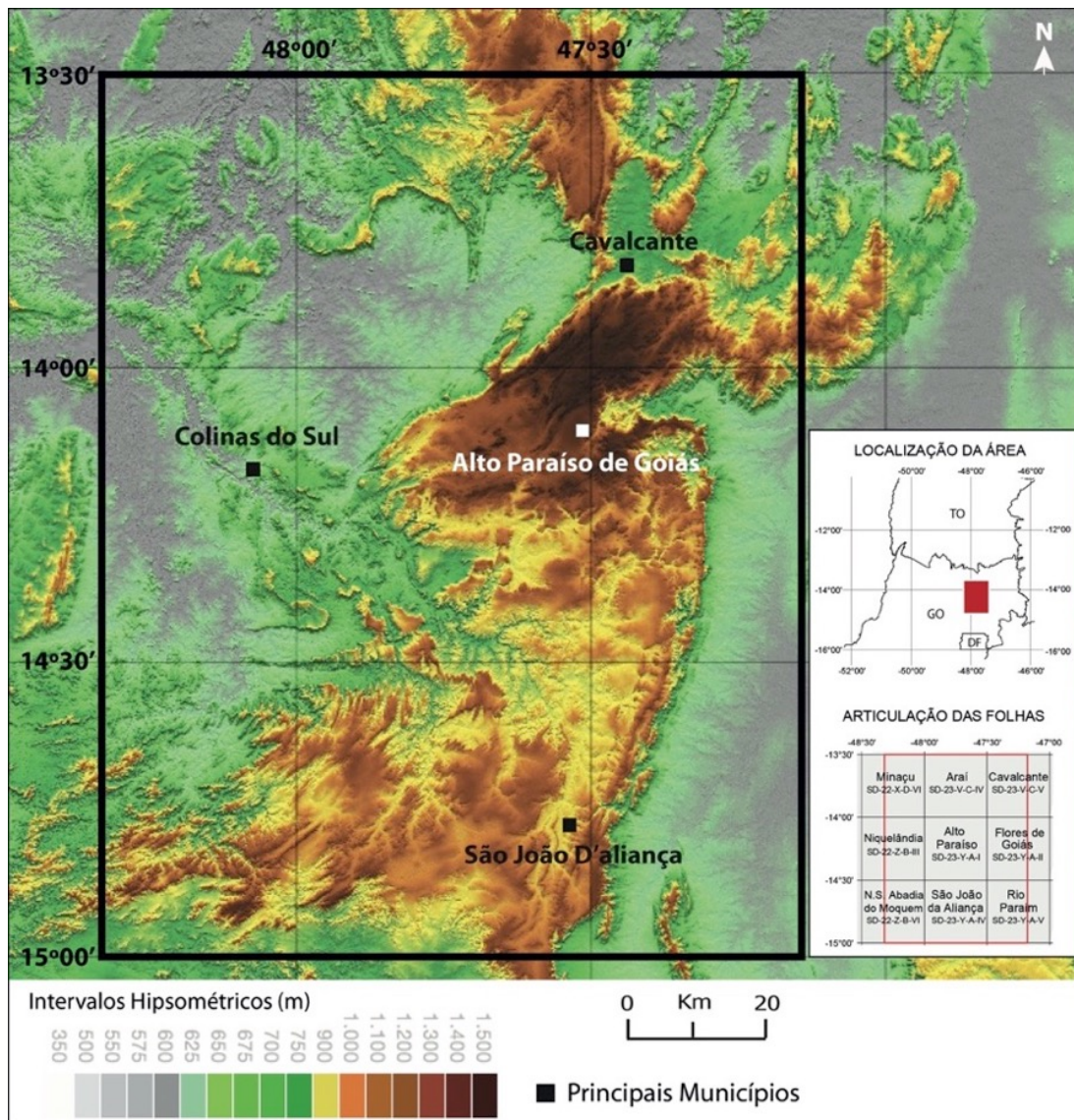


Figura 1 - Localização da área no estado de Goiás, articulação das folhas cartográficas e principais municípios sobre mapa hipsométrico da área de estudo. Exemplo de mapa secundário ou derivado, aplicado à compartimentação geomorfológica.

Em regiões deformadas, a erosão diferencial ocorre não só por diferenças de resistência física e química dessas rochas, mas também pelos diferentes estilos deformacionais impressos em cada uma delas, já que o comportamento reológico de cada rocha é distinto. Deve-se considerar ainda a profundidade a que essas rochas, hoje expostas, estavam submetidas durante a deformação, que implicava em variadas condições de temperatura e pressão confinante, o que certamente teve papel marcante no estilo deformacional impresso.

Além do estilo deformacional e do metamorfismo, a posição das estruturas e o tempo de atuação dos processos denudacionais também são fundamentais para a análise da erosão

diferencial. Como exemplo pode-se citar o caso da evolução do relevo de regiões compostas por rochas sedimentares ou metassedimentares dobradas que passa do tipo Jurássico, no início de evolução, até apalachiano, após amplo período sob denudação.

O controle geotectônico do relevo na região estudada é suficientemente marcante a ponto de permitir que as feições e estilos estruturais sejam os principais critérios de análise geomorfológica na área. Apesar da importância dos critérios geotectônicos, mostra-se fundamental o uso de critérios tradicionais como drenagem, solo e altitude, que atuam como guias para a identificação de padrões regionais onde as informações geológico-estruturais são escassas.

3. Contexto Geológico Regional

A área de estudo está localizada na porção centro-norte da zona externa da Faixa de Dobramentos e Cavalgamentos Brasília (Dardenne 2000), que por sua vez está inserida na porção oriental da Província Estrutural Tocantins, gerada pela orogênese Brasileira no Neoproterozoico (Pimentel *et al.* 1999) em regime de dobras e falhas de empurrão em escamas de descolamento, ora rasas ora profundas, com maior ou menor participação do embasamento.

A Faixa Brasília (Figura 2) recobre o nordeste e sudeste de Goiás, o sudeste do Estado de Tocantins, o Distrito Federal e a porção ocidental de Minas Gerais e é dividida em dois segmentos, de acordo com sua estruturação: setentrional e meridional, separados por uma estrutura regional denominada Inflexão dos Pirineus. Essa divisão foi proposta devido ao fato destes segmentos possuírem peculiaridades estratigráficas, estruturais, metamórficas e metalogénicas, proporcionando uma tectônica complexa e distinta nas porções norte e sul (Fonseca 1995, Freitas-Silva & Campos 1998).

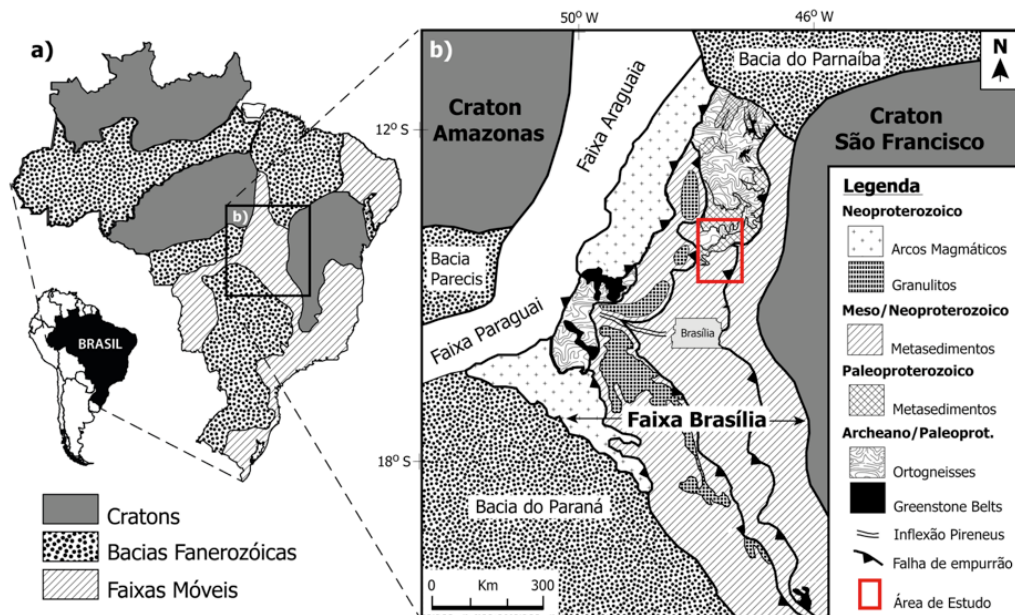


Figura 2 - Localização da área de estudo no mapa geológico da Faixa de Dobramentos e Cavalgamentos Brasília.

4. Materiais e Métodos

Tricart (1965) ao tratar da concepção e princípios de realização da Carta Geomorfológica ressalta as diferentes categorias de fenômenos representados segundo a escala adotada. Como exemplo, as cartas em pequena escala, como 1:1.000.000 ou 1:500.000, se orientam essencialmente para os fenômenos morfoestruturais, mostrando as anticlinais resultantes de dobramentos, seus *monts* ou *combes*, ou ainda os horsts e os grabens de processos de falhamentos associados a eventos tafrogenéticos.

A escala de análise usada na compartimentação proposta é de 1:250.000, considerada a escala ideal para a aplicação do método em regiões deformadas, dada a capacidade de identificação e diferenciação de estruturas tectônicas tanto semi-regionais como regionais. Estima-se que em escalas menores que 1:250.000 a contribuição da compartimentação geomorfológica em estudos de análise tectônica pode ser eficaz, podendo ser usada sem limites até as escalas globais. Deve-se considerar, contudo, que quanto menor a escala de trabalho, menor será a capacidade de detalhamento de ambientes e processos tectônicos. Por outro lado, mapas de compartimentação geomorfológica com escalas maiores ou iguais a 1:100.000 devem ser prioritariamente utilizados para as aplicações práticas da geomorfologia, com destaque para análise ambiental e avaliação de usos prioritários do território.

As fontes de dados selecionadas para o presente trabalho abrangem a área das seguintes quadrículas em escala 1:100.000: Minaçu (SD.22-X-D-VI), Araí (SD.23-V-C-IV), Cavalcante (SD.23-V-C-V), Niquelândia (SD.22-Z-B-III), Alto Paraíso (SD.23-Y-A-I), Flores de Goiás

(SD.23.Y-A-II), Nossa Senhora da Abadia do Moquém (SD.22-Z-B-VI), São João da Aliança (SD.23-Y-A-IV) e Rio Paraim (SD.23-Y-A-V) (Figura 1). Foram usados os dados SRTM 1-arco (resolução de 30 metros), imagens ALOS Palsar, Landsat 8, além de mapas derivados - hipsométrico, padrão de relevo, densidade de drenagem, forma da drenagem e de solos (Figura 3). Esses produtos foram usados em conjunto afim de balizar o mapeamento das feições de relevo já que determinados produtos salientam melhor algumas feições em detrimento de outras.

Demek (1976, apud Casseti, 1990) propõe método para o mapeamento geomorfológico onde frisa a importância do uso de informações geológicas detalhadas. O método proposto por Demek se assemelha em alguns pontos ao método aqui proposto e por isso é apresentado a seguir, todavia não vai adiante na interpretação de relações temporais e deslocamentos relativos, já que não compartilha dos objetivos do presente trabalho em usar a compartimentação geomorfológica como subsídio para a análise da evolução tectônica de uma área. Demek (1976) propõe os seguintes passos para o mapeamento de morfoestruturas:

- a)** análise das cartas geológicas e tectônicas de áreas em estudo (em escalas pequenas e grandes), com a transferência dos principais falhamentos para uma determinada base;
- b)** análise de cartas topográficas, em iguais escalas, com o objetivo de se elaborar uma carta das rupturas tectônicas e das formas de relevo lineares, e uma carta dos elementos do relevo segundo seus atributos morfográficos e morfométricos;
- c)** elaboração de perfis geológico-geomorfológicos, com a intenção de se definirem níveis regionais e elaboração de uma estratigrafia das formas;
- d)** interpretação de fotografias aéreas procurando especificar a gênese dos elementos do relevo;
- e)** levantamento de campo para teste e correção das interpretações, valorizando-se caminhamentos previamente definidos e utilizando-se, eventualmente, de sobrevoos no caso de áreas de difícil acesso. Nesta fase pode-se incluir coleta de materiais para posterior análise em laboratório;
- f)** integração da informação obtida em campo. A carta das formas de relevo resultante, considerando seus aspectos morfográficos e morfométricos, é revista, assumindo um caráter genético, dada a existência de elementos importantes para explicar a origem das formas e esculturação do modelado.

Para fins de análise qualitativa, o presente estudo faz uso da integração de produtos cartográficos primários e secundários (derivados) e de imagens orbitais. Os critérios usados para a definição de compartimentos geomorfológicos foram a hipsometria, textura, padrão de relevo, formas regionais de relevo, padrão e densidade da drenagem, classes de solos predominantes,

além de lineamentos contínuos e compostos.

A extração de lineamentos foi amparada por produtos SRTM em escala de cinza, bem como coloridos (resolução de 30 metros), além de composições ternárias de cenas Landsat 8. Neste método, a textura do DEM reflete os menores elementos visualmente distinguíveis na imagem (falhas, flancos e eixos de dobras, superfícies de cume, vales e padrões de drenagem). Variações nos padrões de relevo e drenagem também são atributos básicos para a análise de imagem. Os elementos de textura podem ter distribuições aleatórias ou ordenadas, dependendo da composição e estruturação das rochas subjacentes às formas de relevo. Os elementos de textura podem assim ser agrupados para formar unidades morfoestruturais (Santos et al. 2009).

Os lineamentos extraídos foram diferenciados pelo tipo de estrutura que representam e o regime tectônico no qual estão inseridos já que objetivou discriminar seus diferentes padrões genéticos incluindo sistemas de cisalhamento transcorrentes, sistemas de empurrões de muito baixo ângulo, eixos de dobras, intrusões aflorantes e sub-aflorantes, foliações e acamamentos. A técnica foi balizada pelos dados geológicos regionais e locais levantados em campo e conhecidos da literatura.

A partir dos diferentes critérios de extração de estruturas lineares mencionados, foi gerado um mapa de lineamentos sobre imagem SRTM exibindo os diversos tipos de lineamentos extraídos (Figura 4). O mapa de lineamentos foi amplamente usado para definir os limites dos compartimentos, que coincidem com formas encaixadas de drenagem, limites de domínios pedológicos e litológicos.

Os compartimentos geomorfológicos gerados (Figura 6) foram posteriormente comparados com dados geológico-estruturais da área, a fim de aferir a eficácia do método em identificar compartimentos com diferentes propriedades litológicas e deformacionais que possam ser úteis para a análise da evolução tectônica da região.

Conforme descrito na seção de discussões, os produtos gerados foram interpretados quanto à temporalidade das estruturas que controlam o relevo por meio de suas relações de corte. Os compartimentos geomorfológicos foram interpretados de acordo com o padrão deformacional interno e limítrofe afim de buscar indícios de deslocamento relativo bem como a caracterização da deformação que controla o relevo em diferentes níveis crustais.

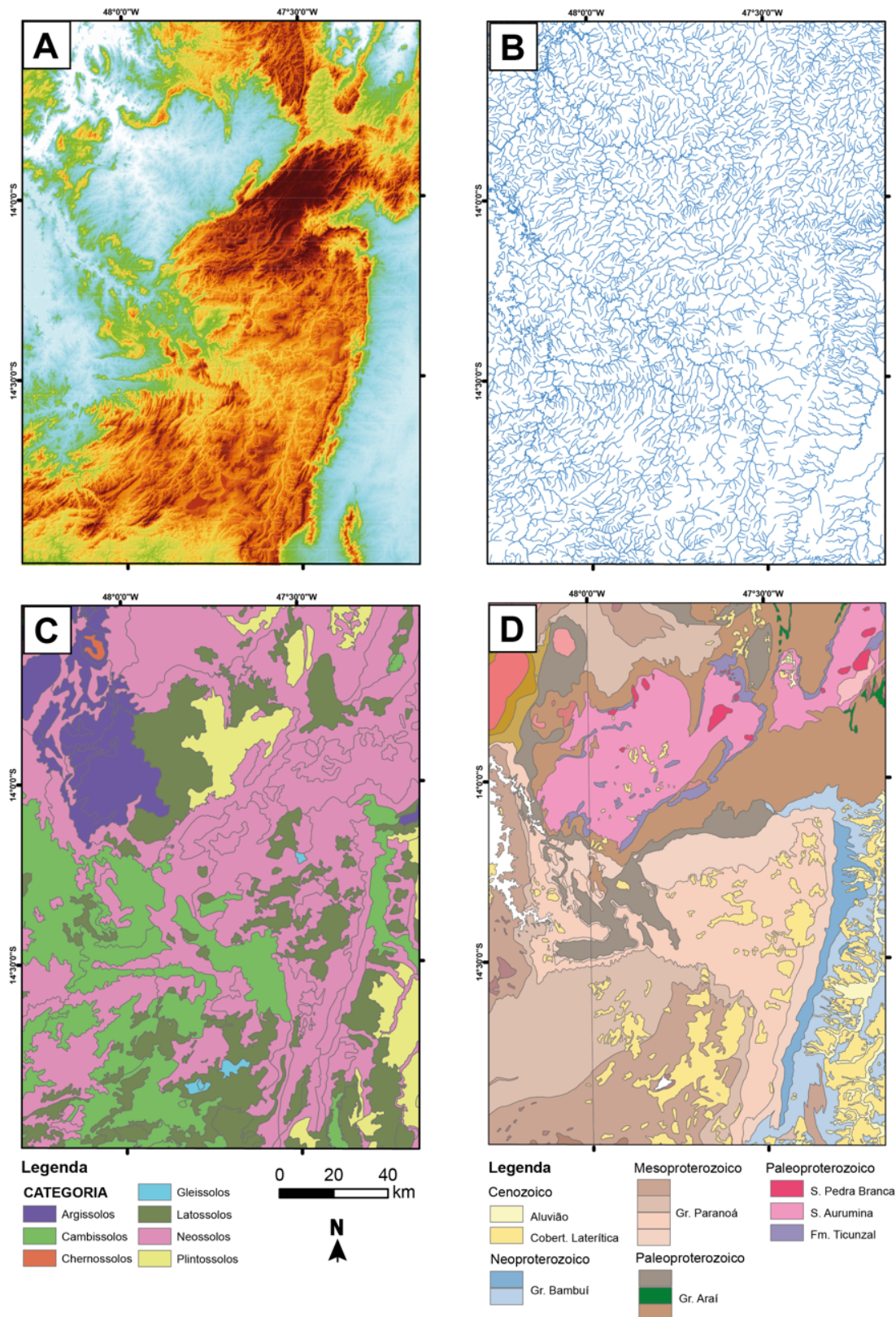


Figura 3 - Mapas (A) hipsométrico, (B) de rede de drenagem, (C) pedológico e (D) geológico da área (fontes: USGS, IBGE, Embrapa, CPRM, respectivamente). Mapas temáticos utilizados como critérios para a compartimentação geomorfológica.

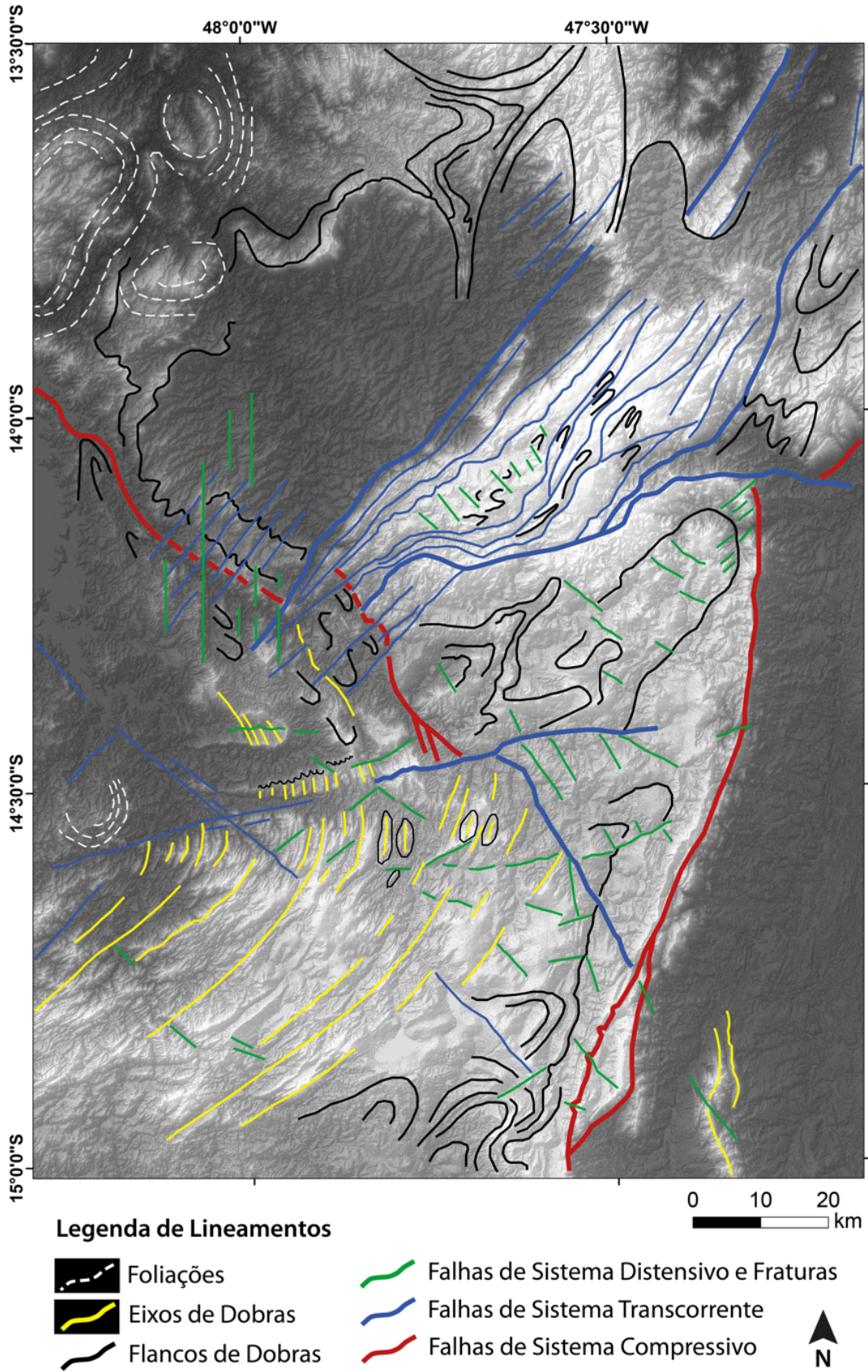


Figura 4 - Mapa de lineamentos morfoestruturais sobre SRTM 1-arco em escala de cinza sombreado.

5. Resultados

5.1. Interpretação dos Lineamentos Morfoestruturais

A forma como as estruturas estudadas ficam impressas na paisagem foi analisada com o objetivo de se estabelecer padrões entre o tipo de estrutura/regime tectônico e o seu comportamento frente aos agentes de dissecação do relevo na área estudada e estão sumarizadas na tabela 1. Os nomes das feições de relevo citadas estão indicados na Figura 5 e foram extraídos das cartas topográficas disponibilizadas no banco de dados geográficos do Exército Brasileiro. Os nomes dos compartimentos geomorfológicos citados estão indicados na Figura 6.

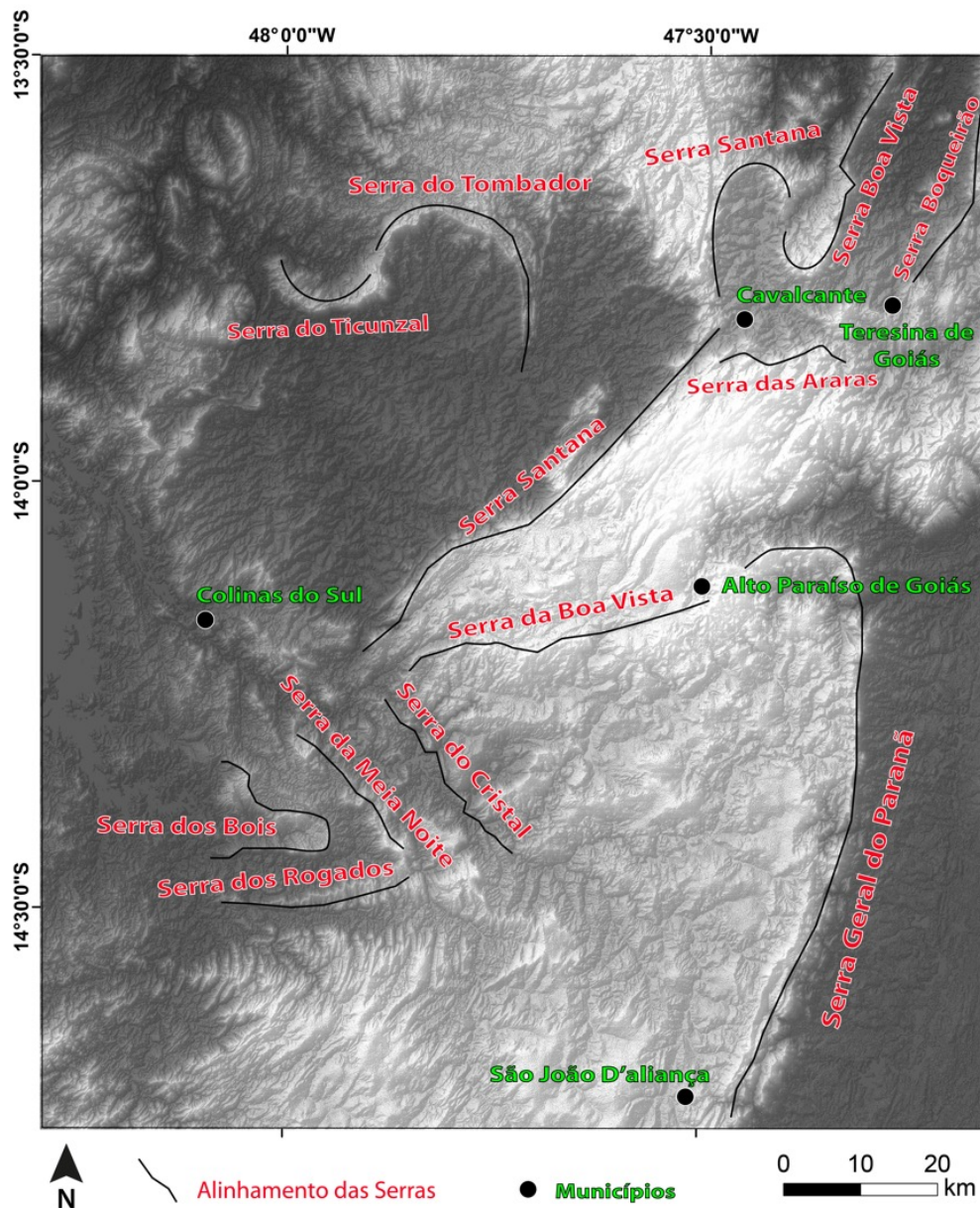


Figura 5 - Mapa das principais feições de relevo da área de estudo sobre produto SRTM sombreado.

As falhas de regime compressivo (reversas e empurrões) ocorrem controlando o curso dos rios, ora hospedando o rio ao longo de seu traço principal (por exemplo, Falha do Rio Paranã, Falha do Rio Maranhão), ora limitando lateralmente a calha de um rio (por exemplo, Falha da Serra dos Cristais e o rio Tocantinzinho). Essas falhas geram escarpas pronunciadas e de grade continuidade lateral (exemplo: Serra do Paranã, Serra dos Cristais) e imprimem lineamentos longos e contínuos, são pouco sinuosos em falhas de mais alto ângulo como falhas reversas ou rampas de empurrão e muito sinuosos em falhas de baixo ângulo como frentes de empurrões ou patamares.

As estruturas transcorrentes, ou zonas de cisalhamento (ZC) também podem hospedar cursos d'água (exemplo, Ribeirão São Miguel na ZC da Serra da Boa Vista ou ZC Colinas-Alto Paraíso), todavia exercendo menor controle sobre as drenagens, dando liberdade para que os rios não se encaixem nos seus traços principais com muita precisão ou não se mantenham neles por longos trechos. Essas estruturas geram escarpas de alta declividade (exemplo, Serra de Santana, Serra da Boa Vista, Serra do Boqueirão, e outras) e imprimem lineamentos longos, contínuos e retilíneos podendo ser localmente sinuosos.

As estruturas rúpteis tardias de regime distensivo (falhas normais) ocorrem como traços retilíneos ou quase retilíneos e descontínuos, podem ter direções preferenciais coincidentes em todos os domínios mas podem também ocorrer com maior frequência em domínios onde afloram níveis crustais mais rasos, ocorre repetição paralela de direção similar (exemplo, Grupo Paranoá ao centro e a sul da área estudada).

As estruturas que representam flancos de dobras em quartzitos do Grupo Araí proporcionam erosão diferencial preferencialmente ao longo dos eixos das antiformes, gerando escarpas de serras curvas obedecendo a curvatura do padrão de dobramento (exemplo, Serra do Ticunzal, Serra do Tombador, Serra Santana de Cavalcante, Serra Boa Vista de Cavalcante).

As estruturas que representam eixos de dobras ou foliação plano-axial de dobras geram três tipos de feições de relevo: planaltos ondulados não erodidos, chapadas recortadas (Serra dos Bois e Serra da Meia Noite) e escarpas onduladas perpendicularmente ao eixo das dobras onde os sinformes são os pontos baixos das serras e as antiformes os pontos altos (Serra dos Rogados) e imprimem lineamentos muito sinuosos e contínuos com alta amplitude em dobras com caimentos de eixo horizontalizado e baixa amplitude em dobras com caimentos de eixo verticalizados.

As foliações que circundam corpos ígneos intrusivos, formam domos de arqueamento causado pelo processo de intrusão (exemplo, Compartimento Baixo Rio Preto) e imprimem

traços circulares a semicirculares descontínuos e concêntricos, frequentemente com evidência de mergulho de camada centrífugo, para fora do corpo intrusivo. Essa feição pode ser confundida com aquelas impressas por domos estruturais e braqui-anticlinais (exemplo, a norte do Compartimento São João), todavia esse último caso ocorre maior continuidade dos lineamentos, salvo em condições erosivas acentuadas.

Tabela 1 – Relação entre estruturas geológicas e os tipos de lineamentos extraídos.

Estruturas Geológicas	Características dos Lineamentos	Exemplos
Falhas frontais de regime compressivo	Longos e contínuos, pouco sinuosos em falhas reversas ou rampas e muito sinuosos em empurrões ou patamares	Falha do Rio Paranã, Falha do Rio Maranhão, Falha da Serra dos Cristais.
Zonas transcorrentes compressivas e rampas laterais	Longos e contínuos, retilíneos a levemente sinuosos	Serra de Santana, Serra da Boa Vista, Serra do Boqueirão
Falhas distensivas rúpteis tardias	Retilíneos e curtos, repetição paralela	Compartimento Alto Tocantinzinho (1B)
Franco de dobras	Muito sinuosos e contínuos, alta amplitude em caimentos de eixo suaves e baixa amplitude em caimentos de eixo verticalizados.	Serra do Ticunzal, Serra do Tombador, Serra Santana de Cavalcante, Serra Boa Vista de Cavalcante
Eixos de dobras	Descontínuos em dobras com caimento de eixo e contínuos em dobras com eixo horizontal	Compartimento São João D’aliança (1A)
Corpos intrusivos	Concêntricos e descontínuos	Compartimento Baixo Rio Preto (3B)
Domos estruturais e braqui-anticlinais.	Concêntricos e contínuos	Compartimento São João D’aliança (1A)

5.2. Compartimentação Geomorfológica e Geotectônica

A seguir são descritas as principais características de cada um dos 16 compartimentos geomorfológicos obtidos pelo método proposto. As áreas de abrangência, bem como o número de cada compartimento, estão representadas na Figura 6.

Quatro conjuntos de compartimentos são propostos, incluindo: 1) Planaltos Dissecados,

2) Vales, 3) Vales com Residuais e 4) Escarpas. Os critérios utilizados para a compartimentação incluem: padrão de relevo, grau de dissecação do relevo, hipsometria, densidade de drenagem, tipos de solos e substrato rochoso. Nos compartimentos de escarpas predominam Neossolos Litólicos e Cambissolos associados a amplas exposições rochosas.

Em função de peculiaridades e variações locais hipsométricas, do grau de dissecação, densidade de drenagem, grau de rugosidade do relevo e substrato rochoso os compartimentos foram classificados e denominados a partir de uma designação geográfica local e são descritos a seguir.

5.3. Compartimentos Geomorfológicos

1A) Planalto Dissecado de São João D'Aliança: caracterizado por padrão de relevo suave ondulado, cotas entre 1200 e 1300 m, moderada densidade de drenagem, predominância de Latossolos, com Cambissolos subordinados e contexto geológico associado ao Grupo Paranoá em um trem de dobras fechadas oeste com eixo curvo e ondulante e vergência variando desde NE até SE que suavizam em direção ao extremo leste do compartimento até se tornarem monoclinal com mergulho para W. Nível crustal médio-superior.

1B) Planalto Dissecado do Alto Tocantininho: apresenta padrão de relevo suave ondulado a ondulado, cotas variáveis entre 1100 e 800 m, predominância de Cambissolos e contexto geológico marcado por uma ampla dobra sinclinal suave no Grupo Paranoá. O eixo do rio que dá nome a este compartimento apresenta padrão em cânion em diferentes trechos. Nível crustal médio-superior.

1C) Planalto Dissecado do Alto Rio Preto: marcado por um padrão de relevo ondulado a forte ondulado, com vales incisos, presença de hog backs e cuestras, com predominância de Neossolos Litólicos, cotas variáveis desde maiores que 1200 a 1676 m e substrato associado a metapsamitos do Grupo Araí. Este compartimento é preservado entre dois corredores regionais de cisalhamento. Nível crustal médio-inferior.

1D) Planalto Dissecado do Rio da Prata: apresenta padrão de relevo ondulado, predominância de Neossolos Litólicos, com amplas exposições rochosas, cotas superiores a 1150 metros, presença de vales incisos, moderada densidade de drenagem com padrão paralelo e substrato geológico representado pelo Grupo Araí, presença de dobras suaves com eixo predominante de direção aproximada N-S. Nível crustal relativo intermediário. Nível crustal médio-inferior.

2A) Vale do Rio Paranã: caracterizado por padrão de relevo suavemente ondulado, presença de Latossolos e Cambissolos, moderada densidade de drenagem, cotas inferiores a 600 metros e

substrato geológico correspondente a pelitos e arcóseos do topo do Grupo Bambuí, com ocorrências subordinadas de rochas carbonáticas da seção mediana do Grupo Bambuí. Nível crustal raso.

2B) Vale do Médio Rio Preto/Rio das Almas: definido por padrão de relevo suave ondulado, com cotas inferiores a 600 metros, predominância de Cambissolos cascalhentos (fragmentos de veios de quartzo), moderada densidade de drenagem dendrítica e substrato representado por granitos da Suíte Aurumina que representam o embasamento regional. Nível crustal profundo.

3A) Vale com Residuais do Baixo Rio Tocantinzinho: composto por relevo com padrão ondulado a forte ondulado com cotas entre 550 a 1000 metros, substrato geológico composto por rochas dos grupos Paranoá e Araí em que as residuais são sustentadas pelas principais camadas de quartzitos do Grupo Paranoá dobradas com eixo de direção predominante NW-SE. Nível crustal médio.

3B) Vale com Residuais do Baixo Rio Preto: apresenta padrão de relevo suave ondulado a ondulado com cotas entre 350 a 1000 metros, substrato geológico compostos por rochas com diferente resistência à denudação, incluindo metassedimentos de baixo grau do Grupo Araí e intrusões graníticas responsáveis pela manutenção das residuais. Nível crustal médio-inferior.

3C) Vale com Residuais do Ribeirão das Pedras: definido por padrão de relevo ondulado a forte ondulado com cotas entre 350 a 1000 metros, substrato geológico compostos por quartzitos, metaconglomerados e metavulcânicas do Grupo Araí e granitos intrusivos, sendo que as rochas de natureza psamítica e psefíticas são responsáveis pelas residuais. Nível crustal médio-inferior.

4A) Escarpa do Paranã: apresenta padrão de relevo escarpado a forte ondulado, amplitude de relevo variável de 600 a 1100 metros, predominância de Neossolos Litólicos, suportadas por quartzitos e metaconglomerados. É controlada por estrutura de falha de empurrão de baixo ângulo que coloca em contato tectônico os grupos Paranoá e Bambuí. É responsável pelo soerguimento do nível crustal médio-superior (Gr. Paranoá) acima do nível cristal raso (Gr. Bambuí).

4B) Escarpa da Serra da Boa Vista: caracterizada por padrão de relevo forte ondulado a escarpado, sendo suportada por quartzitos da seção basal do Grupo Paranoá e tem como controle secundário zona de cisalhamento regional de direção ENE. Sua amplitude de relevo pode alcançar localmente mais de 400 metros. Marca a transição stratigráfica entre os grupos Paranoá e Araí.

4C) Escarpa da Serra de Santana: apresenta padrão de relevo escarpado, amplitude de relevo variável de 500 a 1400 metros, predominância de Neossolos Litólicos, suportadas por quartzitos

e metaconglomerados. Estruturalmente é controlada por zona de cisalhamento em recuo erosivo. A feição marca a transição entre o nível crustal médio-inferior (Gr. Araí) e profundo (embasamento).

4D) Escarpa do Boqueirão: formada pelas serras do Boqueirão e Ticunzal, obedece a morfologia de flancos de dobras suaves nos quartzitos resistentes do Grupo Araí, tendo em sua base xistos da Formação Ticunzal. Localmente podem separar superfícies de cotas entre 1100 m no topo e 650 m na base. Marca a transição entre o nível crustal médio-inferior (Gr. Araí) e profundo (embasamento).

4E) Escarpas de Cavalcante: composta pelas serras das Araras, Santana e Boa Vista circundando a sede do município de Cavalcante. Obedecem à morfologia de flancos de dobras suaves em quartzitos do Grupo Araí e evoluem por erosão diferencial ao longo dos eixos com direção aproximada N-S. Estas escarpas podem alcançar 400 metros de desnível separando superfícies de cotas médias entre 1200 a 800 m. Representam a transição entre o nível crustal médio-inferior (Gr. Araí) e profundo (embasamento).

4F) Escarpas de Teresina de Goiás: é formada pela Serra do Boqueirão e parte da Serra Boa Vista compondo um corredor que se estende para norte no município de Teresina de Goiás. São controladas pelas mesmas zonas de cisalhamento que formam as serras da Boa Vista e Santana a sudoeste e são responsáveis por desníveis de mais de 400 metros com cotas que variam de 1100 a 650 m. Representam a transição entre o nível crustal médio-inferior (Gr. Araí) e profundo (embasamento).

4G) Escarpa da Serra do Cristal: apresenta declividade entre 50 e 60% paralela aos estratos das rochas do Grupo Paranoá, com desnível de até 400 m (desnível entre 1200 e 800). É controlada pela Falha da Serra do Cristal, que define sua direção e morfologia. Encontra-se segmentada e deslocada por zona de cisalhamento de direção NE formando uma ampla escarpa intermitente no relevo. Não marca a transição entre grupos ou níveis crustais, mas entre domínios deformacionais completamente distintos que afetam as rochas do Grupo Paranoá.

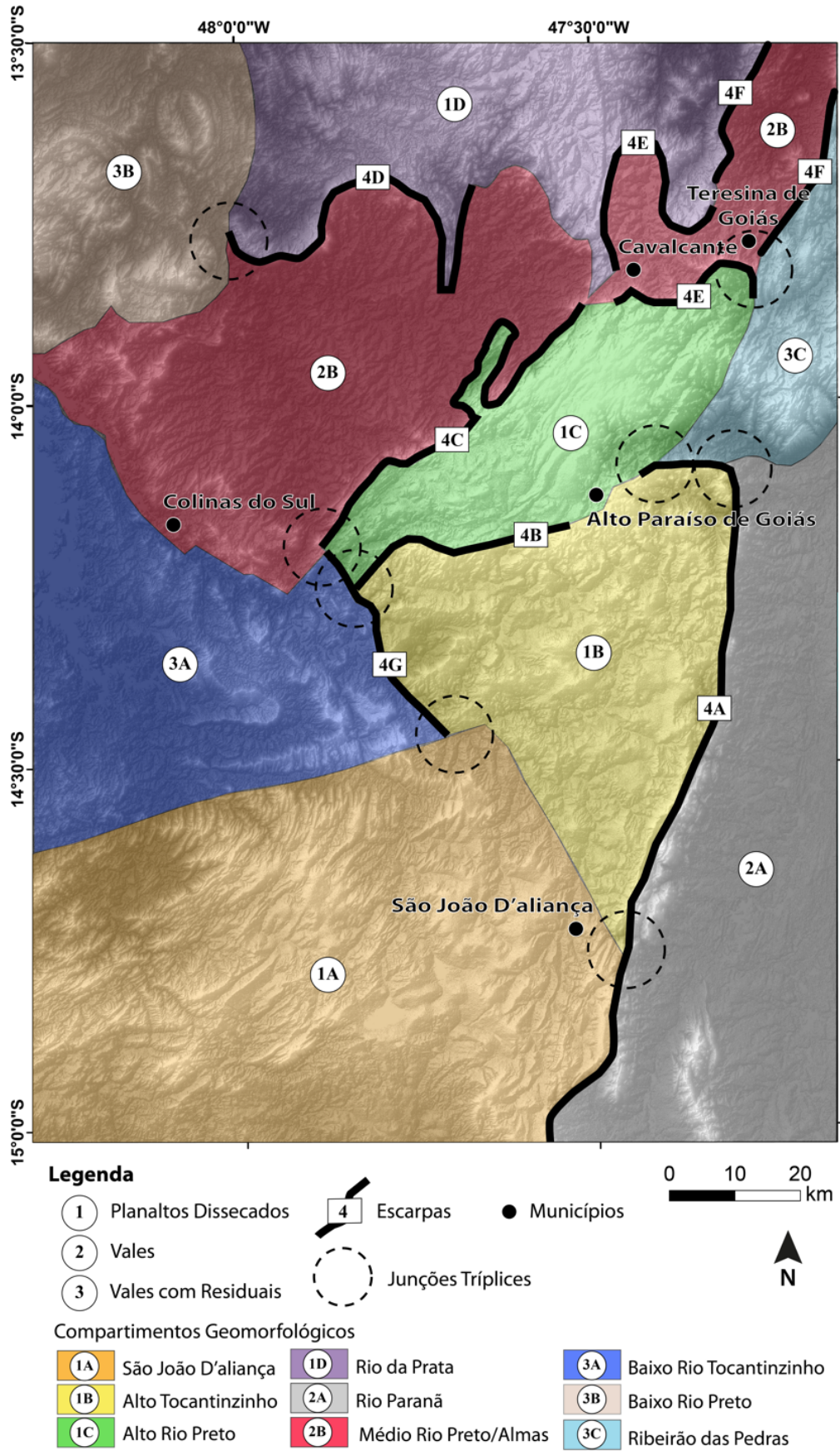


Figura 6 - Mapa de compartimentos geomorfológicos.

6. Discussões

O método de compartimentação geomorfológica proposto separa domínios com estilos deformacionais distintos de acordo com o nível crustal exposto ou com os tensores de deformação locais e prevê limites de acordo com as estruturas tectônicas mais proeminentes.

As características morfográficas dos limites dos compartimentos geomorfológicos traçados, bem como o padrão de lineamentos que delimita e preenche cada compartimento, possibilitam uma análise primária das relações temporais entre as estruturas que controlam o relevo permitindo estabelecer noções básicas sobre a evolução tectônica da área estudada.

Uma análise elementar prévia da evolução tectônica se faz extremamente útil como ponto de partida para o aprofundamento do trabalho já que auxilia na identificação de áreas-alvo para mapeamento geológico de detalhe e permite o levantamento de problemas geológicos específicos para os quais se podem eleger ferramentas auxiliares como a termocronologia, geocronologia, gravimetria, magnetometria, entre outras. Além disso, com o uso dos critérios adotados, os compartimentos geomorfológicos gerados já servem como balizadores para a delimitação dos compartimentos tectônicos em estudos posteriores.

6.1. Análise do Mapa de Lineamentos

A análise do mapa de lineamentos (Figura 4) consiste em avaliar a relação de corte entre cada tipo e família de lineamentos e com isso inferir, quando possível, a temporalidade dos eventos geradores dessas estruturas, conforme descrito a seguir.

Os lineamentos de direção NE-SW (destacados em azul) no norte da área são de caráter transcorrente, cortam e deslocam todos os demais lineamentos, exceto aqueles de direção Norte-Sul totalmente retilíneos (marcados em verde) e por isso são interpretados como gerados e/ou reativados em fase compressiva tardia.

O lineamento que marca a Falha da Serra do Cristal (em vermelho) encontra-se deslocado na direção nordeste do seu traço principal, que é formado pelo lineamento da mesma cor que vem de noroeste e encaixa no rio Tocantinzinho (Lineamento Baixo Tocantinzinho). Estes dois lineamentos, onde se unem e são deslocados pela família de lineamentos nordeste, separam dois domínios deformacionais distintos a nordeste e a sudoeste de seus traços, a notar pela mudança brusca (perpendicular) de direção dos eixos de dobras entre um lado e outro. Infere-se, portanto, tratar-se de feição interligada com falhas profundas, que representam descontinuidades suficientemente significativas ao ponto de particionar a deformação. O mecanismo de desacoplamento (*decoupling*) apresentado por Handy *et al.* (2005) exemplifica como a partição

da deformação ocorre em função de descontinuidades originadas em falhas profundas.

O lineamento que marca a Serra Geral do Paranã (em vermelho na porção leste do mapa) é cortado por todas as demais estruturas e, portanto, é interpretado como a primeira estrutura regional gerada na área. Este lineamento aparentemente era contínuo com o lineamento (também em vermelho) mais a norte de direção NE e foi interrompido e deslocado pela estrutura transcorrente que gerou o lineamento que controla a Serra da Boa Vista e segue por Alto Paraíso de Goiás até cruzar e deslocar o lineamento da Serra Geral do Paranã.

Os lineamentos que caracterizam zonas de cisalhamento regionais (destacados em azul), especificamente aqueles que coincidem com as escarpas das serras da Boa Vista e de Santana são importantes, pois delimitam a rotação no sentido horário do bloco cujo substrato é associado ao Grupo Araí, em contato a norte com rochas graníticas do embasamento cristalino e ao sul com rochas do Grupo Paranoá.

A última feição a se instalar na região deve ter sido a família de lineamentos descontínuos e retilíneos (em verde), dos quais os mais tardios foram aqueles de direção N-S, já que são totalmente retilíneos e não são truncados por nenhuma outra estrutura regional e devem representar falhas normais de abatimento do orógeno Brasília.

6.2. Análise do Mapa de Compartimentos Geomorfológicos

A análise do mapa de compartimentos tectônicos (Figura 6) deve ser realizada em conjunto com a análise do mapa de lineamentos. A forma dos compartimentos, a presença de endentações, os tipos de contatos entre os compartimentos, bem como os tipos de estruturas que os limitam e que os integram internamente, são os elementos que devem ser considerados nesta avaliação. As principais informações relativas aos compartimentos estão resumidas na tabela 2 e detalhados a seguir.

O compartimento **1A** (Planalto Dissecado de São João D'Aliança) apresenta forma aproximadamente retangular e é delimitado por estruturas que mostram evidências de falhas de rasgamento de movimentação sinistral a norte e nordeste que definem limites retilíneos e a leste por falha de empurrão que imprime limite ligeiramente curvo. É constituído internamente por dobras isoclinais a fechadas a oeste e suaves a leste, até próximo à Serra Geral do Paranã. Essas dobras possuem eixo curvo, principalmente a oeste, com caimento para sul e sudoeste, formando um padrão semicircular peculiar. A análise das estruturas que delimitam e compõem o bloco formado por esse compartimento permite sugerir que as rochas foram encurtadas no sentido E-W e avançaram nas direções nordeste, leste e sudeste em relação aos compartimentos

imediatamente a norte.

Tabela 2 – Características principais dos compartimentos geomorfológicos.

Tipo	Sigla	Limites Principais	Controles Erosivos Principais
Planaltos Dissecados	1A	Retilíneos e curvos: Falhas de Rasgamento e Falhas de Empurrão.	Dobras, Falhas de Rasgamento
	1B	Retilíneos e curvos: Zonas Transcorrentes e Falhas de Empurrão.	Litológico, Falhas Normais
	1C	Retilíneos e curvos: Zonas Transcorrentes	Litológico, Zonas Transcorrentes
	1D	Curvos: Foliações e Dobras	Dobras, Zonas Transcorrentes
Vales	2A	Retilíneos e Curvos: Falhas de Empurrão	Dobras, Litológico, Falhas de Empurrão
	2B	Retilíneos e curvos: Dobras e Zonas Transcorrentes	Dobras, Zonas Transcorrentes
Vales com Residuais	3A	Retilíneos: Falhas de Empurrão e Falhas de rasgamento	Dobras, Zonas Transcorrentes
	3B	Curvos: Domos intrusivos	Corpos Graníticos Elipsoidais e Foliações Associadas

O compartimento 1B (Planalto Dissecado do Alto Tocantinzinho) apresenta forma aproximada triangular e é delimitado por escarpas acentuadas (exceto a sudoeste) que coincidem com falhas de empurrão a leste e a oeste e zona transcorrente a norte. Internamente possui estruturação de grandes dobras suaves com direção de eixo NE-SW e juntas perpendiculares à direção dos eixos. Trata-se do compartimento que exhibe menor grau de deformação dentre todos os demais. Apresenta endentação formando junção tríplice (indicado por círculo tracejado na figura 6) de compartimentos a sudoeste, resultado da propagação, para leste, dos compartimentos a norte deste.

O compartimento 1C (Planalto Dissecado do Alto Rio Preto) apresenta forma sigmoide e é tanto delimitado quanto estruturado internamente por zonas de cisalhamento de direção nordeste-sudoeste. Ao norte o compartimento é delimitado pela ZC Colinas-Cavalcante, que controla a estruturação da Serra Santana e ao sul é delimitado pela ZC Colinas- Alto Paraíso, que

controla as Serras da Boa Vista e a parte E-W da Serra Geral do Paranã. A ZC Colinas-Alto Paraíso representa estrutura de escala regional e é responsável pela segmentação da Falha do Paranã a leste do município de Alto Paraíso de Goiás. Entre as inúmeras faixas de cisalhamento que o compõe, ocorrem dobras suaves com direção de eixo NE-SW e caimento para SW. O traçado das zonas de cisalhamento, bem como o estilo das dobras internas a este compartimento sugerem rotação dextral deste bloco.

O compartimento 1D (Planalto Dissecado do Rio da Prata) é delimitado ao sul por escarpas formadas por flancos de dobras que encurta o Grupo Araí na direção E-W. Internamente ocorrem predominantemente dobras com eixos sub-horizontais de direção N-S e ao norte da Serra do Tombador (Figura 5) ocorre dobra suave com eixo E-W e caimento para W. A erosão preferencial se dá ao longo dos eixos N-S das antifomes, provocando padrão ondulado no traçado das cristas de relevo que formam o limite sul do compartimento. O limite leste do compartimento é controlado pela extensão da ZC Colinas-Cavalcante que forma a Serra Boa Vista no extremo nordeste da área de estudo. Para oeste o compartimento é delimitado pela variação do padrão do relevo gerada pelas intrusões graníticas da Suíte Pedra Branca, formando vale com residuais marcado por lineamentos elípticos que circundam os corpos intrusivos.

O compartimento 2A (Vale do Rio Paranã) é alongado na direção N-S e é delimitado a oeste pela Serra Geral do Paranã. Internamente ocorrem cristas de serras onduladas de direção geral N-S formadas por rochas resistentes (quartzitos) do Grupo Paranoá. Os baixios são sustentados por rochas do Grupo Bambuí onde predominam sequências pelíticas e carbonáticas encurtadas na direção E-W.

O compartimento 2B (Vale do Médio Rio Preto/Rio das Almas) é delimitado por escarpas íngremes que marcam o contato entre o embasamento cristalino que compõe a maior extensão do compartimento (Suíte Aurumina e Formação Ticunzal) com as rochas quartzíticas resistentes do Grupo Araí. Internamente é cortado por estruturas paralelas à ZC Colinas-Cavalcante que predominam na porção sul do domínio. O Vale do Médio Rio Preto representa porção mais elevada do embasamento cristalino e foi gerado por erosão diferencial facilitada pelas estruturas da ZC Colinas-Cavalcante na sua porção sul e pelos eixos de anticlinais na porção norte do domínio.

O compartimento 3A (Vale com Residuais do Baixo Rio Tocantinzinho) apresenta forma triangular e possui junções tríplices a sul com os compartimentos 1A e 1B (indicado por círculo tracejado na figura 6) e a leste com os compartimentos 1C e 2B, onde apresenta endentação formada pela ZC Colinas-Cavalcante. É delimitado a norte por estrutura importante de direção

NW-SE que encaixa o rio Tocantinzinho e limita domínios deformacionais bastante distintos. A leste é limitado pela falha da Serra do Cristal, que também limita domínios deformacionais bastante distintos e aparenta ser a continuidade para sudeste da ZC Colinas-Serra da Mesa, deslocada pelo conjunto de zonas de cisalhamento Colinas-Cavalcante e Colinas-Alto Paraíso.

O compartimento 3B (Vale com Residuais do Baixo Rio Preto) possui limites curvos e é internamente marcado por lineamentos elípticos que circundam corpos graníticos intrusivos elipsoidais. O padrão erosivo é controlado pelos corpos intrusivos e pelas foliações geradas ao redor das intrusões nas rochas sedimentares encaixantes.

7. Conclusões

A abordagem adotada para a compartimentação do relevo aplicou critérios genéticos baseados em informações geológicas para a distinção das formas de relevo e lineamentos, pois é focada na avaliação geotectônica preliminar. Entretanto, para a composição de mapas de compartimentação geomorfológica para fins aplicados outros critérios ou parâmetros devem ser considerados, incluindo: comprimento de rampas, declividade, formas das encostas e tipos de solos.

A compartimentação geomorfológica da região da Chapada dos Veadeiros em quatro grandes conjuntos com respectivos subconjuntos é considerada adequada aos objetivos do presente trabalho. Os planaltos com diferentes graus de dissecação, os vales e os vales com residuais compõem regiões com contextos geológicos e geotectônicos distintos, tanto com relação ao nível crustal exposto quanto com relação ao padrão deformacional impresso. As escarpas compõem zonas de transição entre os demais compartimentos e por isso, têm área restrita e são alinhadas segundo as estruturas que as controlam (formam faixas estreitas e alongadas). As áreas que abrangem os limites entre domínios geomorfológicos e principalmente os limites tríplexes representam regiões de elevada importância para estudos detalhados.

O processo de checagem de áreas-chave em campo é essencial em trabalhos de evolução tectônica, pois estes estudos geralmente abrangem grandes áreas, impossíveis de serem percorridas em sua totalidade em trabalhos de detalhamento. Todavia a identificação das áreas-chave demanda estudos preliminares.

O método de mapeamento geomorfológico em escala regional com critérios e informações geológicas detalhadas provou ser uma ferramenta importante para a análise preliminar em estudos de evolução tectônica de uma região e pode ser aplicada a outras regiões em que o substrato controla de forma significativa o modelado do relevo. A aplicação do método

na área de estudo permitiu a geração de produtos como os mapas de lineamentos e compartimentos tectônicos, que permitem a identificação de áreas-chave onde devem ser realizados trabalhos de mapeamento estrutural de detalhe, bem como perfis-chave onde devem ser mapeadas seções geológico-estruturais, com o intuito de levantar dados de campo para análise mais aprofundada da evolução tectônica local e regional.

Agradecimentos

Ao Instituto de Geociências da Universidade de Brasília pelos meios destinados a viabilizar os trabalhos de campo e de escritório. Ao corpo editorial da Revista Brasileira de Geomorfologia e aos revisores anônimos que contribuíram para o aperfeiçoamento do artigo.

Referências Bibliográficas

- AB'SABER, A. N. (1972). Participação das depressões periféricas e superfícies aplainadas na compartimentação do planalto brasileiro: considerações finais, conclusões. *Geomorfologia/USP*, 28, São Paulo, p. 1-38.
- ALVES, F.C.; ROSSETI, D.F. (2015). Análise morfoestrutural e neotectônica na porção norte da Bacia Paraíba (PB). *Revista Brasileira de Geomorfologia*, v.16, n.4, (Out-Dez) p.559-578, 2015. DOI: <http://dx.doi.org/10.20502/rbg.v16i4.710>
- BEZERRA, F.H.R.; SAADI, A.; MOREIRA, J.A.M.; LINS, F.A.P.L.; NOGUEIRA, A.M.B.; MACEDO, J.W.P.; LUCENA, L.F. & NAZARÉ JR., D. (1993). Estruturação neotectônica do litoral de Natal-RN, com base na correlação entre dados geológicos, geomorfológicos e gravimétricos. *In: Simpósio Nacional De Estudos Tectonicos*, Belo Horizonte-MG, 1993. Anais, Sociedade Brasileira de Geologia - SBG/MG, Belo Horizonte- MG, 1993, p. 317-321.
- CASSETI, V. (1990). Introdução à geomorfologia. Disponível em: <http://www.funape.org.br/geomorfologia>.
- CHRISTOFOLETTI, A. (1980). *Geomorfologia*. S. Paulo: Editora Edgard Blucher Ltda.
- CLARK, M. & SMALL, J. (1982). *Slopes and weathering*. New York: Cambridge University Press. 110p.
- COSTA, J. A. V; FALCÃO, M. T. (2011). Compartimentação morfotectônica e implicações de evolução do relevo do hemigráben do Tacutu no estado de Roraima. *Revista Brasileira de Geomorfologia* v.12, n.1, p.85-94, 2011. DOI: <http://dx.doi.org/10.20502/rbg.v12i1.221>
- CORDEIRO, B. M.; FACINCANI, E. M.; PARANHOS FILHO, A. C.; BACANI, V. M.; ASSINE, M. L. (2010). Compartimentação geomorfológica do leque fluvial do rio Negro, borda sudeste da Bacia do Pantanal (MS). *Revista Brasileira de Geociências*, v. 40, n. 2, p. 175-183.
- DARDENNE, M.A. (2000). The Brasília Fold Belt. *In: U.G. Cordani, E.J. Milani, A. Thomaz Filho &*

- D.A. Campos. 2000. Tectonic evolution of South America. 31 st International Geological Congress, Rio de Janeiro, Sociedade Brasileira de Geologia. p. 231-263.
- DARDENNE, M.A.; CAMPOS, J.E.G. (2001). Parque Nacional da Chapada dos Veadeiros – GO. In: Sítios Geológicos e Paleontológicos do Brasil – SIGEP. Título da pesquisa disponível em <http://www.unb.br/ig/sigep/sitio096/sitio096.htm>. Acesso em 15/03/2001.
- DAVIS, W. M. (1899). The geographical cycle. *Geographical Journal*, 14:481-504.
- DEMEK, J. (1976). *Handbuch der geomorphologischen Detailkartierung*. Ferdinand Hirt. Viena.
- FONSECA, M.A.; DARDENNE, M.A.; UHLEIN, A. (1995). Faixa Brasília, Setor Setentrional: estilos estruturais e arcabouço tectônico. *Revista Brasileira de Geociências* 25(4):267-278.
- FREITAS-SILVA, F.H. & CAMPOS, J.E.G. (1998). *Geologia do Distrito Federal. Inventário Hidrogeológico e dos Recursos Hídricos Superficiais do Distrito Federal*. Brasília. Livro. IEMA/SEMATEC/UnB. Vol. 1, Parte I. 86p.
- GILBERT, G.K. (1877). *Geology of the Henry Mountains: U.S. Geographic and Geologic Survey of the Rocky Mountain region*, p. 4 [2d ed. 1880].
- GOIÁS. (2005). Agência Ambiental de Goiás. Mapa de Solos. Projeto Radam Brasil. Base de dados: Determinação de áreas prioritárias para unidades de preservação: - Cons. imagem/WWF. 08/07/2005. Disponível em: <http://www.sieg.go.gov.br/>
- GOMES, M.P.; VITAL, H. (2010). Revisão da Compartimentação Geomorfológica da Plataforma Continental Norte do Rio Grande do Norte, Brasil. *Revista Brasileira Geociências*, 40(3):321-329.
- HACK, J.T. (1960). Interpretation of erosional topography in humid temperate regions. *American Journal of Science* 258A, p. 80-97.
- HANDY, M.R.; BABIST, J, WAGNER, R, ROSENBERG, C.L., KONRAD-SCHMOLKE, M. (2005). Decoupling and its relation to strain partitioning in continental lithosphere: insight from the Periadriatic fault system (European Alps). *Geological Society London Special Publication* 243:249–276. DOI:10.1144/GSL.SP.2005.243.01.17
- IBGE, Fundação Instituto Brasileiro de Geografia e Estatística. (1995). *Manual Técnico de Geomorfologia*. (Coordenação Bernardo de Almeida Nunes et al.). Série Manuais Técnicos em Geociências. Número 5, Rio de Janeiro.
- MARENT, B.R.; VALADÃO, R.C. (2015). Compartimentação geomorfológica dos planaltos escalonados do sudeste de Minas Gerais - Brasil. *Revista Brasileira de Geomorfologia*, São Paulo, v.16, n.2, (Abr-Jun) p.255-270. DOI: <http://dx.doi.org/10.20502/rbg.v16i2.634>
- MARQUES NETO, R.; PEREZ FILHO, A. (2013). Papel da tectônica na dinâmica e evolução da paisagem na Serra da Mantiqueira, sudeste do Brasil. In: *Encuentro de Geógrafos de América Latina*, Lima-Peru, 2013. *Anales do XIV Encuentro de Geógrafos de América Latina*. Lima-Peru, p. 1-17. Disponível em: < <http://www.egal2013.pe/10-geografia-fisica/>>. Acesso em: 3 jun 2015.
- MORISAWA, M. (1975). Tectonics and geomorphic models. in: MELHORN, W.N. & FLEMAL,

- R.C.(edits.). Theories of landform development. G. Allen & Unwin, London, p.199- 216.
- O'LEARY D.W.; FRIEDMAN J.D.; POHN H.A. (1976). Lineament, linear, lineation: some proposed new standards for old terms. *Bulletin of Geological Society of America* 87:1463-1469.
- OLLIER, C. 1981. *Tectonics landforms*. Longman, London.
- PENCK, W. (1924). *Die morphologische Analyse. Ein Kapietel der physikalischen Geologie*. Engelhorn's Nachf., Stuttgart, 283 p. (Tradução da obra em em língua inglesa por CZECH, H. & BOSWELL, K.C. (1953). *Morphological analysis of landforms*. MacMillan and Co., London, 430 p.).
- PIMENTEL M.M., FUCK R.A., BOTELHO N.F. (1999). Granites and the geodynamic history of the Brasília Belt, central Brazil: a review. *Lithos* 46:463-483. DOI: 10.1016/S0024-4937(98)00078-4
- SAADI, A. (1991). *Ensaio sobre a morfotectônica de Minas Gerais*. Belo Horizonte-MG, IGC/UFMG. Tese de Professor Titular, 300 p.
- SAADI, A. & TORQUATO, J.R. (1992). Contribuição à neotectônica do Estado do Ceará. *Revista de Geologia, Fortaleza-CE*. vol. 5:5-38
- SAADI, A. (2013). Modelos morfogenéticos e tectônica global: Reflexões conciliatórias. *Geonomos*, 6(2):55-63. DOI: <http://dx.doi.org/10.18285/geonomos.v6i2.170>
- SANTOS, L.J.C.; OKA-FIORI, C.; CANALI, N.E.; FIORI, A.P.; SILVEIRA, C.T.; SILVA, J.M.F. (2009). Morphostructural Mapping of Paraná State, Brazil. *Journal of Maps*, 5:1, p.170-178, DOI: 10.4113/jom.2009.1059
- SOIL SURVEY STAFF. (1999). *Soil taxonomy: A basic system of soil classification for making and interpreting soil surveys*. 2nd edition. Natural Resources Conservation Service. U.S. Department of Agriculture Handbook, 436 p.
- SOUZA FILHO, P. W. M.; QUADROS, M. L. E. S.; SCANDOLARA, J. E.; SILVA FILHO, E. P.; REIS, M. R. (1999). Compartimentação morfoestrutural e neotectônica do sistema fluvial Guaporé-Mamoré-Alto Madeira, Rondônia, Brasil. *Revista Brasileira de Geociências*, 29(4):469-476.
- TAVARES, B.A.C.; CORRÊA, A.C.B; DE LIRA, D.R.; CAVALCANTI, L.C.S. (2014). Compartimentação geomorfológica e morfotectônica do Gráben do Cariatá, Paraíba, a partir de imageamento remoto. *Revista Brasileira de Geomorfologia* - v. 15, nº 4 (Out-Dez) p.523-538, 2014. DOI: <http://dx.doi.org/10.20502/rbg.v15i4.531>
- TOPODATA. Banco de Dados Geomorfométricos do Brasil. Disponível em: <http://www.dsr.inpe.br/topodata/index.php>. Acesso em 01 ago 2014.
- TRICART, J. (1965). *Principes et méthodes de le geomorphologie*. Paris:Masson Ed., 201p.
- TRICART, J. (1978). *Géomorphologie applicable*. Paris: Masson, 204 p.
- USGS. United States Geological Survey. Earth Explorer. Disponível em: <http://earthexplorer.usgs.gov/>

CAPÍTULO 3

Artigo 2:

**Proterozoic intracontinental basin succession in the São Francisco Craton
western margin constrained by detrital zircon geochronology**

Proterozoic intracontinental basin succession in the São Francisco Craton western margin constrained by detrital zircon geochronology

Marco Antonio Caçador Martins-Ferreira¹ marcoc@unb.br (Corresponding Author); **Farid Chemale Jr²**; **Airton Natanael Coelho Dias³**; **José Eloi Guimarães Campos¹**

1. Instituto de Geociências, Universidade de Brasília, Campus Universitário Darcy Ribeiro, Asa Norte, 70.910-900 Brasília, DF, Brazil.
2. Programa de Pós-Graduação em Geologia. Universidade do Vale do Rio dos Sinos, 93022-000 São Leopoldo, RS, Brazil
3. Departamento de Física, Química e Matemática, UFSCar - Campus Sorocaba, 18052-780 Sorocaba, SP, Brazil

Abstract

The age and stratigraphic position of Paleoproterozoic to Mesoproterozoic covers in the western border of São Francisco Craton (SFC) are still uncertain. Based on detailed sedimentologic-stratigraphic survey, combined with U-Pb and Lu-Hf zircon dating, we present new proposal for the stratigraphy and correlation with similar sequences, situated in other regions of the SFC, the Chapada Diamantina and Northern Espinhaço. Our study demonstrates that the so-called Traíras Formation (Araí Group) has a maximum depositional age of 1543 ± 31 Ma. Zircon distribution pattern shows dominant Rhyacian source and minor contribution of Statherian, Calymmian, Neoarchean and Palearchean sources. The Rhyacian zircon grains have dominant positive $\epsilon_{\text{Hf}}(t)$ signature, suggesting a great contribution from the juvenile Paleoproterozoic terranes, most probably from the Goiás Massif, located west of the study area.

The Calymmian maximum depositional age, calculated for the upper Araí Group (Traíras Formation), leads to the conclusion that the Serra da Mesa Group (ca. 1.55 Ga maximum depositional age) and the Traíras Formation have coexisted in time, possibly forming a larger sag basin, respected their lateral heterogeneities. The data presented also allow a new geotectonic approach for the Goiás Massif, since it necessarily had to be connected to the São Francisco Craton during the Mesoproterozoic, excluding the possibility that this massif was accreted to the São Francisco Craton in the Brasiliano Collage, as previously believed. The study also sheds light to the always debated tectonic relationship between the Serra da Mesa Group and Serra da Mesa Granite (ca. 1.55 and 1.57 Ga, respectively).

Furthermore, based on the new data, we present a stratigraphic revision for the actual Araí and Paranoá groups, proposing to raise the Traíras Formation to group status and its former members to formation status, representing an individual sag-type basin, different from the Araí rift-type basin. We also propose the encompassing of all three Proterozoic intracontinental

extensional pulses (Araí, Traíras and Paranoá groups) as the Veadeiros Supergroup, respectively chronocorrelated to the lower, middle and upper Espinhaço Supergroup.

Keywords: Araí Group, Calymmian sag basin, Traíras Group, Veadeiros Supergroup, São Francisco Craton

1. Introduction

The São Francisco Craton contains mostly undeformed or faintly deformed Proterozoic covers in the cratonic core, whereas at the cratonic margins these rocks are deformed in different degrees, increasing towards the Brasiliano orogenic belts that border the craton (Almeida, 1977, Trompette et al. 1992, Chemale et al. 1993, Alkmim et al. 2001). Two major tectono-stratigraphic units can be recognized, the Late Paleoproterozoic-Mesoproterozoic Espinhaço and the Neoproterozoic São Francisco supergroups. In the last decade, these units were defined based on geochronology, geochemistry and geophysical data in light of plate tectonics view, as Statherian-Tonian intracontinental basins (1.8 Ga to 0.93 Ga) comprising three major sedimentary cycles or megasequences, as proposed by Chemale et al. (2012), Guadagnin et al. (2015) and Guadagnin and Chemale (2015). They show that these basins generally correspond to successor and poly-historic depositional sites filled by a series of unconformity-bounded units that record distinct subsidence pulses, distributed over relatively long time periods.

Many researchers tried to correlate the chronostratigraphic units of the Espinhaço Supergroup with similar sequences that outcrop in the adjacent Brasilia orogenic belt, as the Araí, Serra da Mesa and Paranoá Groups (e.g.: Guadagnin and Chemale, 2015 and references therein). Lately, Guadagnin and Chemale (2015), based on available stratigraphic description and few published geochronological data, suggested that these units affected by the Brasiliano Orogeny in the Brasilia Belt should have also similar stratigraphic distribution and sedimentary-stratigraphic evolution as those in the Espinhaço Supergroup.

Earlier studies, applying the concept of passive margin sequence stratigraphy characterize the Araí Group as a typical rift-post-rift basin and many speculate a possible correlation between the Araí upper sequence (Traíras Formation) and Serra da Mesa Group (Marini et al. 1984a; Fuck et al. 1988; Pimentel et al. 1991; Schobbenhaus 1996; Dardenne 2000; Marques 2009), but that hypothesis was never demonstrated by geochronological data. Indeed, the correlation between those basins is often difficult due to metamorphism, deformation and lateral stratigraphic variations. Besides deformation and metamorphism in the Araí and Serra da

Mesa Group units, we handle with few reliable geochronological constraints, since they form a major sedimentary package. Age and provenance data, allied to stratigraphic analysis can shed light to the problem by positioning the studied basins in space and time (e.g. Dickinson and Gehrels, 2009; Andersen, 2014).

This paper presents new detrital zircon U-Pb ages and Lu-Hf isotopes from the Araí and Paranoá cratonic sequences and compares it to published geochronological and stratigraphic data to evaluate (1) maximum depositional ages of these sequences, (2) detrital zircon age distributions, (3) potential sediment source areas, and (4) the tectonic and sedimentary evolution of the São Francisco craton basins and basement.

2. Geological Setting and Sampling

The São Francisco Craton hosts a series of Proterozoic intracontinental sedimentary sequences separated by unconformities (Figure 1). Most of these sequences were affected by Neoproterozoic Brasiliano orogeny and their correlation has become difficult due to deformation and metamorphism. In the western margin of the São Francisco Craton occurs the Neoproterozoic Northern Brasília Belt (Dardenne, 2000) where the Paleo/Mesoproterozoic sequences include the Araí, Serra da Mesa, Natividade and Paranoá Groups (Figure 2).

These sequences are underlain by Paleoproterozoic crystalline basement formed by different terranes, mainly the Goiás Massif, to the West and the Cavalcante-Almas-Natividade block, to the east. The Goiás Massif is commonly interpreted as an exotic block accreted to the São Francisco Craton during the Neoproterozoic (Fuck et al. 1994, 2017; Pimentel et al. 2000 and others). It is formed mainly by Archean granite-greenstone terrains and Paleoproterozoic orthogneiss. Geochronological and isotopic data for the Goiás massif indicate crystallization age around 2.14 Ga for orthogneiss and $\epsilon\text{Nd}(t)$ values between +2 and -7. The Cavalcante-Almas-Natividade block was formed during two main magmatic pulses, the first varying from 2.3 to 2.4 Ga, with positive $\epsilon\text{Nd}(t)$ values ranging from +2 to close to zero and a second main event from 2.15 to 2.2 Ga with generally slightly negative $\epsilon\text{Nd}(t)$ values ranging mainly from -1 to -3 (Fuck et al. 2014). That basement was intruded by anorogenic granites at 1.77 Ga (Pimentel et al. 1991) known as Pedra Branca Suite and by a second pulse at 1.57 or 1.55 Ga (Rossi et al., 1992 and Kitajima, 2001, respectively) known as Serra da Mesa Suite (Pimentel and Botelho, 2001).

Recently, Cordeiro and Oliveira (2017) have presented geologic and geochronologic data that allow the division of the Goiás Massif (Figure 1) into two domains, the Crixás-Goiás and Campinorte domains. These authors have also expanded the term Goiás Massif to comprise the

whole Brasília Belt crystalline basement older than 1.8 Ga, previously called by many different names (e.g. Cavalcante-Almas-Natividade block), and added two more domains, Cavalcante-Arraias and Almas-Conceição do Tocantins domains. The proposed domain boundaries are intimately related to regional Rhyacian crustal evolution and their division facilitate terrane location, thus these names are adopted in this work.

The Serra da Mesa Group corresponds to a succession of mainly micaschist with minor quartzite and marble deposited in a marine shelf setting. The maximum depositional age is 1557 ± 22 Ma (Marques, 2009). The group has been speculatively correlated to the Araí Group post-rift sequence (Delgado et al., 2003; Marques, 2009; Pimentel et al. 2011), but no consistent data has been presented to confirm that hypothesis. The Serra da Mesa Group basement is represented by the Goiás Massif. The Serra da Mesa Granites have intruded the Brasília Belt basement between 1.57 and 1.61 Ga (Pimentel et al. 1991; Rossi et al. 1992) and the Serra da Mesa basal quartzite contain 1.57 Ga zircons.

The Natividade Group is an intracontinental sequence composed of marble, quartzite, meta-greywacke and phyllite, considered to be a northern extension of the Araí Group post-rift phase (Costa et al. 1976; Marini et al. 1984b; Silva et al. 2005). The only geochronological data available for the Natividade Group refer to Pb-Pb detrital zircon ages that define a maximum depositional age of 1779 ± 6 Ma (Silva et al. 2005).

The Araí Group is historically interpreted as a rift-post-rift sequence deposited over a Paleoproterozoic crystalline basement. It was divided by Tanizaki et al. (2015) in four tectonosequences, pre-rift (Água Morna Formation), rift (Arraias Formation), transitional (Caldas Formation) and post-rift (Traíras Formation), by applying the method of tectonic system tracts proposed by Prosser (1993). The transitional sequence is marked by a series of intraformational conglomerates, suggesting an unconformity that represents sufficient time lapse for lithification, uplift and erosion to take place before the reestablishment of subsidence and the onset of the Traíras Formation basin. Previous geochronological data for the Araí Group comes from U-Pb ages of rift-related bimodal volcanic rocks (Buracão Member) that only occur in the rift sequence (Arraias Formation) and was dated 1771 ± 2 Ma (Pimentel et al. 1991). A prior study on detrital zircon ages from Araí rocks (Marques, 2009) at a different study area did not find zircons younger than 2.0 Ga, maintaining the maximum depositional age for the whole Araí Group (both Arraias and Traíras formations) at 1771 ± 2 Ma.

The Paranoá Group covers the Araí Group, and is separated from it by an unconformity marked by the São Miguel Formation conglomerates. The group represents a shallow water

marine succession composed mainly of sandstones and siltstones with subordinated limestone (Alvarenga et al. 2012, Campos et al. 2013). Available geochronological data for the Paranoá Group consists in detrital zircon U-Pb ages determining a maximum depositional age of 1560 ± 10 Ma and diagenetic xenotime U-Pb age suggesting deposition at 1042 ± 22 Ma (Matteini et al. 2012).

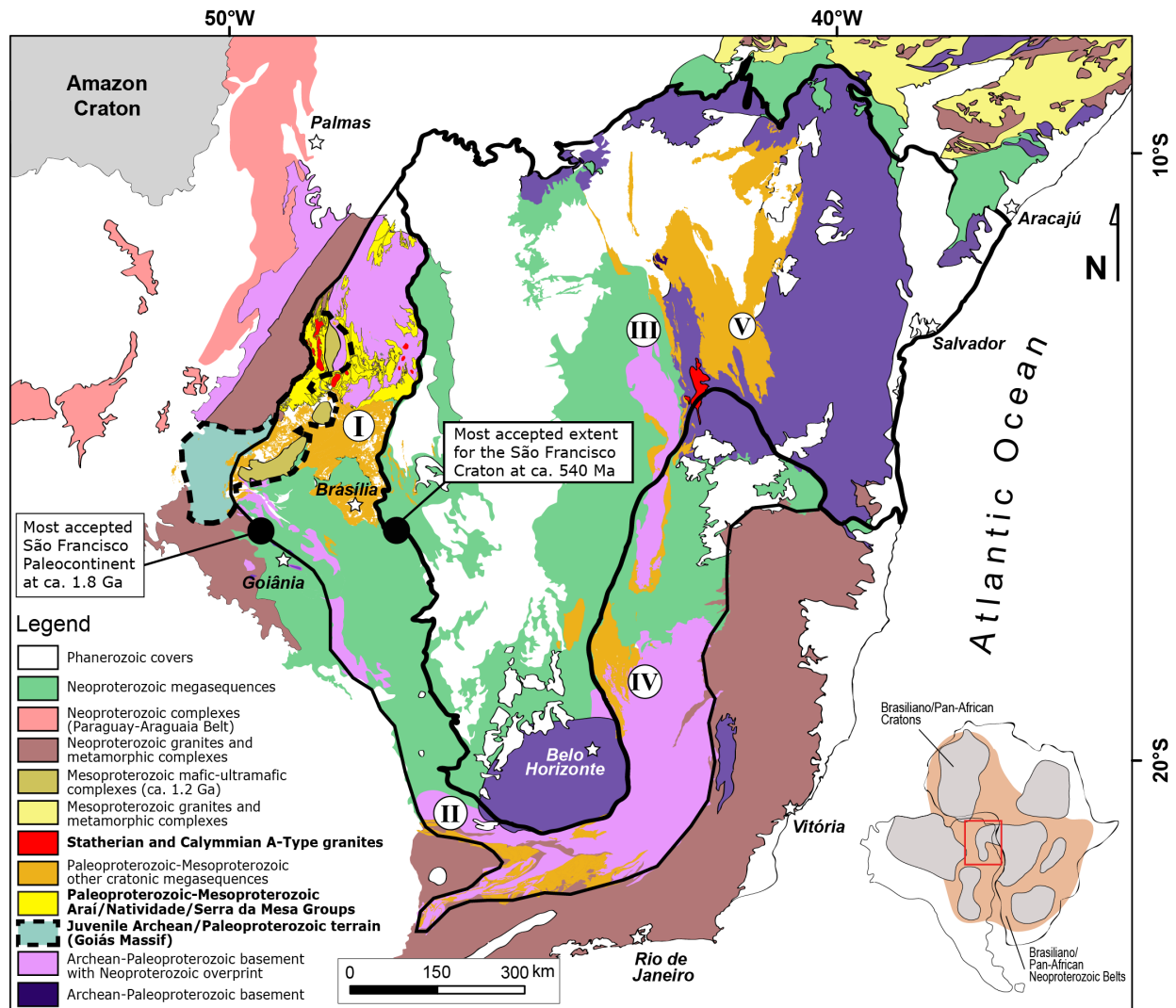


Figure 1 – Geologic map showing location of main intracontinental basins in the São Francisco Craton interior and margins. The black lines indicate the most accepted extent of the Craton after ca. 540 Ma and extent of the continental paleocontinent at ca. 1.8 Ga. Dotted blue line indicates the limits of the Goiás Massif according to Fuck et al. (1994). I) Northern Brasília Belt (Veadeiros Supergroup); II) Northern Espinhaço range; III) Diamantina Plateau; (Chapada Diamantina); IV) Southern Brasília Belt and V) Southern Espinhaço range (modified from Guadagnin and Chemale, 2015).

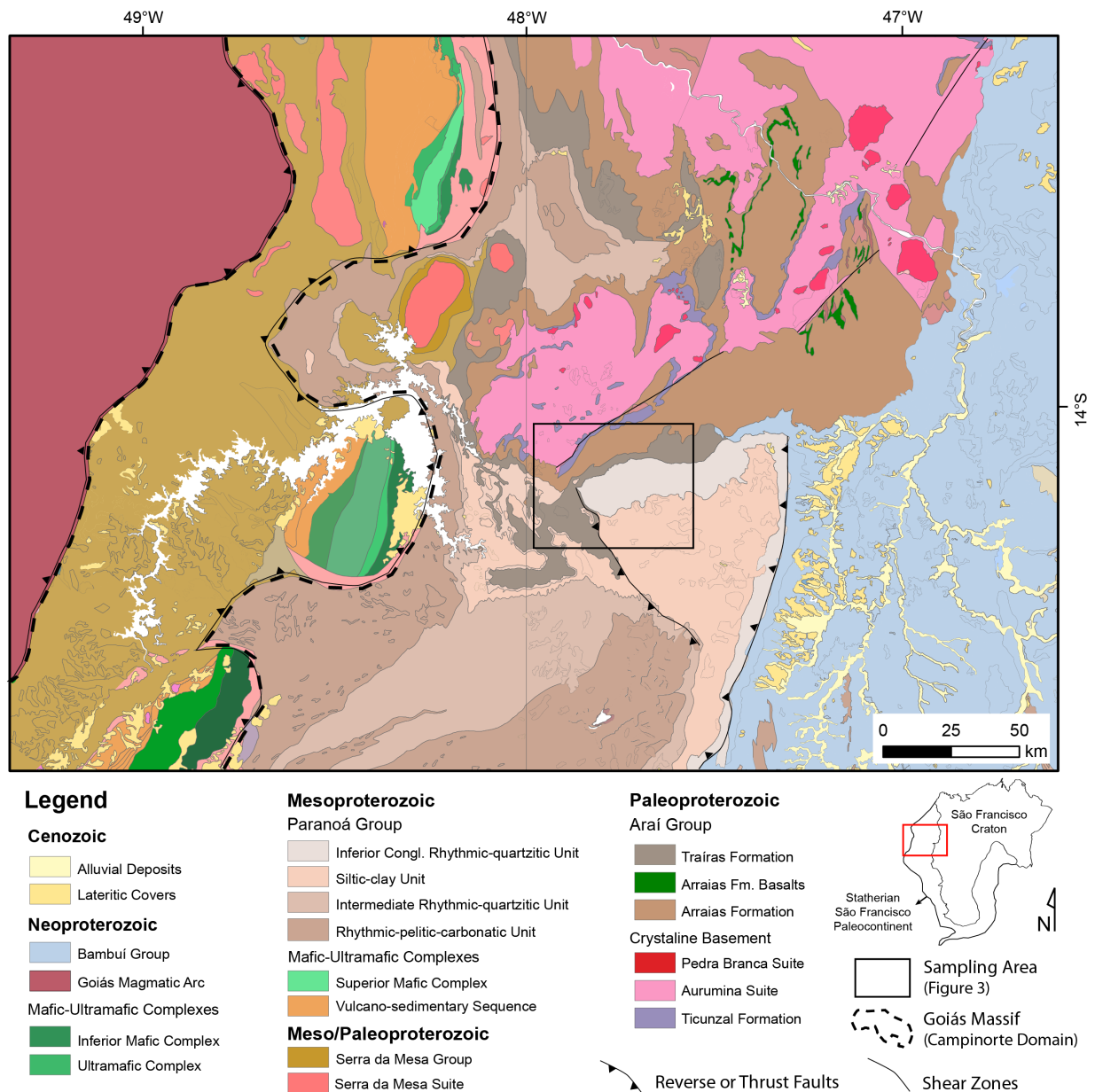


Figure 2 – Geologic map of the study area. Sampling area (detailed in figure 3) is outlined in black rectangle. The Goiás Massif - Campinorte domain (after Cordeiro and Oliveira 2017) is outlined in black dashed lines (modified from CPRM 2008).

2.1. Sampling

Twelve samples were collected from different stratigraphic levels of both Araí and Paranoá Groups and are shown on 1:100.000 scale unpublished geologic map (Figure 3). It is important to highlight, though, that all the samples were very well positioned on stratigraphic succession by local 1:25.000 scale geologic mapping.

Samples TFC 01, TFC 07 and TFC 10 are representative of Paranoá Group, São Miguel Formation, a basal conglomerate and sample TFC 12 represents Serra do Paranã Formation quartzite, also from Paranoá Group. Samples TFC 02, TFC 04, TFC 05 and TFC 06 are

representative of Araí Group, Traíras Formation quartzites and greywackes. Sample TFC 03 did not present satisfactory analytic results and TFC 08, TFC 09, TFC 11 did not contain enough zircons (Table 1).

Table 1 - Analyzed samples and respective rocktypes, formations and UTM Z23S coordinates.

Sample	UTM_X	UTM_Y	Group	Formation - Member	Rocktype	Zircons
TFC_01	181543	8428261	Paranoá	Ribeirão São Miguel	Conglomerate	44
TFC_02	182596	8427837	Araí	Traíras – Boqueirão	Graywacke	13
TFC_03	184895	8421872	Araí	Traíras – Rosário	Quartzite	55
TFC_04	188062	8425939	Araí	Traíras – Rio Preto	Quartzite	10
TFC_05	193811	8430754	Araí	Traíras – Rosário	Quartzite	60
TFC_06	197858	8431200	Araí	Traíras – Rosário	Quartzite	11
TFC_07	198998	8430014	Paranoá	Ribeirão São Miguel	Conglomerate	20
TFC_08	188620	8408704	Paranoá	Serra da Meia Noite	Quartzite	3
TFC_09	189652	8409467	Paranoá	Ribeirão Piçarrão	Siltite	1
TFC_10	193205	8411983	Paranoá	Ribeirão São Miguel	Conglomerate	23
TFC_11	200709	8414800	Paranoá	Ribeirão Piçarrão	Siltite	0
TFC_12	203264	8419879	Paranoá	Serra do Paranã	Quartzite	29

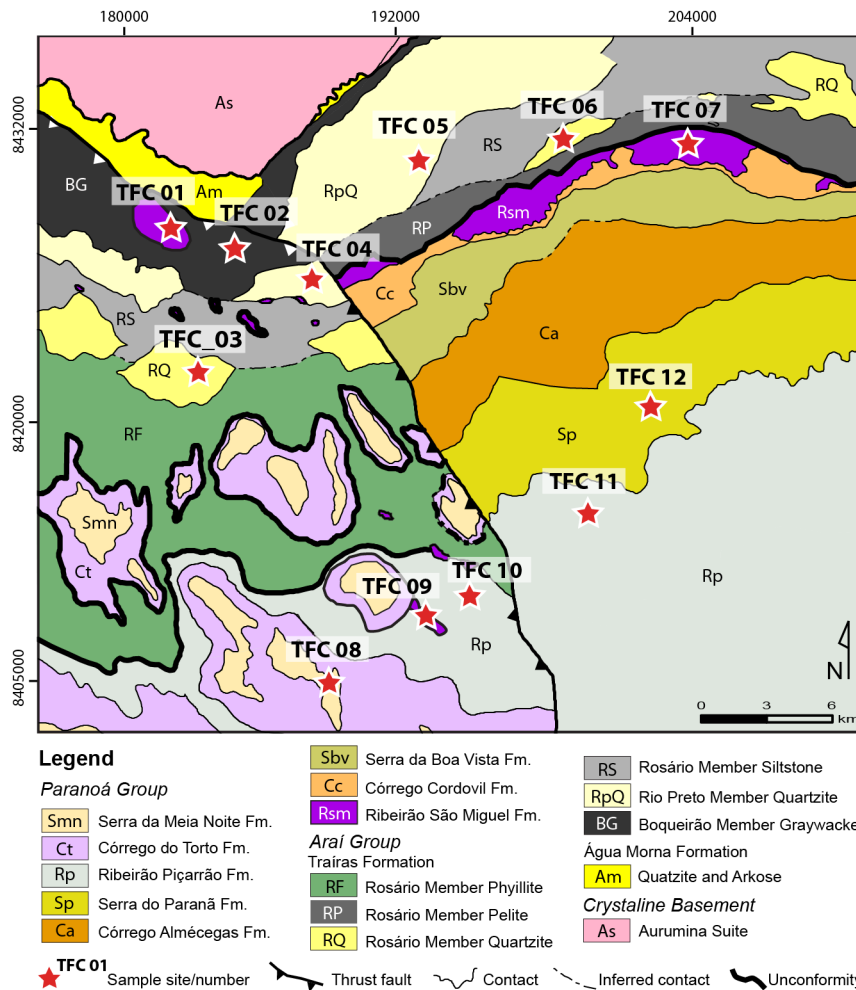


Figure 3 – Detail geologic map with sample sites and numbers indicated. Group, formation and member names are adopted from Tanizaki et al. (2015) for the Araí Group and Campos et al. (2013) for the Paranoá Group. See map location on figure 2.

A detailed description of each sample effectively used as provenance data in this study is presented below. Group, formation and member names are adopted from Tanizaki et al. (2015) for the Araí Group and Campos et al. (2013) for the Paranoá Group.

Araí Group Samples

Sample TFC 02 is a Greywacke from the Boqueirão Member, that represents the Traíras Formation (Araí Group) basal member and is often deposited directly over the crystalline basement. Composed of calciferous siltstone interlayered with quartzite and greywacke. The quartzite is interlayered with carbonatic material and at the top of the unit a sequence composed of well sorted quartzites, phyllites and marbles.

Sample TFC 04 is a quartzite derived from the Rio Preto Member, Traíras Formation (Araí Group) composed mainly of pure medium well sorted quartzite. Frequently coarse to very

coarse horizons occur. Cross stratification is abundant and locally conglomerate and siltstone may occur.

Samples TFC 05 and TFC 06 are quartzites derived from the Rosário Member, Traíras Formation, (Araí Group) composed by anchimetamorphic siltstone sequence, a sand/silt rhythmic sequence and a sand/silt/carbonate sequence.

Paranoá Group Samples

Samples TFC 01, TFC 07 and TFC 10 are conglomerates derived from Ribeirão São Miguel Formation (Paranoá Group), that represents the Paranoá Group basal formation. Most often occurs as matrix supported conglomerate with sandy matrix and carbonatic cement and rarely occurs as clast supported conglomerate. The clasts are angular or sub-rounded and most often represent metasilstones, fine quartzite and fine marble. The clasts may vary from millimeter to decimeter scale, but occasionally meter scaled boulders up to 8 meters wide occur. Based on clast size variation and in the sedimentary structures present, four different facies were distinguished: massive breccia facies, coarse to medium massive rudite facies, fine rudite facies and pelitic facies. The collected samples belong to the most frequent facies, the fine rudite facies.

Sample TFC 04 is a quartzite derived from the Serra do Paranã Formation (Paranoá Group), composed by medium to coarse quartzite layers. At the top layers, fine feldspathic conglomerate channels occur. Tabular cross-stratification is common.

3. Material and Methods

3.1. Sample Preparation

The Conglomerate samples were first prepared by separating the clasts from the matrix, only the matrix fraction was then prepared by the same procedure as all the other samples. The heavy minerals separation process was conducted for 12 samples. All samples were crushed and milled using jaw crusher and ring mill apparatus and concentrated first by hand panning, secondly by Frantz® magnetic separator and last by heavy liquids. Hand-picked zircons from the non-magnetic fraction were mounted randomly in epoxy and polished until the minerals were just revealed, photomicrographed in transmitted and reflect light, and imaged using scanning electron microscope for backscattered electrons (BSE) and cathode-luminescence (CL) images.

3.2. U-Pb method

Zircon grains were dated by the LA-SF-ICP-MS (monocollector ICP-MS) method at the Geochronology Laboratory of Universidade Federal de Ouro Preto. A Thermo-Finnigan Element 2 sector field ICP-MS coupled to a CETAC213 ultraviolet laser system was used. Laser spot size was set for 20 μm and data were acquired in peak jumping mode during 20 s background measurement followed by 20 s sample ablation.

Raw data were corrected for background signal and laser-induced elemental fractional and instrumental mass discrimination were corrected by the reference zircon (GJ-1) (Jackson et al. 2004). The common Pb correction was based on the Pb composition model (Stacey & Kramers 1975). To evaluate the accuracy and precision of the laser-ablation results, the GJ-1 (Jackson et al. 2004) and Plešovice zircon (337 ± 1 Ma; Sláma et al. 2008) were analyzed. Santos et al. (2017) described the detailed analytical methods and data treatment. The age calculation was carried out using Isoplot-Ex (Ludwig 2003).

3.3. Lu-Hf method

A Photon-Machines (ArF excimer laser 193 nm) laser ablation system coupled to a Neptune MC-ICP-MS at the Isotope Laboratory of Universidade Federal de Ouro Preto (UFOP) was used in this study to measure Lu, Yb and Hf isotopic signatures in zircons. The standards used during Hf analysis were 91500 and Mud Tank (Woodhead and Hergt 2005) and Temora zircon (Black et al. 2003). The $^{176}\text{Hf}/^{177}\text{Hf}$ accepted values for the standards were reproduced within error during all analytical session, yielding within-run results of 0.282511 ± 0.000019 (n=20, 2SD) for Mud tank; 0.282321 ± 0.000024 (n=20, 2SD) for 91500 and 0.282704 ± 0.000020 (n=8, 2SD) for Temora.

A laser energy of approximately $5 \text{ J}/\text{cm}^2$ with repetition rate of 5 Hz and spot size of 50 μm were used for all analyses. To improve the sensibility, N (~ 0.080 l/min) was mixed with Ar and He (+sample) gas in a gas mixer (Squid) before entering the torch. The typical signal intensity was 10 V for ^{180}Hf . A user-selected interval of approximately 60 data points covering the sample transient peak was used for calculation of the Hf ratio with mass bias correction using the exponential law. The isotopes ^{175}Lu , ^{171}Yb and ^{173}Yb were monitored during analysis and their relative abundances were used to calculate ^{176}Lu and ^{176}Yb interferences, which were subtracted from ^{176}Hf . The data were corrected in an Excel spreadsheet offline using $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$ (Patchett et al. 1981), $^{176}\text{Lu}/^{175}\text{Lu} = 0.2658$ (JWG in-house value), and $^{176}\text{Yb}/^{173}\text{Yb} = 0.796218$ (Chu et al. 2002) (see Gerdes and Zeh 2006).

4. Results and Discussion

The U-Pb analytic results for each sample (Table 2) are presented in histograms with relative probability plots in figure 4. The age data are clustered by formation and summarized in figure 5, where five main age peaks are evident. ϵ Hf versus age diagram, calculated from Lu-Hf and U-Pb data, also clustered by formation, are presented in Figure 6. Basin stratigraphy, U-Pb age intervals and ϵ Hf signature are used to correlate the studied units to the Espinhaço Supergroup first-order sequences and to indicate possible sediment sources as described below.

Table 2 - Summary of U-Pb zircon data for studied samples obtained by LA-SF-ICP-MS method.

Sample TFC-01				Isotope ratios ^b								Ages (Ma)						
Spot	Pb206*	Pb207	U	²⁰⁷ Pb*/	±	²⁰⁷ Pb*/	±	²⁰⁶ Pb*/	±	RH	²⁰⁶ Pb/	±	²⁰⁷ Pb	±	²⁰⁶ Pb/	±	Conc	
number	cps	cps	ppm	Th/U _a	²⁰⁶ Pb*	1s	²³⁵ U	1s	²³⁸ U	1s	O	²³⁸ U	1s	²³⁵ U	1s	²⁰⁷ Pb	1s	
15	17699	2482	28	0,30	0,1402	0,0018	7,520	0,104	0,401	0,005	0,92	2174	23	2175	12	2230	22	97
16	31945	4278	53	0,40	0,1339	0,0016	7,313	0,094	0,395	0,005	0,94	2147	22	2150	12	2150	21	100
17	35800	4879	59	0,56	0,1363	0,0017	7,150	0,099	0,391	0,005	0,92	2126	23	2130	12	2181	22	98
18	69859	7744	134	0,60	0,1109	0,0015	4,670	0,065	0,315	0,004	0,87	1763	19	1762	12	1813	24	97
19	81257	10909	134	0,32	0,1343	0,0014	6,911	0,085	0,385	0,005	0,99	2099	22	2100	11	2154	19	97
20	33503	4681	54	0,66	0,1397	0,0017	7,507	0,103	0,402	0,005	0,94	2179	24	2174	12	2224	21	98
21	26218	3578	44	0,60	0,1365	0,0018	7,154	0,106	0,392	0,005	0,90	2130	24	2131	13	2183	24	98
22	33062	4498	54	0,49	0,1360	0,0018	7,138	0,106	0,391	0,005	0,90	2128	24	2129	13	2177	24	98
23	26560	2964	52	1,20	0,1116	0,0013	4,811	0,064	0,322	0,004	0,92	1802	19	1787	11	1826	22	99
24	39589	7442	52	0,49	0,1880	0,0021	12,880	0,170	0,515	0,007	0,98	2677	28	2671	12	2725	19	98
25	53214	7625	85	0,43	0,1433	0,0016	7,728	0,101	0,406	0,005	0,97	2196	24	2200	12	2267	20	97
26	25235	3276	43	0,80	0,1298	0,0016	6,697	0,091	0,380	0,005	0,93	2075	22	2072	12	2096	22	99
34	38720	5282	64	0,30	0,1364	0,0016	7,051	0,096	0,389	0,005	0,95	2116	23	2118	12	2182	21	97
35	64304	8309	102	0,41	0,1292	0,0016	6,420	0,086	0,370	0,004	0,89	2031	21	2035	12	2087	23	97
37	52388	9853	65	0,43	0,1881	0,0020	12,897	0,162	0,515	0,006	0,99	2676	27	2672	12	2725	18	98
38	25459	3596	37	0,47	0,1412	0,0021	7,682	0,119	0,406	0,005	0,84	2195	24	2195	14	2243	27	98
39	79870	9930	132	0,18	0,1243	0,0014	6,388	0,082	0,385	0,005	0,96	2101	22	2031	11	2019	20	104
40	66333	9048	108	0,36	0,1364	0,0017	7,112	0,099	0,391	0,005	0,92	2126	23	2126	12	2182	22	97
42	30099	4155	49	0,57	0,1380	0,0016	7,206	0,097	0,393	0,005	0,95	2137	23	2137	12	2203	21	97
43	23653	3385	37	0,27	0,1431	0,0017	7,713	0,102	0,406	0,005	0,95	2198	23	2198	12	2265	21	97
44	40663	5622	67	0,31	0,1383	0,0016	7,261	0,099	0,395	0,005	0,94	2147	23	2144	12	2206	21	97
45	28352	3694	49	0,34	0,1303	0,0016	7,064	0,093	0,390	0,005	0,93	2122	22	2120	12	2102	21	101
53	15358	2095	26	0,61	0,1364	0,0022	7,031	0,120	0,388	0,006	0,84	2114	26	2115	15	2182	29	97
54	39099	4136	84	0,27	0,1058	0,0012	4,147	0,054	0,295	0,004	0,95	1665	18	1664	11	1728	21	96
55	15136	2288	24	0,45	0,1512	0,0025	8,042	0,144	0,415	0,006	0,85	2236	29	2236	16	2359	30	95
56	36610	7987	42	0,27	0,2182	0,0026	16,620	0,229	0,571	0,008	0,96	2913	31	2913	13	2967	20	98
57	32806	4419	57	0,74	0,1347	0,0016	6,570	0,093	0,376	0,005	0,93	2056	23	2055	12	2160	23	95
58	60029	9346	85	0,68	0,1557	0,0019	9,229	0,122	0,444	0,005	0,93	2369	24	2361	12	2409	21	98
59	133755	13626	307	0,29	0,1019	0,0012	3,846	0,052	0,283	0,004	0,93	1608	18	1603	11	1659	22	97
60	34329	3745	69	0,93	0,1091	0,0014	4,832	0,065	0,320	0,004	0,91	1790	19	1791	11	1784	23	100
61	15194	1739	31	0,57	0,1145	0,0020	4,879	0,090	0,320	0,005	0,78	1790	23	1799	16	1871	33	96
62	52194	7008	90	0,39	0,1343	0,0015	6,526	0,085	0,374	0,005	0,97	2050	22	2049	11	2155	20	95
63	47036	6203	77	0,33	0,1319	0,0015	6,681	0,085	0,381	0,005	0,96	2081	22	2070	11	2123	20	98
64	46967	6603	71	0,69	0,1406	0,0016	7,548	0,097	0,404	0,005	0,94	2188	22	2179	11	2234	21	98

65	29236	4037	47	0,50	0,1381	0,0016	7,186	0,094	0,393	0,005	0,95	2135	23	2135	12	2203	21	97
66	33282	4614	51	0,23	0,1386	0,0017	7,446	0,100	0,401	0,005	0,91	2174	23	2167	12	2210	22	98
67	27927	3969	48	0,45	0,1421	0,0020	7,081	0,113	0,390	0,005	0,89	2121	25	2122	14	2253	26	94
68	43877	6069	70	0,42	0,1383	0,0015	7,268	0,093	0,395	0,005	0,96	2147	22	2145	11	2206	20	97
76	23278	3183	38	0,40	0,1367	0,0016	6,972	0,093	0,388	0,005	0,94	2113	23	2108	12	2186	21	97
77	52433	7659	71	0,38	0,1461	0,0036	8,702	0,206	0,431	0,006	0,64	2309	29	2307	22	2300	41	100
78	14936	2117	21	0,23	0,1417	0,0034	7,919	0,182	0,413	0,007	0,71	2228	31	2222	21	2249	41	99
79	31947	4352	47	0,47	0,1362	0,0022	7,378	0,121	0,399	0,005	0,79	2166	24	2158	15	2180	29	99
80	29742	3233	65	0,85	0,1087	0,0015	4,489	0,067	0,309	0,004	0,89	1734	20	1729	12	1778	25	98
81	71837	9567	113	0,33	0,1332	0,0014	7,180	0,088	0,410	0,005	0,99	2217	23	2134	11	2140	19	104
82	45476	6311	72	0,23	0,1388	0,0015	7,318	0,094	0,397	0,005	0,96	2157	23	2151	11	2212	20	98
83	46386	4631	106	0,37	0,0998	0,0011	3,673	0,048	0,278	0,003	0,94	1582	17	1566	11	1621	22	98
84	84817	10896	145	0,75	0,1285	0,0014	6,235	0,078	0,366	0,004	0,97	2011	21	2009	11	2077	20	97
85	43837	5912	72	0,42	0,1349	0,0015	7,073	0,091	0,391	0,005	0,96	2127	22	2121	11	2162	20	98
87	44104	6056	73	0,71	0,1373	0,0017	7,211	0,101	0,397	0,005	0,93	2157	24	2138	13	2194	22	98
88	28791	3816	45	0,50	0,1325	0,0028	7,357	0,148	0,400	0,006	0,72	2168	27	2156	18	2132	36	102
89	20512	2781	31	0,43	0,1356	0,0029	7,156	0,151	0,393	0,006	0,73	2136	28	2131	19	2171	38	98
90	45130	6363	64	0,33	0,1410	0,0022	7,896	0,125	0,413	0,005	0,80	2228	24	2219	14	2239	27	99
98	49235	9146	66	0,52	0,1858	0,0030	13,015	0,227	0,517	0,008	0,84	2686	32	2681	16	2705	27	99
99	52914	10061	65	0,58	0,1901	0,0021	12,972	0,164	0,515	0,006	0,97	2678	27	2678	12	2743	19	98
100	21891	3036	36	0,40	0,1387	0,0017	7,098	0,097	0,390	0,005	0,93	2122	23	2124	12	2211	22	96
101	43098	5284	75	0,49	0,1226	0,0019	6,241	0,097	0,366	0,005	0,81	2009	22	2010	14	1994	27	101
102	37634	5017	55	0,40	0,1333	0,0029	6,827	0,150	0,383	0,006	0,67	2091	26	2089	20	2142	39	98
103	58938	9557	87	0,42	0,1622	0,0031	9,913	0,201	0,459	0,007	0,77	2435	32	2427	19	2478	33	98
104	51020	7370	81	0,33	0,1445	0,0019	7,827	0,115	0,411	0,005	0,90	2220	25	2211	13	2281	23	97
105	72920	9989	113	0,63	0,1370	0,0017	7,100	0,095	0,391	0,005	0,90	2127	22	2124	12	2189	22	97
106	48507	6537	72	0,37	0,1348	0,0022	7,286	0,121	0,396	0,005	0,77	2151	23	2147	15	2161	29	100
107	47131	6401	72	0,79	0,1358	0,0019	7,143	0,105	0,393	0,005	0,84	2135	22	2129	13	2174	25	98
108	18109	3008	25	0,57	0,1661	0,0023	10,588	0,156	0,473	0,006	0,88	2497	27	2488	14	2519	24	99
109	52286	7364	79	0,57	0,1408	0,0018	7,768	0,105	0,412	0,005	0,90	2222	23	2204	12	2238	22	99

Sample TFC-02					Isotope ratios ^b								Ages (Ma)					
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/	±	²⁰⁷ Pb*/	±	²⁰⁶ Pb*	±	RH	²⁰⁶ Pb/	±	²⁰⁷ Pb	±	²⁰⁶ Pb/	±	Conc
number	cps	cps	ppm	Th/U _a	²⁰⁶ Pb*	1s	²³⁵ U	1s	/	²³⁸ U	1s	²³⁸ U	1s	²³⁵ U	1s	²⁰⁷ Pb	1s	
1	75324	10004	699	0,28	0,1328	0,0014	6,878	0,076	0,384	0,004	0,87	2096	17	2096	10	2135	22	98
2	167571	22384	1565	0,15	0,1336	0,0011	6,938	0,065	0,385	0,004	0,98	2102	17	2104	8	2146	18	98
3	113354	28881	2019	0,23	0,2548	0,0013	21,501	0,197	0,647	0,006	0,92	3215	24	3162	9	3215	16	100
5	190548	34896	2440	0,20	0,1831	0,0012	12,894	0,120	0,529	0,005	0,90	2739	21	2672	9	2681	17	102
4	72939	10130	708	0,22	0,1389	0,0012	7,615	0,075	0,406	0,004	0,95	2196	17	2187	9	2213	19	99
11	109392	14863	1039	0,36	0,1359	0,0012	7,262	0,072	0,396	0,004	0,94	2149	17	2145	9	2175	20	99
6	61759	8312	581	0,09	0,1346	0,0012	7,237	0,072	0,399	0,004	0,95	2163	17	2141	9	2159	19	100
10	120668	15702	1098	0,25	0,1301	0,0012	6,616	0,063	0,378	0,004	0,97	2069	16	2062	8	2100	19	99
8	124737	16916	1183	0,21	0,1356	0,0012	5,407	0,052	0,323	0,003	1,02	1803	15	1886	8	2172	20	83
19	168282	22105	1545	0,11	0,1314	0,0012	6,822	0,064	0,384	0,004	0,97	2096	16	2089	8	2116	19	99
15	79986	11136	779	0,28	0,1392	0,0014	6,603	0,071	0,379	0,004	0,95	2073	18	2060	9	2218	22	93
16	180303	34061	2381	0,20	0,1889	0,0012	12,808	0,117	0,512	0,005	0,91	2666	20	2666	9	2733	17	98
17	95564	13001	909	0,39	0,1360	0,0013	7,287	0,083	0,396	0,004	0,90	2152	19	2147	10	2177	21	99

20	53483	7681	537	0,24	0,1436	0,0012	7,454	0,072	0,399	0,004	0,98	2166	17	2168	9	2271	19	95
24	116615	15625	1092	0,18	0,1340	0,0012	7,054	0,069	0,389	0,004	0,95	2117	17	2119	9	2151	19	98
25	129044	16245	1136	0,37	0,1259	0,0012	6,299	0,061	0,369	0,003	0,95	2026	16	2019	8	2041	19	99
28	224780	55563	3885	0,29	0,2472	0,0013	20,904	0,194	0,628	0,006	0,99	3141	23	3135	9	3167	16	99
26	69355	6959	487	0,23	0,1003	0,0013	3,684	0,037	0,275	0,003	0,94	1568	13	1568	8	1630	21	96
21	118755	15850	1108	0,26	0,1335	0,0014	7,191	0,086	0,395	0,004	0,86	2144	19	2136	11	2144	23	100
22	92680	12777	893	0,27	0,1379	0,0013	7,531	0,080	0,403	0,004	0,91	2183	18	2177	9	2201	20	99
18	156871	22101	1545	0,30	0,1409	0,0012	7,901	0,072	0,430	0,004	0,91	2305	18	2220	8	2238	18	103
13	58028	7565	529	0,21	0,1304	0,0012	6,729	0,069	0,382	0,004	0,93	2086	17	2077	9	2103	20	99
14	124671	17241	1205	0,35	0,1383	0,0014	7,511	0,083	0,401	0,004	0,86	2175	18	2175	10	2206	22	99
23	151972	25394	1775	0,19	0,1671	0,0013	10,682	0,121	0,474	0,005	0,90	2500	21	2496	10	2529	20	99
12	166055	22075	1543	0,16	0,1329	0,0013	7,572	0,087	0,418	0,004	0,87	2252	19	2182	10	2137	22	105
29	125172	17399	1216	0,21	0,1390	0,0017	7,571	0,109	0,406	0,005	0,80	2196	21	2182	13	2215	27	99
33	117796	16289	1139	0,17	0,1383	0,0012	7,656	0,076	0,410	0,004	0,93	2216	17	2192	9	2206	20	100
30	64157	8950	626	0,34	0,1395	0,0012	7,517	0,077	0,400	0,004	0,94	2171	18	2175	9	2221	20	98
31	87008	8607	602	0,43	0,0989	0,0013	3,682	0,037	0,277	0,003	0,94	1577	13	1568	8	1604	21	98

Sample TFC-04					Isotope ratios ^b								Ages (Ma)					
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/	±	²⁰⁷ Pb*/	±	²⁰⁶ Pb*	±	RH	²⁰⁶ Pb/	±	²⁰⁷ Pb	±	²⁰⁶ Pb/	±	Conc
number	cps	cps	ppm	Th/U ^a	²⁰⁶ Pb*	1s	²³⁵ U	1s	²³⁸ U	1s	O	²³⁸ U	1s	²³⁵ U	1s	²⁰⁷ Pb	1s	
13	71174	8408	127	0,50	0,1181	0,0014	5,488	0,071	0,344	0,004	0,94	1905	20	1899	11	1928	22	99
14	55416	7921	71	0,41	0,1429	0,0037	8,173	0,201	0,418	0,006	0,61	2251	28	2250	22	2263	59	99
15	40032	3870	86	0,41	0,0967	0,0017	3,598	0,063	0,273	0,004	0,74	1558	18	1549	14	1561	27	100
16	82119	10606	124	0,35	0,1292	0,0018	6,529	0,096	0,373	0,004	0,82	2043	21	2050	13	2086	30	98
17	12958	1663	20	1,07	0,1283	0,0028	6,480	0,137	0,372	0,006	0,71	2039	26	2043	19	2075	45	98
18	53154	7268	87	0,50	0,1367	0,0028	7,220	0,152	0,394	0,006	0,73	2143	28	2139	19	2186	44	98
19	72478	19848	72	0,39	0,2738	0,0037	22,951	0,355	0,648	0,009	0,89	3221	35	3225	15	3328	45	97
20	38392	4232	69	0,59	0,1102	0,0018	4,684	0,077	0,314	0,004	0,76	1759	19	1764	14	1803	29	98
22	87077	11642	136	0,47	0,1337	0,0022	6,886	0,117	0,384	0,005	0,77	2093	23	2097	15	2147	35	97
30	101240	11060	217	0,54	0,1092	0,0017	4,242	0,073	0,295	0,004	0,78	1665	20	1682	14	1787	28	93
31	66061	13370	69	0,80	0,2024	0,0043	15,612	0,324	0,559	0,008	0,67	2861	32	2853	20	2846	60	101
32	49486	9579	54	0,50	0,1936	0,0040	14,256	0,292	0,535	0,007	0,68	2761	31	2767	19	2773	58	100
33	36324	4012	76	0,49	0,1105	0,0016	4,255	0,066	0,292	0,004	0,82	1650	19	1685	13	1807	26	91
34	60205	6557	115	0,38	0,1089	0,0020	4,334	0,080	0,300	0,004	0,70	1690	19	1700	15	1781	32	95
36	51525	5136	110	0,62	0,0997	0,0022	3,779	0,082	0,277	0,004	0,64	1575	20	1588	17	1618	36	97
37	76362	14171	93	0,59	0,1856	0,0028	12,353	0,201	0,502	0,006	0,79	2622	28	2632	15	2703	41	97
38	33362	3282	72	0,40	0,0984	0,0023	3,660	0,084	0,272	0,004	0,63	1553	20	1563	18	1594	37	97
39	30381	3053	65	0,53	0,1005	0,0021	3,617	0,075	0,271	0,004	0,66	1547	19	1553	16	1633	33	95

Sample TFC-05					Isotope ratios ^b								Ages (Ma)					
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/	±	²⁰⁷ Pb*/	±	²⁰⁶ Pb*	±	RH	²⁰⁶ Pb/	±	²⁰⁷ Pb	±	²⁰⁶ Pb/	±	Conc
number	cps	cps	ppm	Th/U ^a	²⁰⁶ Pb*	1s	²³⁵ U	1s	²³⁸ U	1s	O	²³⁸ U	1s	²³⁵ U	1s	²⁰⁷ Pb	1s	
13	90186	24355	92	0,67	0,2701	0,0030	24,137	0,306	0,659	0,008	0,98	3263	32	3274	12	3306	36	99
14	28259	3837	49	0,67	0,1358	0,0018	7,230	0,104	0,394	0,005	0,91	2141	24	2140	13	2174	28	98

15	16211	1967	36	0,71	0,1213	0,0028	6,411	0,141	0,368	0,006	0,77	2019	29	2034	19	1976	46	102
16	37851	5243	66	1,53	0,1385	0,0016	7,346	0,099	0,394	0,005	0,94	2140	23	2154	12	2209	26	97
17	19695	2496	37	0,27	0,1267	0,0017	6,147	0,091	0,363	0,005	0,90	1996	23	1997	13	2053	28	97
18	54088	7355	97	0,25	0,1360	0,0024	7,127	0,133	0,390	0,006	0,81	2123	27	2127	17	2177	39	98
19	50401	9031	69	0,75	0,1792	0,0023	12,038	0,170	0,498	0,007	0,93	2604	28	2607	13	2645	33	98
20	42510	5703	74	0,42	0,1342	0,0016	7,367	0,094	0,395	0,005	0,95	2147	22	2157	11	2153	25	100
21	8520	1843	10	0,03	0,2163	0,0035	16,818	0,278	0,573	0,008	0,89	2920	35	2925	16	2953	48	99
22	47122	8050	62	0,42	0,1708	0,0021	11,101	0,143	0,481	0,006	0,93	2530	25	2532	12	2566	31	99
23	43475	7819	58	0,81	0,1799	0,0023	12,263	0,169	0,502	0,006	0,93	2622	28	2625	13	2651	33	99
24	38743	10965	38	0,93	0,2830	0,0031	25,885	0,322	0,679	0,008	0,99	3340	32	3342	12	3380	37	99
25	36247	4507	73	0,52	0,1243	0,0019	5,808	0,097	0,351	0,005	0,85	1939	24	1948	15	2019	31	96
26	107365	19838	147	0,38	0,1848	0,0022	12,478	0,174	0,507	0,007	0,95	2646	29	2641	13	2696	32	98
27	29122	4063	49	0,25	0,1395	0,0016	7,407	0,097	0,395	0,005	0,95	2146	23	2162	12	2221	26	97
35	5614	747	10	0,50	0,1331	0,0025	6,849	0,132	0,383	0,006	0,76	2088	26	2092	17	2139	40	98
36	90903	24679	94	0,43	0,2715	0,0028	23,913	0,296	0,657	0,008	1,00	3254	32	3265	12	3315	34	98
37	26356	3320	47	0,50	0,1260	0,0018	6,454	0,094	0,370	0,005	0,86	2031	22	2040	13	2042	29	99
38	64147	10593	91	0,60	0,1651	0,0024	10,618	0,165	0,471	0,006	0,87	2487	28	2490	14	2509	37	99
39	103279	18785	133	0,58	0,1819	0,0020	12,477	0,150	0,506	0,006	0,97	2639	25	2641	11	2670	29	99
40	33766	4328	62	0,70	0,1282	0,0022	7,423	0,124	0,399	0,005	0,82	2165	25	2164	15	2073	36	104
41	22489	4795	24	0,38	0,2132	0,0040	16,185	0,296	0,563	0,009	0,83	2878	35	2888	17	2930	55	98
42	106370	28294	124	0,84	0,2660	0,0033	22,338	0,323	0,638	0,009	0,95	3180	35	3199	14	3283	40	97
43	19040	2607	29	0,31	0,1369	0,0026	7,247	0,135	0,393	0,006	0,77	2138	26	2142	17	2189	41	98
44	43508	9984	50	0,28	0,2295	0,0024	17,970	0,223	0,589	0,007	1,00	2985	30	2988	12	3048	32	98
46	48524	6124	137	0,30	0,1262	0,0016	3,779	0,051	0,222	0,003	0,89	1293	14	1588	11	2046	26	63
47	56278	13320	65	0,39	0,2367	0,0034	18,975	0,298	0,602	0,009	0,92	3038	35	3041	15	3098	44	98
48	6175	687	13	0,73	0,1113	0,0022	4,550	0,093	0,310	0,004	0,71	1742	22	1740	17	1820	35	96
49	18616	2051	40	1,25	0,1102	0,0015	4,542	0,065	0,308	0,004	0,88	1729	19	1739	12	1802	24	96
50	49338	9105	69	0,35	0,1845	0,0020	12,030	0,159	0,498	0,007	0,99	2606	28	2607	12	2694	30	97
51	20787	3658	30	0,25	0,1760	0,0027	11,210	0,186	0,482	0,007	0,89	2535	31	2541	15	2615	40	97
52	33839	4535	58	0,72	0,1340	0,0016	6,891	0,090	0,383	0,005	0,94	2090	22	2098	12	2151	25	97
60	39260	4983	78	0,56	0,1269	0,0028	6,070	0,138	0,361	0,006	0,74	1985	29	1986	20	2056	46	97
61	15499	2132	24	0,43	0,1376	0,0034	7,292	0,171	0,392	0,006	0,71	2134	30	2148	21	2197	53	97
62	13627	1878	23	0,33	0,1378	0,0019	7,292	0,107	0,394	0,005	0,89	2142	24	2148	13	2200	30	97
63	33604	5699	51	0,89	0,1696	0,0021	10,089	0,144	0,460	0,006	0,94	2441	27	2443	13	2554	31	96
64	24130	6627	23	0,44	0,2746	0,0034	24,560	0,325	0,663	0,008	0,96	3278	33	3291	13	3333	41	98
65	24687	3410	43	0,32	0,1381	0,0018	7,268	0,105	0,394	0,005	0,91	2141	24	2145	13	2204	29	97
66	22173	2893	39	0,29	0,1305	0,0017	6,407	0,089	0,367	0,005	0,90	2014	22	2033	12	2104	27	96
67	11791	1608	20	0,34	0,1364	0,0030	7,169	0,155	0,389	0,006	0,75	2120	29	2133	19	2182	48	97
68	100650	27409	100	0,34	0,2723	0,0029	23,923	0,290	0,654	0,008	0,98	3243	30	3265	12	3319	35	98
69	57480	14881	66	0,11	0,2589	0,0029	20,665	0,283	0,622	0,008	0,99	3116	34	3123	13	3240	36	96
70	61002	10924	80	0,52	0,1791	0,0020	11,906	0,147	0,495	0,006	0,97	2593	26	2597	12	2644	29	98
71	48884	13446	46	0,50	0,2751	0,0041	24,498	0,371	0,664	0,009	0,87	3281	34	3288	15	3335	50	98
72	42665	7261	57	0,44	0,1702	0,0021	11,067	0,144	0,481	0,006	0,93	2530	25	2529	12	2559	31	99
73	38959	5092	68	0,57	0,1307	0,0020	6,530	0,106	0,372	0,005	0,83	2039	23	2050	14	2107	33	97
75	27721	3726	51	0,68	0,1344	0,0018	6,756	0,101	0,379	0,005	0,90	2073	24	2080	13	2156	29	96
83	28557	3686	53	1,19	0,1291	0,0016	6,810	0,090	0,381	0,005	0,94	2083	22	2087	12	2086	25	100
84	40497	7371	54	0,32	0,1820	0,0020	12,101	0,153	0,498	0,006	0,98	2606	26	2612	12	2671	29	98
85	14233	1972	21	0,38	0,1386	0,0041	7,537	0,213	0,402	0,008	0,67	2176	35	2177	25	2209	66	99
86	53062	7588	87	0,39	0,1430	0,0021	7,902	0,126	0,413	0,006	0,85	2228	26	2220	14	2264	34	98

87	27688	5175	34	0,74	0,1869	0,0025	12,889	0,178	0,513	0,006	0,90	2669	27	2672	13	2715	36	98
89	52593	13707	55	0,46	0,2606	0,0030	22,336	0,296	0,641	0,008	0,96	3193	32	3199	13	3250	38	98
91	55728	12414	62	0,48	0,2228	0,0025	17,373	0,217	0,581	0,007	0,96	2951	28	2956	12	3001	34	98
93	45916	5745	86	0,50	0,1251	0,0015	6,032	0,084	0,361	0,005	0,92	1988	22	1980	12	2030	25	98
94	21187	2927	33	0,41	0,1382	0,0030	7,313	0,153	0,395	0,006	0,73	2145	28	2150	19	2204	47	97
95	38238	5292	66	0,47	0,1384	0,0016	7,462	0,097	0,396	0,005	0,95	2152	22	2168	12	2207	26	98
96	12979	1835	19	0,27	0,1414	0,0034	7,725	0,182	0,407	0,007	0,71	2200	31	2200	21	2244	55	98
97	13999	1812	22	0,31	0,1294	0,0042	6,521	0,199	0,373	0,007	0,64	2044	34	2049	27	2090	67	98
105	114618	16061	190	0,39	0,1401	0,0018	7,574	0,107	0,405	0,005	0,89	2191	23	2182	13	2229	28	98
106	88910	11689	158	0,55	0,1315	0,0020	6,458	0,102	0,369	0,005	0,82	2023	23	2040	14	2118	32	96
107	52021	9596	61	0,46	0,1845	0,0040	12,799	0,268	0,512	0,008	0,71	2666	33	2665	20	2693	58	99
108	50842	6835	85	0,69	0,1344	0,0016	6,973	0,090	0,387	0,005	0,93	2109	22	2108	11	2157	25	98
109	32717	4276	57	0,35	0,1307	0,0016	6,514	0,086	0,372	0,005	0,92	2039	21	2048	12	2107	25	97
110	64554	8423	117	0,63	0,1305	0,0015	6,563	0,084	0,373	0,005	0,95	2042	21	2054	11	2104	24	97
111	67992	9015	120	0,59	0,1326	0,0015	6,720	0,086	0,379	0,005	0,95	2074	22	2075	11	2133	24	97
112	64822	8058	122	0,41	0,1243	0,0014	5,909	0,078	0,356	0,004	0,94	1965	21	1963	11	2019	23	97
113	46521	5394	106	0,82	0,1159	0,0017	5,133	0,081	0,329	0,005	0,88	1836	22	1842	13	1895	27	97
114	66267	8889	117	0,41	0,1341	0,0021	6,836	0,112	0,382	0,005	0,83	2085	24	2090	15	2153	33	97
116	31976	6197	38	0,92	0,1938	0,0034	14,022	0,242	0,532	0,007	0,79	2752	31	2751	16	2775	49	99
125	17380	2186	30	1,20	0,1257	0,0027	6,911	0,135	0,384	0,006	0,73	2095	26	2100	17	2039	43	103
126	60204	7440	114	1,25	0,1236	0,0023	6,235	0,115	0,367	0,005	0,76	2014	24	2009	16	2009	37	100
127	48026	5946	93	0,66	0,1238	0,0018	5,837	0,092	0,353	0,005	0,85	1950	22	1952	14	2012	29	97
128	33110	10536	27	0,63	0,3182	0,0071	31,736	0,689	0,732	0,012	0,78	3541	46	3542	21	3561	80	99
129	60854	13702	67	0,45	0,2252	0,0026	17,507	0,221	0,582	0,007	0,94	2957	28	2963	12	3018	35	98
130	90209	13050	148	0,44	0,1447	0,0017	7,906	0,106	0,410	0,005	0,93	2216	23	2220	12	2284	27	97
144	41720	5328	61	0,46	0,1277	0,0016	6,222	0,086	0,365	0,005	0,90	2007	22	2008	12	2067	26	97
145	51415	10438	49	1,15	0,2030	0,0035	14,980	0,260	0,546	0,008	0,81	2808	32	2814	17	2851	49	99
146	33217	6367	35	0,45	0,1917	0,0023	13,301	0,181	0,520	0,007	0,95	2701	29	2701	13	2757	33	98
147	16354	2092	24	0,79	0,1279	0,0018	6,098	0,094	0,360	0,005	0,86	1981	23	1990	13	2070	30	96
148	5147	537	9	0,50	0,1044	0,0027	5,049	0,117	0,326	0,005	0,69	1821	25	1828	20	1704	44	107
149	14220	3019	14	0,59	0,2123	0,0028	15,633	0,228	0,556	0,007	0,93	2850	31	2855	14	2923	39	97
150	30048	4779	37	2,31	0,1591	0,0019	9,556	0,129	0,449	0,006	0,94	2391	26	2393	12	2446	29	98
151	22586	2609	36	0,46	0,1155	0,0018	5,034	0,081	0,325	0,004	0,82	1816	21	1825	14	1888	29	96
152	48056	8313	55	0,45	0,1730	0,0021	11,223	0,151	0,482	0,006	0,92	2538	26	2542	13	2587	32	98
153	54375	10301	60	0,46	0,1894	0,0020	12,635	0,164	0,509	0,007	0,99	2654	28	2653	12	2737	29	97
154	42070	7843	48	0,35	0,1864	0,0034	12,882	0,245	0,512	0,008	0,83	2667	34	2671	18	2711	49	98
156	60626	11437	69	0,45	0,1886	0,0027	13,276	0,210	0,520	0,007	0,89	2698	31	2700	15	2730	39	99
157	29484	5202	34	0,33	0,1764	0,0023	11,734	0,166	0,491	0,006	0,90	2576	27	2583	13	2620	34	98
158	19081	2540	26	0,12	0,1331	0,0022	6,796	0,115	0,382	0,005	0,80	2088	24	2085	15	2139	35	98
159	18885	1797	42	0,57	0,0952	0,0027	3,372	0,095	0,261	0,005	0,64	1494	24	1498	22	1531	44	98
160	31220	5733	32	0,54	0,1836	0,0035	12,094	0,232	0,498	0,007	0,78	2606	32	2612	18	2686	51	97

Sample TFC-06		Isotope ratios ^b										Ages (Ma)						
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/	±	²⁰⁷ Pb*/	±	²⁰⁶ Pb*	±	RH	²⁰⁶ Pb/	±	²⁰⁷ Pb	±	²⁰⁶ Pb/	±	Conc
number	cps	cps	ppm	Th/U _a	²⁰⁶ Pb*	1s	²³⁵ U	1s	²³⁸ U	1s	O	²³⁸ U	1s	²³⁵ U	1s	²⁰⁷ Pb	1s	
18	59269	6877	123	0,64	0,1160	0,0022	4,800	0,095	0,320	0,005	0,76	1792	24	1785	17	1896	36	94
19	59157	7261	101	0,62	0,1227	0,0013	6,014	0,072	0,361	0,004	0,96	1986	20	1978	10	1996	22	99

20	34685	4530	56	0,54	0,1306	0,0016	6,823	0,089	0,384	0,005	0,92	2095	21	2089	12	2106	25	99
21	75902	9339	129	1,00	0,1230	0,0013	6,052	0,072	0,363	0,004	0,97	1995	20	1983	10	2001	21	100
22	42668	4178	94	0,60	0,0979	0,0011	3,688	0,046	0,279	0,003	0,92	1584	16	1569	10	1585	18	100
23	138144	28240	174	0,31	0,2044	0,0022	13,508	0,178	0,528	0,007	0,99	2731	29	2716	12	2862	31	95
24	67479	12195	80	0,40	0,1807	0,0020	12,311	0,145	0,504	0,006	0,96	2630	25	2628	11	2660	29	99
25	16646	1620	36	0,45	0,0973	0,0019	3,506	0,068	0,267	0,004	0,68	1526	18	1529	15	1573	31	97
26	17066	1844	37	0,33	0,1081	0,0052	4,479	0,207	0,307	0,008	0,57	1726	40	1727	38	1767	86	98
35	67684	7328	134	0,41	0,1083	0,0011	4,509	0,053	0,313	0,004	0,98	1758	18	1733	10	1770	18	99
36	65372	17583	64	0,46	0,2690	0,0048	23,626	0,434	0,655	0,010	0,87	3247	41	3253	18	3300	59	98
37	44489	5451	74	0,70	0,1225	0,0016	6,075	0,084	0,367	0,004	0,88	2015	21	1987	12	1993	26	101
38	84445	14357	121	0,27	0,1700	0,0018	10,550	0,130	0,469	0,006	0,99	2479	25	2484	11	2558	27	97
40	116073	12742	232	0,41	0,1098	0,0014	4,798	0,067	0,321	0,004	0,89	1795	19	1785	12	1796	23	100
41	26243	2615	60	0,89	0,0996	0,0013	3,648	0,050	0,275	0,003	0,88	1568	17	1560	11	1617	20	97
42	62009	5985	134	0,40	0,0965	0,0011	3,660	0,045	0,280	0,003	0,94	1594	16	1563	10	1558	17	102
43	99854	12690	170	0,22	0,1271	0,0013	6,238	0,074	0,363	0,004	0,97	1998	20	2010	10	2058	22	97
44	59701	5954	139	0,46	0,0997	0,0011	3,515	0,042	0,266	0,003	0,97	1519	16	1531	10	1619	17	94
52	59039	5955	138	0,36	0,1009	0,0013	3,763	0,054	0,277	0,003	0,88	1575	18	1585	11	1640	21	96
53	113453	28955	103	0,37	0,2552	0,0028	21,867	0,257	0,638	0,007	0,95	3180	28	3178	11	3217	36	99
54	75095	9238	124	0,42	0,1230	0,0014	6,019	0,072	0,364	0,004	0,95	1999	19	1979	10	2000	22	100
55	111119	20106	133	0,45	0,1809	0,0019	12,224	0,141	0,502	0,006	0,97	2623	24	2622	11	2662	28	99
56	62417	11208	82	0,70	0,1796	0,0024	12,012	0,177	0,496	0,007	0,90	2598	28	2605	14	2649	36	98
57	68328	7622	131	0,52	0,1116	0,0013	4,950	0,064	0,329	0,004	0,93	1832	19	1811	11	1825	21	100
58	99111	14916	148	0,54	0,1505	0,0030	8,870	0,177	0,432	0,007	0,76	2315	29	2325	18	2352	47	98
59	61800	8058	111	1,24	0,1304	0,0016	6,722	0,091	0,384	0,005	0,94	2093	23	2076	12	2103	25	100
61	74910	9570	136	0,56	0,1278	0,0020	6,214	0,102	0,364	0,005	0,83	2003	24	2006	14	2067	32	97
69	45666	8745	52	0,53	0,1915	0,0022	13,495	0,164	0,525	0,006	0,95	2720	26	2715	11	2755	31	99
70	35648	3939	72	0,55	0,1105	0,0016	4,755	0,073	0,319	0,004	0,84	1786	20	1777	13	1808	26	99
71	30919	5340	38	0,57	0,1727	0,0021	11,193	0,145	0,483	0,006	0,92	2539	25	2539	12	2584	32	98
72	104686	18603	130	0,40	0,1777	0,0018	11,792	0,138	0,495	0,006	0,98	2591	25	2588	11	2632	27	98
73	52961	5690	103	0,48	0,1074	0,0012	4,481	0,056	0,311	0,004	0,92	1744	18	1728	10	1756	20	99
74	45544	5766	77	0,62	0,1266	0,0014	6,201	0,077	0,365	0,004	0,95	2007	20	2005	11	2051	23	98
75	66827	7380	128	0,76	0,1104	0,0012	4,753	0,058	0,321	0,004	0,95	1795	18	1777	10	1807	20	99
76	66151	8678	102	0,43	0,1312	0,0016	6,865	0,085	0,387	0,004	0,91	2110	20	2094	11	2114	25	100
77	37880	4986	63	0,68	0,1316	0,0019	6,913	0,108	0,391	0,005	0,86	2129	24	2100	14	2120	31	100
78	20174	2507	33	0,94	0,1243	0,0024	6,931	0,122	0,390	0,005	0,76	2123	24	2103	16	2018	38	105
86	92699	12877	139	0,17	0,1389	0,0020	7,589	0,111	0,405	0,005	0,84	2194	23	2184	13	2214	31	99
87	95555	15370	143	0,37	0,1608	0,0020	9,947	0,140	0,459	0,006	0,91	2436	26	2430	13	2465	31	99
88	69590	14277	70	0,57	0,2052	0,0045	15,165	0,323	0,552	0,008	0,68	2834	33	2826	20	2868	63	99
89	64010	6296	135	0,19	0,0984	0,0012	3,608	0,047	0,275	0,003	0,87	1568	16	1551	10	1593	20	98
90	52080	5070	112	0,31	0,0974	0,0012	3,677	0,048	0,282	0,003	0,89	1601	17	1567	11	1574	19	102
91	34767	4724	57	0,26	0,1359	0,0016	6,986	0,093	0,387	0,005	0,93	2110	22	2110	12	2175	26	97
92	40712	5270	75	0,44	0,1294	0,0018	6,157	0,095	0,360	0,005	0,87	1981	23	1998	13	2090	29	95
93	50867	6743	86	0,34	0,1326	0,0018	6,797	0,100	0,383	0,005	0,87	2093	23	2085	13	2132	29	98
94	63520	8186	107	0,37	0,1289	0,0016	6,957	0,089	0,389	0,005	0,93	2119	21	2106	11	2083	25	102
103	60189	7935	101	0,54	0,1318	0,0016	6,709	0,090	0,382	0,005	0,93	2085	22	2074	12	2123	25	98
104	26420	2683	65	0,35	0,1016	0,0018	3,594	0,067	0,265	0,004	0,75	1515	19	1548	15	1653	30	92
105	57650	6055	115	0,48	0,1050	0,0013	4,683	0,058	0,321	0,004	0,93	1794	18	1764	10	1715	20	105
106	66237	7583	130	0,65	0,1145	0,0018	5,076	0,083	0,328	0,004	0,80	1829	21	1832	14	1872	29	98
107	38370	3800	91	0,36	0,0990	0,0012	3,437	0,044	0,261	0,003	0,91	1493	16	1513	10	1606	19	93

108	49298	5040	104	0,66	0,1022	0,0015	3,811	0,057	0,279	0,003	0,80	1588	17	1595	12	1665	24	95
109	98677	11893	205	0,24	0,1205	0,0022	4,983	0,101	0,325	0,005	0,75	1816	24	1817	17	1964	36	92
110	52738	6896	83	0,51	0,1308	0,0016	6,801	0,088	0,388	0,005	0,91	2114	21	2086	11	2108	25	100

Sample TFC-07					Isotope ratios ^b								Ages (Ma)					
Spot number	Pb206*	Pb207	U		²⁰⁷ Pb*/ ²⁰⁶ Pb*		²⁰⁷ Pb*/ ²³⁵ U		²⁰⁶ Pb*/ ²³⁸ U		RH O	²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²³⁵ U		²⁰⁶ Pb/ ²⁰⁷ Pb		Conc
	cps	cps	ppm	Th/U _a	1s	1s	1s	1s	1s	1s		1s	1s	1s	1s	1s		
58	108888	14438	1009	0,17	0,1326	0,0013	6,899	0,074	0,385	0,004	0,92	2098	18	2099	9	2133	20	98
52	90916	11552	808	0,24	0,1271	0,0012	6,371	0,063	0,371	0,003	0,95	2035	16	2028	9	2058	20	99
51	76056	11052	773	0,53	0,1453	0,0012	8,714	0,086	0,446	0,004	0,96	2379	19	2309	9	2291	19	104
57	36531	5086	356	0,29	0,1392	0,0013	7,660	0,084	0,407	0,004	0,90	2201	18	2192	10	2218	21	99
62	65578	9604	671	0,25	0,1465	0,0012	8,520	0,084	0,437	0,004	0,96	2338	19	2288	9	2305	19	101
61	91175	12065	843	0,32	0,1323	0,0012	7,076	0,070	0,397	0,004	0,95	2154	17	2121	9	2129	19	101
60	96439	13673	956	0,31	0,1418	0,0012	7,401	0,068	0,408	0,004	0,91	2205	17	2161	8	2249	19	98
59	60848	7859	549	0,70	0,1292	0,0012	6,613	0,066	0,383	0,004	0,94	2089	17	2061	9	2087	20	100
53	94060	12515	875	0,20	0,1331	0,0012	6,778	0,067	0,377	0,004	0,95	2064	16	2083	9	2139	19	96
54	53646	8923	624	0,18	0,1663	0,0014	11,088	0,123	0,494	0,005	0,89	2587	21	2531	10	2521	21	103
55	81497	11753	822	0,51	0,1442	0,0012	8,535	0,085	0,438	0,004	0,94	2343	18	2290	9	2278	19	103
56	41321	4641	324	0,68	0,1123	0,0013	4,802	0,052	0,317	0,003	0,89	1774	15	1785	9	1837	22	97
50	68815	9280	649	0,31	0,1349	0,0012	7,219	0,073	0,400	0,004	0,95	2167	18	2139	9	2162	20	100
49	39117	4321	302	0,51	0,1105	0,0014	4,783	0,054	0,320	0,003	0,86	1789	15	1782	10	1807	23	99
46	40017	5483	383	0,20	0,1370	0,0013	7,157	0,075	0,391	0,004	0,93	2126	18	2131	9	2190	20	97
48	77380	10362	724	0,45	0,1339	0,0012	7,063	0,071	0,391	0,004	0,94	2126	17	2119	9	2150	20	99
47	83400	10755	752	0,17	0,1290	0,0016	6,479	0,082	0,375	0,004	0,80	2055	18	2043	11	2084	25	99
44	49401	6735	471	0,29	0,1363	0,0012	6,867	0,069	0,383	0,004	0,95	2090	17	2094	9	2181	20	96
41	63762	8587	600	0,23	0,1347	0,0012	7,086	0,072	0,390	0,004	0,93	2123	17	2122	9	2160	20	98
45	95276	12724	890	0,25	0,1335	0,0012	7,119	0,071	0,395	0,004	0,94	2145	17	2126	9	2145	19	100
43	90424	12353	864	0,28	0,1366	0,0013	7,648	0,082	0,414	0,004	0,91	2234	18	2191	10	2185	20	102
42	81430	10704	748	0,23	0,1315	0,0012	6,810	0,067	0,389	0,004	0,96	2117	17	2087	9	2117	19	100
27	78916	10362	724	0,30	0,1313	0,0012	6,880	0,068	0,393	0,004	0,96	2136	17	2096	9	2115	19	101
28	79951	13434	939	0,12	0,1680	0,0012	10,660	0,104	0,474	0,004	0,96	2500	20	2494	9	2538	18	99
29	35268	5030	352	0,27	0,1426	0,0013	7,122	0,074	0,389	0,004	0,95	2117	18	2127	9	2259	21	94
32	75010	16982	1187	0,00	0,2264	0,0014	17,998	0,193	0,590	0,006	0,91	2988	23	2990	10	3027	19	99
30	82016	11024	771	0,32	0,1344	0,0015	7,126	0,088	0,391	0,004	0,84	2127	19	2127	11	2156	23	99
31	57124	7654	535	0,50	0,1340	0,0012	7,064	0,073	0,392	0,004	0,93	2131	17	2119	9	2151	20	99
17	39811	5112	357	0,24	0,1284	0,0014	6,531	0,078	0,376	0,004	0,84	2059	18	2050	10	2076	23	99
26	51019	4764	333	0,20	0,0934	0,0014	3,121	0,034	0,248	0,002	0,88	1428	12	1438	8	1496	23	95
40	79056	10615	742	0,31	0,1343	0,0014	7,409	0,083	0,407	0,004	0,87	2200	18	2162	10	2155	22	102
37	78119	11466	802	0,26	0,1468	0,0012	8,594	0,086	0,434	0,004	0,94	2325	18	2296	9	2309	19	101
35	51220	6776	474	0,25	0,1323	0,0018	7,048	0,110	0,393	0,005	0,75	2136	21	2117	14	2129	30	100
33	59553	7882	551	0,25	0,1324	0,0015	7,140	0,092	0,398	0,004	0,82	2160	19	2129	12	2129	25	101
34	79758	10329	722	0,35	0,1295	0,0013	6,625	0,069	0,383	0,004	0,93	2091	17	2063	9	2091	20	100
36	80921	11727	820	0,40	0,1449	0,0015	7,430	0,091	0,400	0,004	0,88	2169	20	2165	11	2287	24	95
38	82927	11138	779	0,43	0,1343	0,0016	7,201	0,095	0,395	0,004	0,78	2147	19	2137	12	2155	26	100
39	43441	5762	403	0,30	0,1326	0,0013	6,866	0,076	0,383	0,004	0,88	2091	17	2094	10	2133	22	98
14	52324	5678	397	0,42	0,1085	0,0016	4,511	0,059	0,308	0,003	0,81	1733	16	1733	11	1775	26	98
18	27044	3772	264	0,20	0,1395	0,0015	7,533	0,093	0,399	0,004	0,83	2166	19	2177	11	2221	24	98

9	45377	6391	447	0,41	0,1408	0,0013	7,579	0,080	0,400	0,004	0,91	2170	18	2182	10	2238	20	97
1	54104	9883	691	0,28	0,1827	0,0015	12,809	0,156	0,517	0,005	0,86	2684	23	2666	12	2677	22	100
10	76362	12145	849	0,51	0,1590	0,0012	9,958	0,101	0,464	0,004	0,94	2459	19	2431	9	2446	19	101
11	67783	9474	662	0,37	0,1398	0,0012	7,575	0,077	0,403	0,004	0,94	2185	18	2182	9	2224	20	98
2	115016	15576	1089	0,35	0,1354	0,0013	7,567	0,084	0,412	0,004	0,88	2224	18	2181	10	2169	21	103
20	67807	9200	643	0,20	0,1357	0,0013	7,109	0,074	0,388	0,004	0,91	2115	17	2125	9	2173	20	97
21	54105	6562	459	0,39	0,1213	0,0013	5,653	0,061	0,352	0,003	0,91	1944	16	1924	9	1975	21	98
12	52614	6897	482	0,28	0,1311	0,0016	7,124	0,095	0,403	0,004	0,81	2183	20	2126	12	2113	25	103
4	72148	9678	677	0,28	0,1341	0,0014	7,169	0,083	0,395	0,004	0,84	2148	18	2132	10	2153	23	100
3	55619	7130	498	0,26	0,1282	0,0016	6,703	0,089	0,385	0,004	0,79	2099	19	2073	12	2073	26	101
13	90598	12137	849	0,31	0,1340	0,0014	7,454	0,085	0,411	0,004	0,86	2218	18	2167	10	2151	22	103
22	53338	5930	415	0,30	0,1112	0,0014	4,761	0,052	0,318	0,003	0,88	1780	15	1778	9	1819	22	98
23	42132	6740	471	0,39	0,1600	0,0016	9,855	0,125	0,457	0,005	0,82	2425	21	2421	12	2455	24	99
24	49087	6350	444	0,20	0,1294	0,0013	6,627	0,072	0,380	0,004	0,89	2076	17	2063	10	2089	21	99
14	93943	9269	648	0,36	0,0987	0,0016	3,728	0,047	0,279	0,003	0,78	1586	14	1577	10	1599	27	99
5	117142	16608	1161	0,19	0,1418	0,0013	6,602	0,066	0,377	0,004	0,98	2062	17	2059	9	2249	20	92
25	140569	19367	1354	0,32	0,1378	0,0016	7,501	0,100	0,401	0,004	0,79	2175	20	2172	12	2199	25	99
15	27771	3665	256	0,32	0,1320	0,0014	6,817	0,082	0,383	0,004	0,84	2088	18	2087	11	2124	23	98
16	71714	9644	674	0,43	0,1345	0,0015	7,136	0,092	0,391	0,004	0,81	2130	19	2128	11	2157	25	99
6	34331	3790	265	0,86	0,1104	0,0015	4,763	0,057	0,320	0,003	0,84	1789	16	1778	10	1806	24	99
7	95026	13432	939	0,23	0,1414	0,0017	7,651	0,108	0,406	0,005	0,80	2195	21	2191	13	2244	26	98
8	22540	3485	244	0,47	0,1546	0,0014	8,745	0,102	0,432	0,004	0,89	2315	20	2311	11	2398	22	97

Sample TFC-08					Isotope ratios ^b								Ages (Ma)					
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/ ²⁰⁶ Pb*		²⁰⁷ Pb*/ ²³⁵ U		²⁰⁶ Pb*/ ²³⁸ U		RH	²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²³⁵ U		²⁰⁶ Pb/ ²⁰⁷ Pb		Conc
number	cps	cps	ppm	Th/U _a	1s	1s	1s	1s	1s	1s	O	1s	1s	1s	1s	1s	1s	
2	118544	16640	1103	0,37	0,1404	0,0012	8,035	0,081	0,417	0,004	0,92	2246	18	2235	9	2232	19	101
3	184438	25330	1678	0,32	0,1373	0,0012	7,676	0,079	0,410	0,004	0,95	2216	18	2194	9	2194	19	101
4	50745	8989	596	0,15	0,1771	0,0014	10,647	0,130	0,469	0,005	0,93	2477	23	2493	11	2626	21	94
6	211695	27105	1796	0,43	0,1280	0,0011	6,669	0,064	0,382	0,004	0,97	2086	17	2069	8	2071	18	101
17	263680	33643	2229	0,16	0,1276	0,0012	7,372	0,076	0,405	0,004	0,91	2082	17	2074	9	2065	20	101
7	64943	9512	630	1,02	0,1465	0,0017	8,271	0,117	0,420	0,005	0,80	2262	22	2262	13	2305	26	98
1	114957	16213	1074	0,22	0,1410	0,0012	7,515	0,080	0,406	0,004	0,95	2195	19	2175	9	2240	19	98

Sample TFC-10					Isotope ratios ^b								Ages (Ma)					
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/ ²⁰⁶ Pb*		²⁰⁷ Pb*/ ²³⁵ U		²⁰⁶ Pb*/ ²³⁸ U		RH	²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²³⁵ U		²⁰⁶ Pb/ ²⁰⁷ Pb		Conc
number	cps	cps	ppm	Th/U _a	1s	1s	1s	1s	1s	1s	O	1s	1s	1s	1s	1s	1s	
16	68862	9814	101	0,21	0,1425	0,0021	7,799	0,126	0,408	0,006	0,88	2206	26	2208	15	2258	34	98
17	36826	4204	63	0,50	0,1141	0,0013	4,980	0,064	0,325	0,004	0,94	1814	19	1816	11	1866	22	97
19	28217	4076	38	0,22	0,1445	0,0018	7,933	0,106	0,412	0,005	0,94	2224	24	2223	12	2281	28	97
20	12009	1566	18	0,87	0,1304	0,0023	7,515	0,123	0,403	0,005	0,83	2181	25	2175	15	2104	36	104
21	28507	3712	40	0,19	0,1302	0,0024	6,432	0,115	0,371	0,005	0,78	2033	24	2037	16	2101	38	97
22	47519	6193	74	0,31	0,1303	0,0017	6,507	0,093	0,373	0,005	0,92	2046	23	2047	13	2102	27	97
23	21982	2506	39	0,42	0,1140	0,0014	4,871	0,068	0,320	0,004	0,90	1790	20	1797	12	1864	24	96

24	58746	8183	89	0,57	0,1393	0,0016	7,175	0,096	0,392	0,005	0,97	2133	24	2133	12	2218	25	96
25	28189	3866	42	0,50	0,1371	0,0018	7,888	0,104	0,410	0,005	0,93	2217	23	2218	12	2192	28	101
26	35200	4982	48	0,37	0,1415	0,0017	7,758	0,100	0,408	0,005	0,94	2206	23	2203	12	2246	27	98
28	29369	5843	29	0,72	0,1990	0,0030	14,259	0,220	0,536	0,007	0,90	2768	31	2767	15	2818	42	98
29	26321	3593	41	0,40	0,1365	0,0019	6,883	0,107	0,383	0,005	0,90	2090	25	2096	14	2183	30	96
30	60837	8497	89	0,38	0,1397	0,0017	7,499	0,104	0,401	0,005	0,94	2174	24	2173	12	2223	27	98
38	41046	5931	57	0,69	0,1445	0,0017	7,895	0,105	0,410	0,005	0,95	2214	24	2219	12	2282	27	97
39	44298	6122	58	0,54	0,1382	0,0019	7,246	0,101	0,394	0,005	0,88	2139	22	2142	12	2205	30	97
40	23994	3010	38	0,41	0,1254	0,0020	7,155	0,105	0,390	0,005	0,86	2124	23	2131	13	2035	32	104
41	19441	2711	29	0,27	0,1394	0,0019	7,262	0,108	0,393	0,005	0,91	2137	24	2144	13	2220	30	96
43	39578	5595	52	0,88	0,1414	0,0017	7,692	0,100	0,407	0,005	0,92	2203	22	2196	12	2244	27	98
44	30560	4248	41	0,48	0,1390	0,0020	7,320	0,107	0,395	0,005	0,86	2145	23	2151	13	2215	32	97
45	73140	12660	78	0,02	0,1731	0,0022	10,976	0,144	0,477	0,006	0,91	2516	25	2521	12	2588	33	97
46	30444	4299	45	0,25	0,1412	0,0017	8,117	0,103	0,416	0,005	0,96	2243	23	2244	12	2242	26	100
48	15089	1889	26	0,67	0,1252	0,0019	6,927	0,106	0,385	0,005	0,87	2100	24	2102	14	2032	31	103
49	20323	2868	28	0,88	0,1411	0,0018	7,630	0,104	0,405	0,005	0,93	2193	24	2188	12	2241	28	98
50	16284	2484	20	0,64	0,1525	0,0032	8,612	0,175	0,426	0,007	0,78	2287	31	2298	19	2375	49	96
51	11287	1519	16	1,27	0,1346	0,0026	6,877	0,132	0,383	0,006	0,77	2091	26	2096	17	2159	42	97
60	122306	17346	174	0,93	0,1418	0,0015	7,646	0,095	0,405	0,005	0,99	2193	23	2190	11	2250	24	97
62	33154	4725	41	0,33	0,1425	0,0030	7,719	0,160	0,406	0,006	0,75	2197	29	2199	19	2258	48	97
63	37984	5837	52	0,50	0,1537	0,0018	8,744	0,120	0,428	0,006	0,95	2298	25	2312	12	2387	28	96
64	26270	3412	38	0,53	0,1299	0,0034	7,909	0,179	0,410	0,007	0,71	2214	30	2221	20	2096	55	106
66	54059	7560	80	0,12	0,1398	0,0017	7,480	0,105	0,401	0,005	0,94	2175	24	2171	13	2225	27	98
67	34655	4788	47	0,59	0,1382	0,0024	7,494	0,132	0,402	0,006	0,81	2179	26	2172	16	2204	38	99
68	135856	22577	171	0,39	0,1662	0,0018	10,238	0,127	0,462	0,006	0,99	2449	25	2457	12	2520	27	97
69	25550	2642	51	0,19	0,1034	0,0012	3,883	0,052	0,282	0,003	0,92	1602	17	1610	11	1686	20	95
72	72775	9833	108	0,20	0,1351	0,0014	6,958	0,086	0,387	0,005	0,99	2108	22	2106	11	2165	23	97
73	15664	2169	22	0,38	0,1385	0,0025	7,180	0,131	0,390	0,006	0,80	2124	27	2134	16	2208	40	96
81	51239	7261	74	0,57	0,1417	0,0016	7,745	0,100	0,407	0,005	0,97	2203	23	2202	12	2248	25	98
87	28389	3790	38	0,41	0,1335	0,0024	6,684	0,120	0,378	0,005	0,78	2065	25	2071	16	2145	39	96
88	54444	7576	78	0,34	0,1392	0,0016	7,294	0,094	0,394	0,005	0,96	2140	22	2148	11	2217	25	97
89	36825	6939	35	0,72	0,1884	0,0042	13,349	0,282	0,519	0,009	0,78	2696	36	2705	20	2729	61	99
90	27750	3699	40	0,35	0,1333	0,0021	7,164	0,111	0,391	0,005	0,84	2127	23	2132	14	2142	34	99
91	31964	4913	47	0,31	0,1537	0,0023	7,974	0,133	0,413	0,006	0,89	2228	28	2228	15	2388	35	93

Sample TFC-11					Isotope ratios ^b							Ages (Ma)						
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/	±	²⁰⁷ Pb*/	±	²⁰⁶ Pb*	±	RH	²⁰⁶ Pb/	±	²⁰⁷ Pb	±	²⁰⁶ Pb/	±	Conc
number	cps	cps	ppm	Th/U ^a	²⁰⁶ Pb*	1s	²³⁵ U	1s	²³⁸ U	1s	O	²³⁸ U	1s	²³⁵ U	1s	²⁰⁷ Pb	1s	
1	114811	15563	1031	0,10	0,1356	0,0013	7,417	0,078	0,402	0,004	0,90	2178	18	2163	9	2171	20	100
2	137356	22782	1510	0,38	0,1659	0,0013	10,439	0,115	0,464	0,005	0,93	2458	21	2475	10	2516	19	98

Sample TFC-12					Isotope ratios ^b							Ages (Ma)						
Spot	Pb206*	Pb207	U		²⁰⁷ Pb*/	±	²⁰⁷ Pb*/	±	²⁰⁶ Pb*	±	RH	²⁰⁶ Pb/	±	²⁰⁷ Pb	±	²⁰⁶ Pb/	±	Conc
number	cps	cps	ppm	Th/U ^a	²⁰⁶ Pb*	1s	²³⁵ U	1s	²³⁸ U	1s	O	²³⁸ U	1s	²³⁵ U	1s	²⁰⁷ Pb	1s	

16	84817	11269	113	0,42	0,1329	0,0019	6,742	0,097	0,380	0,005	0,84	2075	21	2078	13	2136	30	97
17	90530	11575	145	0,55	0,1279	0,0014	6,193	0,080	0,364	0,005	0,98	2003	22	2004	11	2069	22	97
18	125175	28205	108	0,51	0,2253	0,0030	17,599	0,241	0,584	0,007	0,88	2965	29	2968	13	3019	40	98
19	196829	34068	251	1,12	0,1731	0,0017	10,592	0,131	0,469	0,006	1,02	2479	26	2488	11	2588	25	96
20	151964	29778	175	0,27	0,1960	0,0020	13,194	0,167	0,517	0,007	1,02	2687	28	2694	12	2793	28	96
21	63228	8389	94	0,65	0,1327	0,0014	6,841	0,084	0,383	0,005	0,99	2091	22	2091	11	2134	22	98
22	65775	8850	96	0,68	0,1345	0,0014	6,894	0,085	0,384	0,005	0,99	2095	22	2098	11	2158	23	97
23	81275	10612	128	0,24	0,1306	0,0015	6,532	0,089	0,373	0,005	0,95	2043	23	2050	12	2106	24	97
24	112066	21374	117	0,70	0,1907	0,0021	12,887	0,157	0,512	0,006	0,97	2666	26	2671	11	2748	30	97
25	161709	29467	186	0,11	0,1822	0,0023	11,983	0,168	0,496	0,006	0,93	2598	28	2603	13	2673	33	97
26	83346	11293	123	0,41	0,1355	0,0014	6,954	0,086	0,385	0,005	0,99	2099	22	2106	11	2170	23	97
27	72166	13383	76	0,46	0,1854	0,0021	12,348	0,151	0,504	0,006	0,96	2629	25	2631	12	2702	30	97
28	168728	22738	272	0,18	0,1348	0,0014	6,567	0,083	0,374	0,005	1,01	2049	22	2055	11	2161	22	95
29	219447	28504	333	0,26	0,1299	0,0013	6,380	0,075	0,368	0,004	1,01	2021	21	2030	10	2096	21	96
30	81452	10370	136	0,79	0,1273	0,0018	6,167	0,097	0,365	0,005	0,89	2005	24	2000	14	2061	28	97
38	69065	12242	80	0,38	0,1773	0,0020	11,511	0,151	0,487	0,006	0,96	2556	27	2566	12	2627	30	97
39	125090	23771	142	0,38	0,1900	0,0022	13,363	0,184	0,521	0,007	0,95	2704	29	2706	13	2742	32	99
40	64615	8506	96	0,59	0,1316	0,0017	6,700	0,094	0,381	0,005	0,91	2080	23	2073	12	2120	27	98
41	41044	5561	57	0,42	0,1355	0,0017	6,920	0,093	0,384	0,005	0,90	2096	22	2101	12	2170	28	97
42	61773	10377	75	0,63	0,1680	0,0028	10,545	0,184	0,469	0,007	0,84	2480	30	2484	16	2538	42	98
43	63489	13048	59	0,39	0,2055	0,0031	15,078	0,228	0,547	0,007	0,84	2812	29	2820	14	2871	43	98
44	77397	15266	82	0,42	0,1972	0,0023	14,029	0,190	0,533	0,007	0,95	2753	29	2752	13	2804	33	98
45	93067	11714	144	0,78	0,1259	0,0015	6,088	0,080	0,363	0,004	0,94	1994	21	1989	11	2041	24	98
46	169497	21921	249	0,23	0,1293	0,0014	6,291	0,075	0,368	0,004	0,98	2019	20	2017	10	2089	22	97
47	12725	1796	17	0,73	0,1411	0,0022	7,479	0,121	0,399	0,005	0,84	2165	25	2170	14	2241	35	97
48	14404	1788	24	0,47	0,1242	0,0020	6,613	0,106	0,376	0,005	0,86	2058	24	2061	14	2017	32	102
49	261002	49199	291	0,46	0,1885	0,0019	12,824	0,155	0,510	0,006	1,00	2659	26	2667	11	2729	27	97
57	107321	11864	192	0,77	0,1105	0,0012	4,513	0,058	0,308	0,004	0,94	1731	18	1734	11	1808	20	96
58	97460	19647	94	1,54	0,2016	0,0023	14,319	0,179	0,538	0,006	0,93	2774	26	2771	12	2839	33	98
59	92081	9905	180	0,44	0,1076	0,0015	4,363	0,067	0,300	0,004	0,86	1694	20	1705	13	1759	24	96
60	88369	8672	192	0,37	0,0981	0,0011	3,447	0,045	0,263	0,003	0,96	1507	17	1515	10	1589	18	95
61	64149	8274	87	0,50	0,1290	0,0021	6,453	0,103	0,371	0,005	0,78	2035	22	2040	14	2084	33	98
62	180461	22560	305	1,34	0,1250	0,0015	6,128	0,085	0,363	0,005	0,94	1994	22	1994	12	2029	24	98
63	49136	6147	71	0,51	0,1251	0,0017	5,888	0,084	0,355	0,004	0,85	1958	20	1960	12	2030	28	96
64	44397	5751	59	0,64	0,1295	0,0026	6,516	0,127	0,372	0,005	0,71	2037	24	2048	17	2092	42	97
65	35177	4579	51	0,79	0,1302	0,0016	6,332	0,084	0,367	0,004	0,91	2016	21	2023	12	2100	26	96
66	51817	11628	48	0,61	0,2244	0,0025	17,106	0,213	0,576	0,007	0,96	2933	28	2941	12	3013	34	97
67	81583	13520	99	0,51	0,1657	0,0017	10,277	0,126	0,464	0,006	0,99	2459	25	2460	11	2515	26	98
69	63874	12427	76	0,17	0,1946	0,0024	13,759	0,199	0,527	0,007	0,93	2728	30	2733	14	2781	35	98
70	117212	14793	189	0,49	0,1262	0,0014	6,059	0,078	0,360	0,005	0,97	1983	21	1984	11	2046	22	97
71	206462	34983	256	0,41	0,1694	0,0019	10,816	0,144	0,474	0,006	0,96	2502	26	2507	12	2552	29	98
72	134848	21657	188	0,36	0,1606	0,0020	8,966	0,133	0,435	0,006	0,92	2330	27	2335	14	2462	31	95
73	13222	1866	18	0,25	0,1411	0,0020	7,497	0,114	0,399	0,005	0,88	2165	24	2173	14	2241	32	97
74	102124	11133	193	0,21	0,1090	0,0012	4,544	0,057	0,309	0,004	0,98	1734	19	1739	10	1783	19	97
82	14775	1976	22	0,78	0,1337	0,0019	6,741	0,103	0,379	0,005	0,86	2070	23	2078	14	2148	31	96
83	71361	9196	101	0,52	0,1289	0,0017	6,205	0,085	0,364	0,004	0,86	2002	20	2005	12	2083	27	96
84	8116	873	13	0,50	0,1076	0,0039	4,375	0,153	0,301	0,006	0,60	1698	31	1708	29	1759	64	97
85	58473	15122	49	0,34	0,2586	0,0030	21,639	0,275	0,634	0,008	0,94	3164	30	3168	12	3238	38	98
86	28786	3940	42	0,33	0,1369	0,0017	7,067	0,097	0,388	0,005	0,92	2113	23	2120	12	2188	27	97

87	1548	295	2	0,14	0,1906	0,0079	13,078	0,523	0,515	0,014	0,70	2677	61	2685	38	2747	114	97
92	12836	1714	19	0,52	0,1335	0,0025	6,427	0,124	0,369	0,005	0,77	2025	26	2036	17	2145	40	94

^a Th/U ratios are calculated relative to GJ-1 reference zircon

^b Corrected for background and within-run Pb/U fractionation and normalised to reference zircon GJ-1 (ID-TIMS values/measured value)

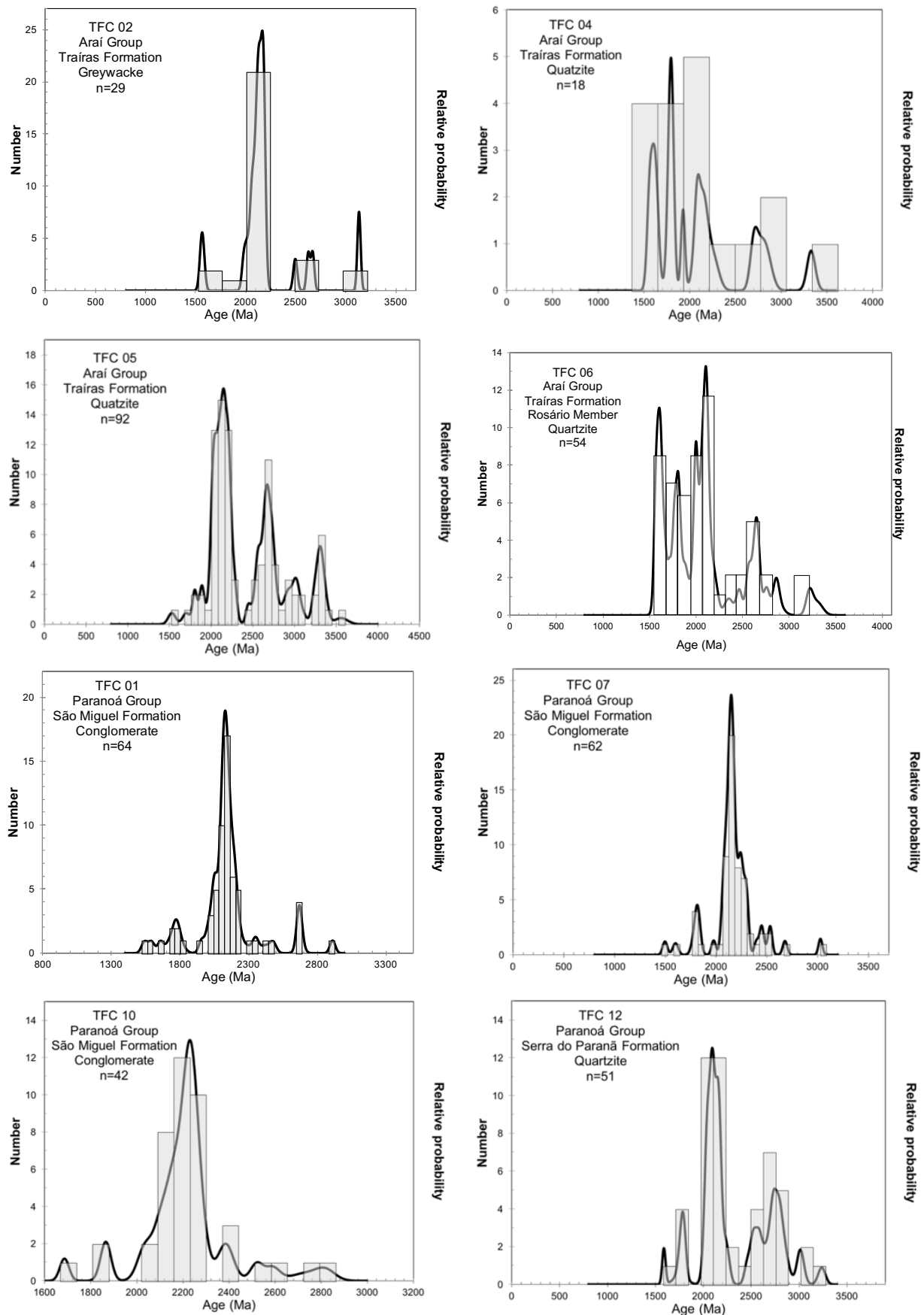


Figure 4 – Histogram and relative probability plots for samples TFC 01 to TFC 12.

Zircon distribution pattern shows dominant Rhyacian source and minor contribution of Statherian, Calymmian, Neoarchean and Paleoproterozoic sources for both the upper Araí (Traíras Formation) and Paranoá groups. The youngest interval is Calymmian with age peak around 1.58 Ga, evidencing provenance from the Serra da Mesa Suite anorogenic granite that is regionally correlated to the Middle Espinhaço magmatism. The second youngest interval is Statherian with age peak at ca. 1.8 Ga, evidencing provenance from the Pedra Branca Suite anorogenic granite, that is regionally correlated to the Lower Espinhaço magmatism. The third and most abundant age interval is Rhyacian and can be correlated to the intense orogenic magmatism, either called Transamazonian Orogeny (Almeida, 1981, Alkmim and Marshak, 1998), that also took place in the Northern Brasília Belt and is mainly represented by the Aurumina Suite.

Nevertheless, most Rhyacian zircon grains have positive $\epsilon_{\text{Hf}}(t)$ signature, which are rarely present in the Aurumina Suite (Cuadros et al. 2017a), suggesting that this important contribution comes mainly from juvenile Paleoproterozoic terranes, most probably from the Goiás Massif, especially the Campinorte domain, located west of the study area, where positive $\epsilon_{\text{Nd}}(t)$ values are common (Giustina et al. 2009a, Fuck et al. 2014). Possible sources from juvenile terranes further to the east are not discarded, though less probable.

Minor contribution of Neoarchean to Paleoproterozoic zircons is observed. Neoarchean and Paleoproterozoic present predominantly negative $\epsilon_{\text{Hf}}(t)$ values, whereas ca. Mesoarchean show some positive results. Positive $\epsilon_{\text{Nd}}(t)$ values from +2 to +3 $\epsilon_{\text{Nd}}(t)$ are found in Mesoarchean ca. 3.0 Ga rocks and 2.8 Ga rocks varying from slightly negative to slightly positive (-1 to 1) $\epsilon_{\text{Hf}}(t)$ in the Goiás Massif Crixás-Goiás Domain, as reported by Borges et al. (2017) and Beghelli Jr (2012), respectively, indicating that the oldest portion of Goiás Massif could be at least one of the Archean zircon sources. Nonetheless, predominant negative $\epsilon_{\text{Hf}}(t)$ values found for Archean zircons require evolved Archean terrains, which can be found south of the São Francisco Craton, at the Quadrilátero Ferrífero (Farina et al. 2016), a possible source for these zircons found in the Paranoá and Traíras sequences.

The ten youngest concordant zircon grains from the Traíras Formation (upper Araí Group) were plotted in a Concordia diagram (Figure 7) and reveal a maximum depositional age of 1543 ± 31 Ma for this marine sequence, finally confirming the assumption that the Upper Araí and the Serra da Mesa groups are chronocorrelated. The data allows to advance on the tectono-stratigraphic understanding by constraining Proterozoic extensional pulses and permitting a more direct correlation to other sequences in the São Francisco Craton and its margins.

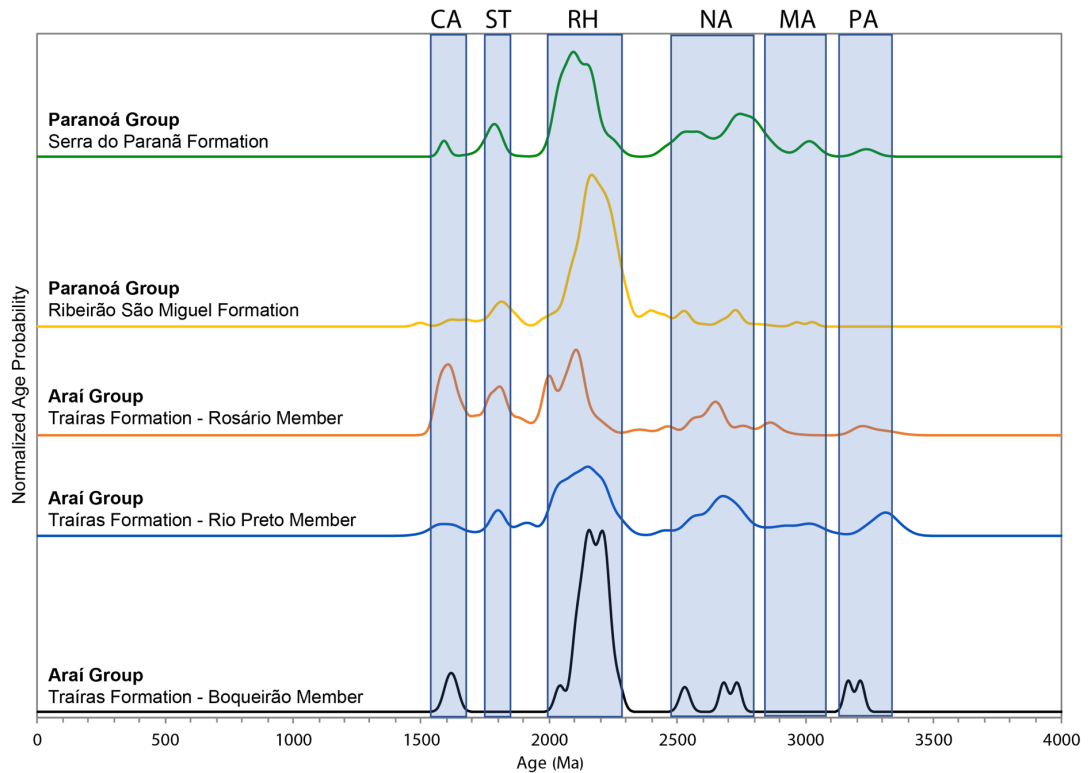


Figure 5 – Normalized age probability plot for all U-Pb data grouped by formation. Main age intervals are identified as: CA = Calymmian; ST = Statherian, RH = Rhyacian, NA = Neoproterozoic, MA = Mesoarchoean; PA = Paleoarchoean.

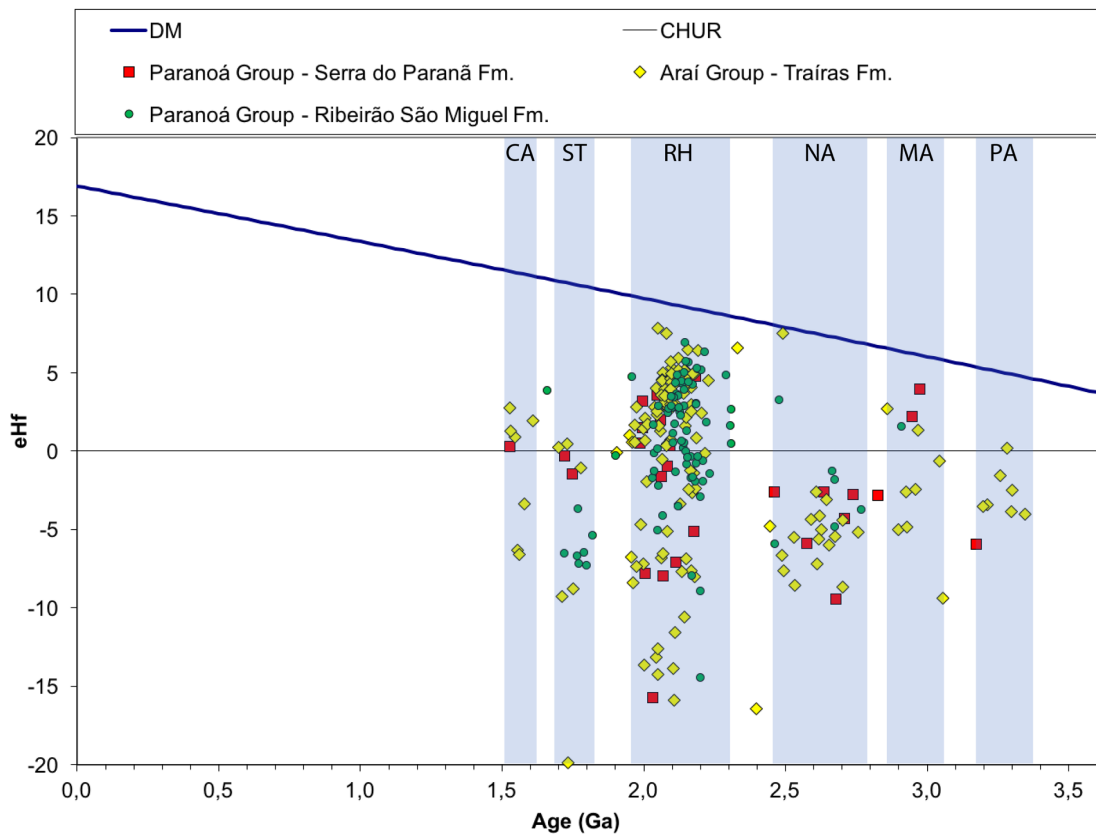


Figure 6 – U-Pb ages versus $\epsilon_{\text{Hf}}(t)$ results plot of detrital zircons for all samples analyzed, distinguished by group and formation. CA = Calymmian; ST = Statherian, RH = Rhyacian, NA = Neoproterozoic, MA = Mesoarchoean; PA = Paleoarchoean.

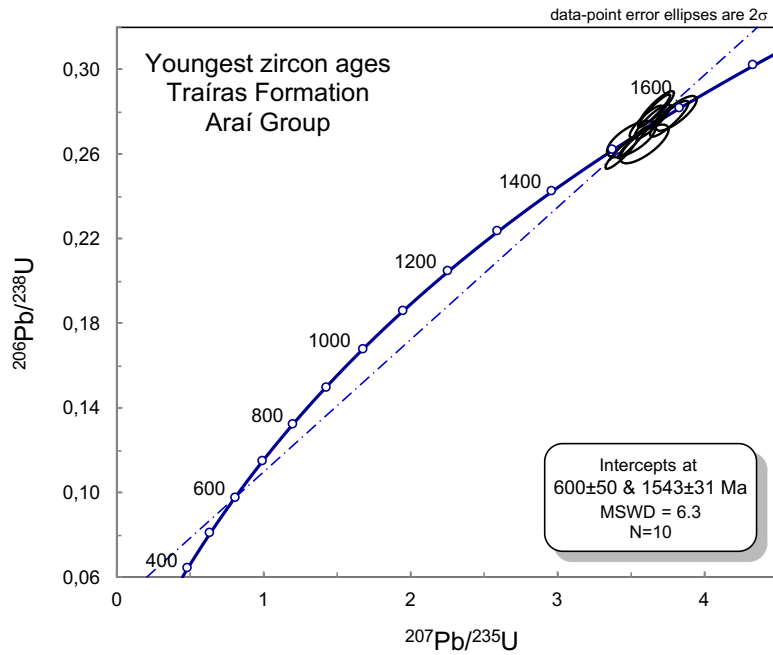


Figure 7 – Concordia diagram for the youngest zircon grains from Traíras Formation, Araí Group.

Based on the presented data and the stratigraphic elucidation it provides, we rescue the early concept of Braun (1980) to separate the rift and sag basins in different groups. According to the Brazilian Stratigraphic Nomenclature Guide (Petri et al. 1986a, 1986b), we propose to raise the Traíras formation to group status, separating it, thus, from the Araí Group, that historically represents the Statherian rift sequence. The Traíras Group, therefore, represents an entirely different Calymmian sag sequence and its previous members can be raised to formation status (Figure 8).

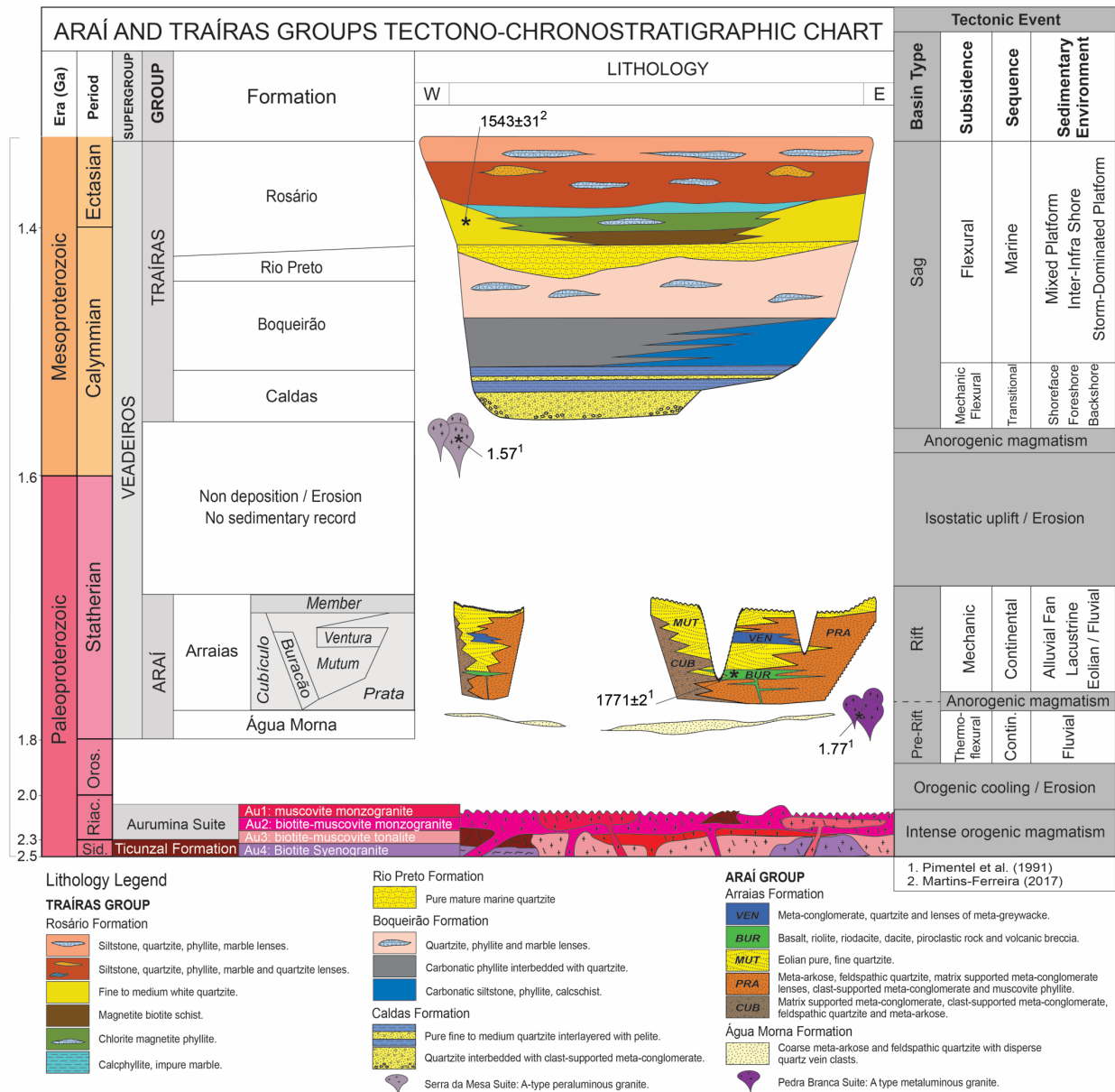


Figure 8 – Tectono-stratigraphic chart Araí and Traíras groups. regional correlation chart of the proposed Veadeiros Supergroup with the Espinhaço Supergroup from Statherian to Tonian Periods, showing basin type, depositional environments, unconformities, formal lithostratigraphic hierarchical units, first-order sequences and main global tectonic events (modified from Tanizaki et al. 2015).

Several sedimentary successions, either spanned by short or long time intervals between each other, are grouped in supergroups as to evidence a certain degree of relationship, be linked tectonic regimes, isostatic cycles, or simply geographical location. Supergroups generally encompass different types of basins formed in a long period of time, spanning from one era up to an entire eon. Examples are the Paleo- to Neoproterozoic Vindhyan Supergroup (Chakraborty and Bhattacharyya 1996) and the Palaeo- to Neoproterozoic Espinhaço Supergroup (Chemale et al. 2012; Alkmim and Martins-Neto 2012).

Aiming to facilitate regional correlation with the other São Francisco Craton sedimentary sequences, we propose the encompassing of all three Proterozoic intracontinental extensional pulses in the western margin as the Veadeiros Supergroup, respectively chronocorrelated to the lower, middle and upper Espinhaço Supergroup first order sequences (Figure 9). The Veadeiros Supergroup represents the Paleo- to Neoproterozoic extensional basin successions in the Northern Brasília Belt, it includes the Araí, Traíras and Paranoá groups and may include the groups interpreted as their lateral correlatives, Serra da Mesa, Natividade and Canastra groups.

The Neoproterozoic Bambuí Group should not be included in the Veadeiros Supergroup for two reasons: first, it represents a much wider basin, not restricted to the Brasília belt, secondly (and mainly), because there is evidence that it was (at least part of it) controlled by the Brasiliano orogeny compressive tectonics, thus diverging from the purely extensional events represented by the Veadeiros Supergroup sequences.

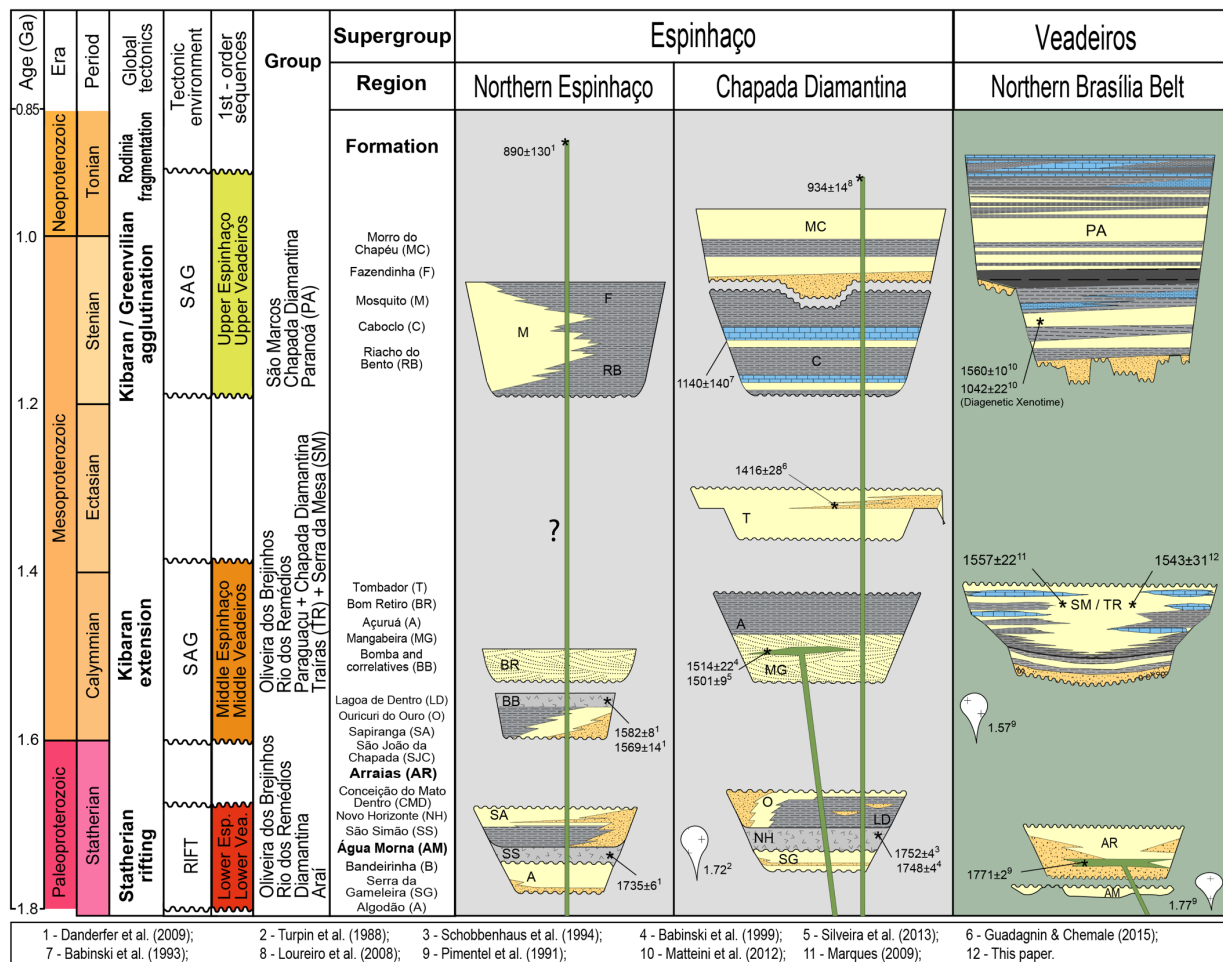


Figure 9 - Chrono-stratigraphic chart confronting the Veadeiros Supergroup sequences with its correlatives from the Espinhaço Supergroup, modified from Guadagnin and Chemale (2015).

Zircons coeval or from the Serra da Mesa Suite are present in the Traíras and Serra da mesa sequences but at the same time show contact metamorphism with Serra da Mesa basal sediments. The characterization of these sequences as formed in intracontinental sag type basins can shed light to this geologic chicken-and-egg problem. Intracontinental basins are usually long lasting and have a complicated subsidence history, frequently containing significant unconformities capable of eroding and reworking its own previous deposits. The assumption admits the hypothesis that the lower portion of Serra da Mesa Group might have been intruded by the Serra da Mesa granite and erosion in a major unconformity might have supplied these zircons to the upper Serra da Mesa and Traíras groups.

The new data suggest that the Traíras and Serra da Mesa groups might represent the same basin system, first because they both contain zircons from the Serra da Mesa Granite, second because the Goiás Massif is at the same time is the basement for the Mesoproterozoic Serra da Mesa Group and source of Rhyacian positive $\epsilon_{\text{Hf}}(t)$ zircons for the Mesoproterozoic Traíras Group. These assumptions claim for a further assertion, that plays against established beliefs: The Goiás Massif cannot be an exotic block accreted to the São Francisco Craton during the Neoproterozoic (Brito Neves and Cordani, 1991; Fuck et al. 1994, 2017; Pimentel et al. 2000; 2004; 2011; Blum et al. 2003; Valeriano et al. 2008; Queiroz et al. 2008; Ferreira Filho et al. 2010), but would more likely have been accreted earlier, probably during the Rhyacian orogeny. Cordeiro and Oliveira (2017) also advocate that hypothesis based on other sound geologic and geochronological data.

This new geotectonic model for the Goiás Massif invariably forces the São Francisco paleocontinent western limit at 1.8 Ga to coincide with the Goiás Massif western limit and locally even a little beyond, considering that the Goiás Magmatic Arc is covering the continental paleoplate borders by thrusts in both northern and southern Brasília Belt segments. According to all the topics discussed above, we are compelled to propose a more accurate São Francisco paleocontinent western limit at ca. 1.8 Ga, outlined in red solid line in figure 10, dashed lines represent inferred limits under the Goiás Magmatic Arc thrusts.

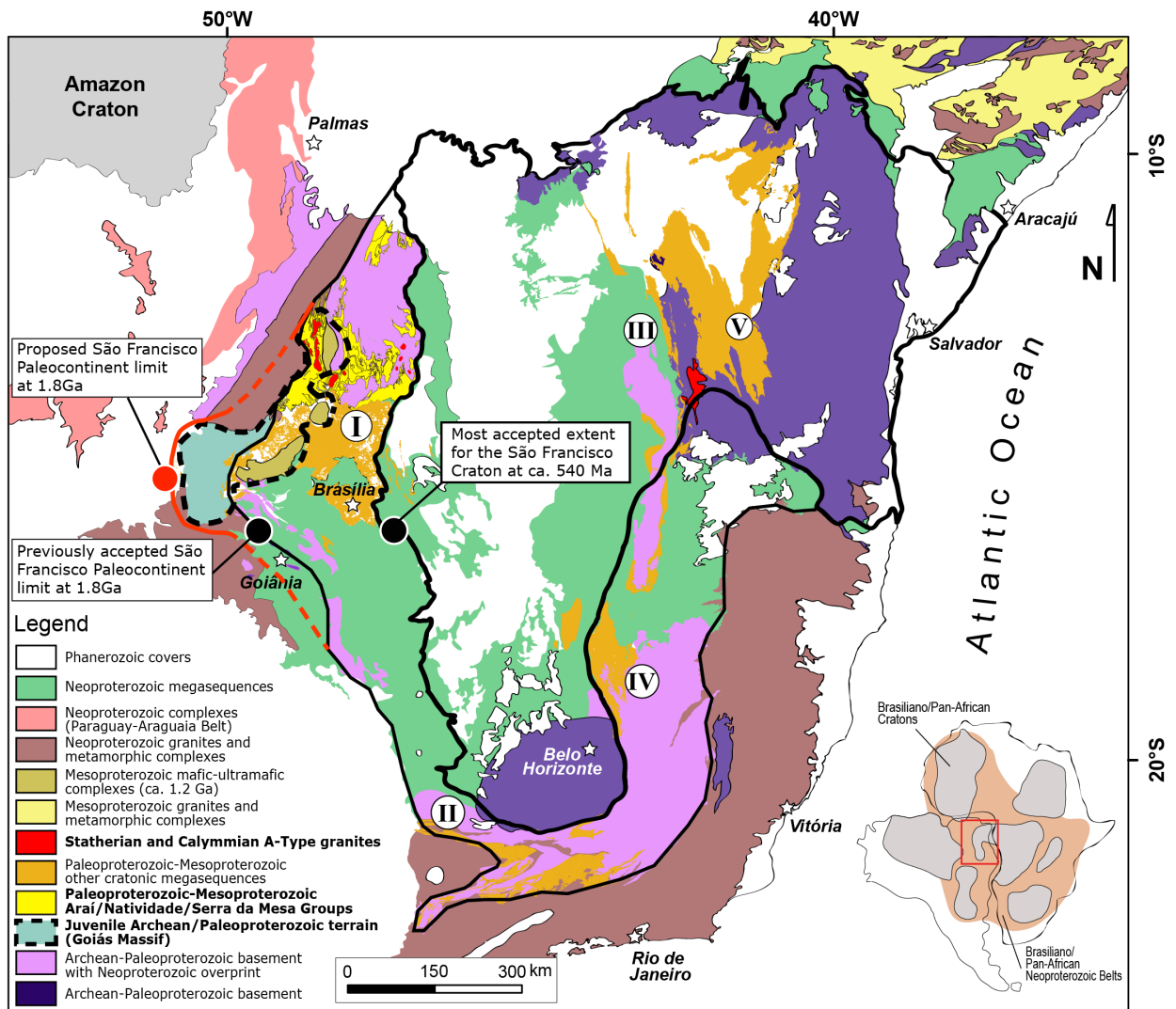


Figure 10 – Geologic map (modified from figure 1) indicating a more accurate São Francisco Palecontinent western limit at 1.8 Ga (red line). The black lines indicate the most accepted extent of the Craton after ca. 540 Ma and the previous paleocontinent limit at ca. 1.8 Ga. Dotted blue line indicates the limits of the Goiás Massif according to Fuck et al. (1994). I) Northern Brasília Belt (Veadeiros Supergroup); II) Northern Espinhaço range; III) Diamantina Plateau; (Chapada Diamantina); IV) Southern Brasília Belt and V) Southern Espinhaço range (modified from Guadagnin and Chemale, 2015).

5. Conclusions

- The maximum depositional age for the Traíras Formation (Araí Group) is 1543 ± 31 Ma.
- The Traíras Formation (upper Araí Group) represents an intracontinental sag-type basin, separated from the rift-related Arraias Formation by a timespan of at least 228 Ma, marked by a major unconformity, followed by conglomerates that rework the Arraias Formation, and as such, the Traíras Formation should be separated from the Araí Group and raised to the group status.

- The Araí Group is restricted to the rift-related sediments and is divided in two formations, Água Morna and Arraias. The Arraias Formation is sub-divided in five members, Prata, Cubículo, Buracão, Mutum and Ventura.
- The Veadeiros Supergroup is proposed to encompass all the Proterozoic extensional first-order sequences in the São Francisco Craton western margin, including the Araí, Traíras and Paranoá groups and their lateral correlatives Serra da Mesa, Natividade and Canastra groups.
- The Traíras and Serra da Mesa groups might have coexisted in time if were not part of the the same sag basin system, respected their lateral heterogeneities.
- The and Serra da Mesa Granite has intruded the Rhyacian basement and sedimentary covers during the first deposition stages of the Traíras and Serra da Mesa groups.
- The Goiás Massif, interpreted as the basement for the Serra da Mesa Group and possible source of Rhyacian positive $\epsilon_{\text{Hf}}(t)$ values. According to this interpretation, it unlikely represents an exotic block accreted to the SF craton during the Neoproterozoic, but would be more likely to have been accreted to the São Francisco craton earlier, possibly during the Rhyacian orogeny.
- The São Francisco continental paleocontinent limit at ca. 1.8 Ga is proposed to coincide approximately with the Goiás Massif western limit. That new limit is proposed.

Acknowledgements

We would like to acknowledge the Geosciences Institute, University of Brasília for providing fieldwork financial and logistic support. The authors are grateful to the Brazilian foundation FAPESP (São Paulo Research Foundation), process number: 2014/13792-5. We also thank Cristiano de Carvalho Lana and Ana Ramalho Alkmim for their help with U-Pb and Lu-Hf analysis.

References

- Alkmim, F.F. and Marshak, S. 1998. Transamazonian Orogeny in the Southern São Francisco Craton Region, Brazil: Evidence for Paleoproterozoic collision and collapse in the Quadrilátero Ferrífero. *Precambrian Research*, 90, 29–58.
- Alkmim, F.F., Marshak, S. Fonseca, M.A., 2001. Assembling West Gondwana in the Neoproterozoic: Clues from the São Francisco craton region, Brazil. *Geology*, v. 29, p. 319–322.
- Alkmim, F.F. and Martins-Neto, M.A., 2012. Proterozoic first-order sedimentary sequences of the São Francisco craton, eastern Brazil. *Marine and Petroleum Geology*, 33, 127–139.

- Almeida, F.F.M., 1977. O Cráton do São Francisco. *Revista Brasileira de Geociências*, 7, 285–295.
- Almeida, F.F.M. 1981. O Cráton do Paramirim e suas relações com o do São Francisco. Simpósio Sobre o Cráton do São Francisco e Suas Faixas Marginais. Salvador, 1981. SBG-Núcleo Bahia, Salvador, Anais, p. 1–10.
- Alvarenga, C.J.S., Dardenne, M.A., Vieira, L.C., Martinho, C.T., Guimarães, E.M., Santos, R.V., Santana, R.O., 2012. Estratigrafia da borda ocidental da Bacia do São Francisco. *Boletim de Geociências Petrobrás* 20, 145-164.
- Andersen, T. 2014. The detrital zircon record: supercontinents, parallel evolution or coincidence? *Precambrian Research* 244, 279–287.
- Beghelli Jr., L.P. Charnockitos e ortognaisses da porção centro-oeste do bloco arqueano de Goiás: dados geoquímicos e isotópicos. MSc Thesis, Universidade de Brasília, 77p., 2012.
- Black, L.P., Kamo, S.L., Allen, C.M., Aleinikoff, J.N., Davis, D.W., Korsch, R.J., Foudoulis, C., 2003. TEMORA 1: A new zircon standard for Phanerozoic U-Pb geochronology. *Chem. Geol.* 200, 155–170. doi:10.1016/S0009-2541(03)00165-7.
- Blum, M.L.B., Moraes, R.A.V., Pires, A.C.B., Jost, H., 2003. Caracterização dos complexos ortognáissicos arqueanos de Goiás por gamaespectrometria aérea. *Revista Brasileira de Geociências* 33, 147-152.
- Borges, C.C.A.; Toledo, C.L.B.; Silva, A.M.; Chemale, F.; Jost, H.; Lana, C.C. 2017. Geochemistry and isotopic signatures of metavolcanic and metaplutonic rocks of the Faina and Serra de Santa Rita greenstone belts, Central Brazil: Evidences for a Mesoarchean intraoceanic arc, *Precambrian Research*, Volume 292, Pages 350-377, <https://doi.org/10.1016/j.precamres.2017.02.017>.
- Botelho N.F., Dantas E.L., Laux J.H., Junges S.L. 2006. The Pleoproterozoic peraluminous Aurumina granite suite, Goiás and Tocantins: whole-rock geochemistry and Sm-Nd isotopic constraints. In: *The Paleoproterozoic Record of the São Francisco Craton, Brazil*. IGCP 509, Brazil, pp. 9-21.
- Brito Neves, B.B., Cordani, U.G., 1991. Tectonic evolution of South America during the Late Proterozoic. *Prec Res* 53, 23-40.
- Campos, J.E.G., Dardenne, M.A., Freitas-Silva, F.H., Martins-Ferreira, M.A.C., 2013. Geologia do Grupo Paranoá na porção externa da Faixa Brasília. *Brazilian Journal of Geology* 43, 461-475.
- Cordeiro, P.F and Oliveira, C.G. 2017. The Goiás Massif: Implications for a pre- Columbia 2.2 to 2.0 Ga continent-wide amalgamation cycle in central Brazil, *Precambrian Research* (2017), doi: <http://dx.doi.org/10.1016/j.precamres.2017.06.021>.
- Costa L. A. M., Portella A. C. P., Nilson A. A., Vale C. R. O., Marchetto C. L. M., Santos E. L., Meneguesso G., Inda H. A. V., Sterna L. H., Marchetto M., Baptista M. B., Fratin O., Mosmann O., Oliveira T. F., Silva W. G. 1976. Projeto Leste Tocantins/Oeste do São Francisco. Relatório Final, Rio de Janeiro, DNPM/CPRM/PROSPEC.

- CPRM. 2008. Org. por Maria Luiza Osório Moreira, Luiz Carlos Moreton, Vanderlei Antônio de Araújo, Joffre Valmório de Lacerda Filho e Heitor Faria da Costa. Programa Geologia do Brasil: Integração, Atualização e Difusão de Dados de Geologia do Brasil. GEOLOGIA do Estado de Goiás e Distrito Federal. Escala 1:500.000. Goiânia: CPRM/SIC - FUNMINERAL, 143 p. il.; + mapa.
- Chakraborty, C. and Bhattacharyya, A., 1996. Fan-delta sedimentation in a foreland moat, Deoland formation, Vindhyan Supergroup, Son valley. In: Bhattacharyya, A. (Ed.), Recent Advances in Vindhyan Geology. Mem. Geol. Soc. India, v. 36, p. 27-48.
- Chemale, F., Alkmim, F.F., Endo, I., 1993. Late Proterozoic tectonism in the interior of the São Francisco craton. In: Findlay, R.H., Unrug, R., Banks, M.R., Veevers, J.J. (Eds.), Gondwana Eight: Assembly, Evolution and Dispersal. Balkema, Rotterdam, pp. 29-41.
- Chemale, F., Dussin, I.A., Alkmim, F.F., Martins, M.S., Queiroga, G., Armstrong, R., Santos, M.N., 2012. Unravelling a Proterozoic basin history through detrital zircon geochronology: the case of the Espinhaço Supergroup, Minas Gerais, Brazil. Gondwana Res. 22, 200-206.
- Chu, N.C., Taylor, R.N., Chavagnac, V., Nesbitt, R.W., Boella, M., Milton, J.A., 2002. Hf isotope ratio analysis using multi-collector inductively coupled plasma mass spectrometry: an evaluation of isobaric interference corrections. J. Anal. At. Spectrom. 17, 1567-1574.
- Cuadros, F.A. 2017. Relações petrogenéticas entre a suíte granítica aurumina e sua encaixante, a formação ticunzal, no setor setentrional da zona externa da faixa Brasília. PhD Thesis. Instituto de Geociências, Universidade de Brasília, Brazil.
- Cuadros, F.A.; Botelho, N.F.; Fuck, R.A.; Dantas, E.L. 2017. The peraluminous Aurumina Granite Suite in central Brazil: An example of mantle-continental crust interaction in a Paleoproterozoic cordilleran hinterland setting?. Precambrian Research, 299, 75-100.
- Dardenne, M.A. 2000. The Brasília Fold Belt. In: U.G. Cordani, E.J. Milani, A. Thomaz Filho & D.A. Campos. 2000. Tectonic evolution of South America. 31st International Geological Congress, Rio de Janeiro, SBG. p. 231-263.
- Dardenne, M. A. Moura, M. A. Dantas, Elton Luiz, 2003. O Granito Serra Dourada (Goiás): um biotita granito intrusivo no Grupo Serra da Mesa. In: VIII Simpósio de Geologia do Centro-Oeste, Cuiabá - MT. Boletim de Resumos. Cuiabá: SBG - Núcleo Centro-Oeste, 2003. v. único. p. 83-84.
- Delgado I.M, Souza.J.D, Silva.L.C, Silveira Filho.N.C, Santos R.A, Pedreira A.J, Guimarães J.T, Angelim L.A, Vasconcelos A.M Gomes I.P, Lacerda Filho J.V, Valente C.R Perrotta M.M, Heinick C.A; 2003 Província Tocantins. In: Bizzi, L.A.; Schobbenhaus, C.; Vidotti, R.M.; Gonçalves,J.H. (Ed.). Geologia, Tectônica e Recursos Minerais do Brasil. Rio de Janeiro: CPRM, 2003. p.281-292.
- Dickinson, W.R., Gehrels, G.E., 2009. Uses of U-Pb ages of detrital zircons to infer maximum depositional ages of strata: a test against a Colorado Plateau Mesozoic database. Earth Planetary Science Letters 288, 115-125.

- Farina, F.; Albert, C.; Martínez Dopico, C.; Aguilar Gil, C.; Moreira, H.; Hippertt, J.P.; Cutts, K.; Alkmim, F.F.; Lana, C. (2016) The Archean-Paleoproterozoic evolution of the Quadrilátero Ferrífero (Brasil): Current models and open questions. *J S Am Earth Sciences* 68:4-21.
- Ferreira Filho, C.F., Pimentel, M.M., Araújo, S.M., Laux, J.H., 2010. Layered intrusions and volcanic sequences in Central Brazil: geological and geochronological constraints for Mesoproterozoic (1.25 Ga) and Neoproterozoic (0.79 Ga) igneous associations. *Precambrian Res.* 183, 617–634.
- Fuck R.A.; Pimentel, M.M; D'el-Rey Silva L.J. 1994. Compartimentação Tectônica na porção oriental da Província Tocantins. In: Congresso Brasileiro de Geologia, 38, Camboriú (SC), Anais, p.215-216.
- Fuck, R.A., Dantas, E.L., Pimentel, M.M., Botelho, N.F., Armstrong, R., Laux, J.H., Junges, S.L., Soares, J.E., Praxedes, I.F. 2014. Paleoproterozoic crust-formation and reworking events in the Tocantins Province, central Brazil: A contribution for Atlantica supercontinent reconstruction. *Precambrian Research*, 244: 53–74.
- Fuck, R.A.; Pimentel, M.M.; Alvarenga C.J.S.; Dantas, E.L. 2017. The Northern Brasília Belt. In: São Francisco Craton, Eastern Brazil. *Tectonic Genealogy of a Miniature Continent*. Pgs 205-220. Editors: Heilbron, Monica, Cordani, Umberto G., Flecha de Alkmim, Fernando. *Regional Geology Reviews*. Springer, 1ed. DOI:10.1007/978-3-319-01715-0.
- Gerdes, A., Zeh, A., 2006. Combined U–Pb and Hf isotope LA-(MC-)ICP-MS analyses of detrital zircons: Comparison with SHRIMP and new constraints for the provenance and age of an Armorican metasediment in Central Germany. *Earth Planet. Sci. Lett.* 249, 47–61. doi:10.1016/j.epsl.2006.06.039.
- Giustina, M.E.S.D., Oliveira, C.G., Pimentel, M.M., Melo, L.V., Fuck, R.A., Dantas, E.L., Buhn, B., 2009a. U–Pb and Sm–Nd constraints on the nature of the Campinorte Sequence and related Paleoproterozoic juvenile orthogneisses, Tocantins Province, Central Brazil. *Geol. Soc. Lond. Spec. Publ.* 323, 255–269.
- Guadagnin, F.; Chemale, F. 2015. Detrital zircon record of the Paleoproterozoic to Mesoproterozoic cratonic basins in the São Francisco Craton, *Journal of South American Earth Sciences*, Volume 60, July 2015, Pages 104-116.
- Guadagnin, F., Chemale Jr., F., Magalhães, J., Santana, A., Dussin, I., Takehara, L., 2015. Age constraints on crystal-tuff from the Espinhaço supergroup - insight into the Paleoproterozoic to Mesoproterozoic intracratonic basin cycles of the Congo-São Francisco Craton. *Gondwana Research* 27, 363-376.
- Jackson, S.E.; Pearson, N.J.; Griffina, W.L.; Belousova, E.A. 2004. The application of laser ablation-inductively coupled plasma-mass spectrometry to in situ U–Pb zircon geochronology. *Chemical Geology*, 211: 47-69

- Kitajima, L. F. W., Ruiz, J. Gehrels, G., GASPAR, J. C. Uranium-Lead ages of zircon megacrysts and zircon included in corundum from Peixe Alkaline Complex (Brazil). In: III Simposio Sudamericano de Geologia Isotopica, 2001, Pucón. III Simposio Sudamericano de Geologia Isotopica, 2001.
- Ludwig, K.R., 2003. User's Manual for Isoplot/Ex version 3.00—A Geochronology Toolkit for Microsoft Excel, No. 4. Berkeley Geochronological Center Special Publication, 70 pp.
- Marini, O.J.; Fuck, R.A.; Dardenne, M.A.; Danni, J.C.M. 1984a. Província Tocantins - Setores Central e Sudeste. In: Almeida, F.F.M. & Hasui, Y. (coord.) - 1984 - O Pré- cambriano do Brasil, p. 205-264. Edit. Edgard Blücher Ltda.
- Marini O.J., Fuck R.A., Danni J.C.M., Dardenne M.A., Loguércio S.O., Ramalho R. 1984b. As faixas de dobramento, Uruaçu, Paraguai-Araguaia e o Maciço Mediano de Goiás. In: C. Schobbenhaus, D.A. Campos, G.R. Derze, H.E. Asmus (coords.), Geologia do Brasil, Brasília, DNPM, pp.251-303.
- Marques, G.C. 2009. Geologia dos Grupos Araí e Serra da Mesa e seu embasamento no sul de Tocantins. MSc Thesis. IG-UnB, 116 p.
- Martins-Ferreira, M.A.C. 2017. Evolução tectônica das bacias Araí e Paranoá na faixa Brasília norte. PhD Thesis. Instituto de Geociências. Universidade de Brasília. Brazil.
- Matteini, M., Dantas, E.L., Pimentel, M.M., Alvarenga, C.J.S., Dardenne, M.A. 2012. U-Pb and Hf isotope study on detrital zircons from the Paranoá Group, Brasília Belt Brazil: constraints on depositional age at Mesoproterozoic-Neoproterozoic transition and tectono-magmatic events in the São Francisco Craton. *Precambrian Res.* 206-207, 168-181.
- Patchett, P.J.; Kuovo, O.; Hedge, C.E.; Tatsumoto, M. 1981. Evolution of continental crust and mantle heterogeneity: evidence from Hf isotopes. *Contrib. Mineral. Petrol.* 78, 279– 297.
- Petri, S.; Coimbra, A.M.; Amaral, G.; Ojeda, Y.H.; Fulfaro, V. J.; Ponçano, W. L. 1986a. Código brasileiro de nomenclatura estratigráfica. *Revista Brasileira de Geociências*, São Paulo, v. 16, n. 4, p. 370-376.
- Petri, S.; Coimbra, A. M.; Amaral, G.; Ponçano, W. L. 1986b. Guia de nomenclatura estratigráfica. *Revista Brasileira de Geociências*, São Paulo, v. 16, n. 4, p. 376-415.
- Pimentel M.M., Rodrigues J.B., Dellagiustina M.E.S., Junges S.L., Matteini M., Armstrong R. 2011. The tectonic evolution of the Neoproterozoic Brasília Belt, central Brazil, based on SHRIMP and LA-ICPMS UePb sedimentary provenance data: A review. *Journal of South American Earth-Sciences*, 31: 345-357.
- Pimentel, M.M.; Jost, H.; Fuck, R.A. 2004. O Embasamento da Faixa Brasília e o arco magmático de Goiás. In: Mantesso-Neto et al (eds) *Geologia do Continente Sul-Americano. Evolução da obra de Fernando Flávio Marques de Almeida*. Editora Becca, São Paulo, pp 356–381.
- Pimentel M.M., Dardenne M.A., Fuck R.A., Viana M.G., Junges S.L., Fischel D.P., Seer H.J., Dantas E.L. 2001. Nd isotopes and the provenance of detrital sediments of the Neoproterozoic Brasília Belt, Central Brazil. *Journal of South American Earth Sciences*, 14:571-585.

- Pimentel, M.M., Botelho, N.F., 2001. Sr and Nd isotopic characteristics of 1.77–1.58 Ga rift-related granites and volcanics of the Goiás tin province, Central Brazil. *An. Acad. Bras. Cienc.* 73, 263–276.
- Pimentel, M.M., Fuck, R.A., Jost, H., Ferreira Filho, C.F., Araújo, S.M. 2000. The basement of the Brasília Fold Belt and the Goiás Magmatic Arc. U.G. Cordani, E.J. Milani, A. Thomaz Filho, D.A. Campos (Eds.), *Proceedings of the 31st Int. Geol. Congr.*, Rio de Janeiro, pp. 195-229.
- Pimentel, M.M., Heaman, L., Fuck, R.A., Marini, O.J. 1991. U-Pb zircon geochronology of Precambrian tin-bearing continental-type acid magmatism in central Brazil. *Precambrian Research* 52, 321–335.
- Prosser, S. 1993. Rift-related linked depositional systems and their seismic expression. *Geological Society, London, Special Publications*, 71:35-66.
- Queiroz, C.L., Jost, H., da Silva, L.C., McNaughton, N.J., 2008. U-Pb SHRIMP and Sm-Nd geochronology of granite-gneiss complexes and implications for the evolution of the Central Brazil Archean Terrain. *Journal of South American Earth Sciences* 26, 100-124.
- Rossi, P.; Andrade, G.F.; Cocherie, A. 1992. The 1.58 Ga type granite of Serra da Mesa (GO): an example of NYF fertile granite pegmatite. In: *Congresso Brasileiro de Geologia*, 37, São Paulo, SBG, Bol. Res. Expandidos, p.389-390.
- Santos, M.M., Lana, C., Scholz, R., Buick, I., Schmitz, M.D., Kamo, S.L., Gerdes, A., Cirfu, F., Tapster, S., Lancaster, P., Storey, C.D., Basei, M.A.S., Tohver, E., Alkmim, A., Nalini, H., Krambrock, K., Fantini, C., Wiedenbeck, M. 2017. A New Appraisal of Sri Lankan BB Zircon as a Reference Material for LA-ICP-MS U-Pb Geochronology and Lu-Hf Isotope Tracing. *Geostandards and Geoanalytical Research*, 1-24. 10.1111/ggr.12167
- Schobbenhaus, C. 1996. As tafrogêneses superpostas Espinhaço e Santo Onofre, Estado da Bahia: Revisão e novas propostas. *Revista Brasileira de Geociências*, 26:265-276.
- Silva W.L., Kotschoubey B., Gallarza M.A. 2005. Grupo Natividade e Grupo Araí: correlatos mas diacrônicos. In: SBGq, *Congresso Brasileiro de Geoquímica*, 10, e *Simpósio de Geoquímica dos Países do Mercosul*, 2, Porto de Galinhas, Pernambuco, Brasil, Anais, CD-ROM.
- Sláma J., Košler J., Condon D.J., Crowley J.L., Gerdes A., Hanchar J.M., Horstwood M.S.A., Morris G.A., Nasdala L., Norberg N., Schaltegger U., Schoene B., Tubrett M.N., Whitehouse M.J. 2008. Plešovice zircon – a new natural reference material for U–Pb and Hf isotopic microanalysis. *Chem. Geol.*, 249: 1-35.
- Stacey, J.S. and Kramers, J.D. (1975). Approximation of Terrestrial Lead Isotope Evolution by a 2-Stage Model. *Earth and Planetary Science Letters* 26(2): 207-221.
- Tanizaki, M.L.N.; Campos, J.E.G; Dardenne, M.A. 2015. Estratigrafia do Grupo Araí: registro de rifteamento paleoproterozoico no Brasil Central. *Brazilian Journal of Geology*, 45(1): 95-108.
- Trompette, R.; Uhlein, A.; Egydio Da Silva, M.; Karmann, I. 1992. The Brasiliano São Francisco craton revisited (central Brazil). *Journal of South American Earth Sciences*, 6, 49–57.

- Valeriano, C.M., Pimentel, M.M., Heilbron, M., Almeida, J.C.H., Trouw, R.A.J., 2008. Tectonic evolution of the Brasília Belt, Central Brazil, and early assembly of Gondwana, Geological Society Special Publication, pp. 197-210.
- Woodhead, J.D., Hergt, J.M., 2005. A Preliminary Appraisal of Seven Natural Zircon Reference Materials for In Situ Hf Isotope Determination. *Geostand. Geoanalytical Res.* 29, 183–195. doi:10.1111/j.1751-908X.2005.tb00891.x

CAPÍTULO 4

Artigo 3:

**Paleorift structure constrained by gravity and stratigraphic data, the
Statherian Araí Rift case**

Paleorift structure constrained by gravity and stratigraphic data, the Statherian Araí Rift case

Marco Antonio Caçador Martins-Ferreira¹; José Eloi Guimarães Campos¹; Monica Giannoccaro Von Huelsen^{1,2}; Brandow Lee Neri²

1. Instituto de Geociências, Universidade de Brasília, Campus Universitário Darcy Ribeiro, Asa Norte, CEP 70.910-900, Brasília - DF - Brasil.
2. Observatório Sismológico. Universidade de Brasília. Campus Universitário Darcy Ribeiro, Asa Norte, CEP 70.910-900, Brasília - DF - Brasil.

Abstract

Gravimetric and stratigraphic data were used to investigate the Araí Paleorift, a failed Statherian continental rift located in the western margin of the São Francisco Craton. The major contribution of this paper is the recognition of paleo-rift system morphology, up to now barely known, over ca. 50 km² area and the progress on the investigation of a possible source for the long-linear low-density/velocity anomaly beneath the rift main axis. Euler deconvolution, tilt, total horizontal gradient amplitude and upward continuation technics were applied to terrestrial gravimetric data in order to investigate the rift's main faults location, depth and orientation and thus constraint the rift installation history. We found that rift faults occur to a maximum depth of ca. 38 km, but major fault throw occurs from 4 to 8 km deep and are attenuated from 8 to 12 km interval, probably the brittle-ductile transition zone at the time of rifting and practically disappearing after 20 km. Stratigraphic data was analyzed in order to identify the rift's horsts and grabens, to locate the main volcanic and plutonic sources. We classify the Araí rift as a passive, three armed failed rift, narrow to divergent type, that produced preferably anorogenic rapakivi-related magmas, most of it still lodged in the crust from surface down to ca. 19 km deep and subsidiary mafic magmatism. The deep occurrence of low-density magmas might be explained by the decompression generated by the paleorift faults, still generating significant fault throw up to 20 km in depth. Correlation to other Statherian rifts shows that the São Francisco Craton was strongly affected by Statherian taphrogenesis together with Siberia, North America and North China. Finally, by comparison of our discoveries to recent examples we find that the Ethiopian rift (East Africa system) is a quite similar example, especially the Tanzania Divergence, and the whole East Africa rift system surrounding the Tanzanian craton is also quite similar to the system formed by the Araí and Espinhaço rift systems, surrounding the São Francisco craton, indicating that Paleoproterozoic extensional systems were not very different from Cenozoic systems.

Keywords: Araí Group, paleorift, continental failed rift, Statherian extension, São Francisco craton.

1. Introduction

Paleorifts are commonly inverted and eroded or covered by subsequent sequences, hindering the study on how rifting processes occurred in the past, leaving the recent examples as

the East Africa Rift as our only global models. Nonetheless, some paleorifts are partially preserved and exposed, offering a good study opportunity, especially if geophysics can be used. Paleorifts may hide important systems for oil and gas generation and trapping (Chekunov et al. 1993), rare metals and rare earth elements (Botelho and Moura 1998), including uranium and gold (Guryanov and Peskov 2017). Furthermore, the study of ancient rifts can shed light on how extensional tectonic processes worked in the past.

During early Statherian, the São Francisco Craton margins witnessed an extensional event that opened several continental rift basins. By late Calymmian, new extension occurred, and the rift basins were followed by broader sag-type basins, locally initiated by rifting processes. Again, in the Ectasian-Stenian transition, widespread extension was registered by the formation of intracontinental sag-type basins and rifts that might have evolved to passive margins. These three extensional episodes are well constrained by stratigraphic and geochronologic data (Alkmim et al. 2012, Danderfer Filho et al. 2015, Guadagnin and Chemale 2015, Martins Ferreira et al. 2017). Pedrosa-Soares and Alkmim (2011) consider all these extensional events as rift-related and speculate about a total of six rifting events, if considered the Congo Craton.

In the Neoproterozoic, the previous extensional sequences were affected by continental collision that imposed compressional deformation and metamorphism which represent significant obstacles for Proterozoic tectono-stratigraphic studies. Nonetheless, in the orogenic external zones, Paleoproterozoic rift-related sedimentary record is preserved in low metamorphic grade and the rifted crystalline basement is largely exposed, providing an excellent paleorift study opportunity.

Despite the good exposures, the São Francisco Craton Proterozoic rift structures are still poorly known and their elements have been recognized only in a few available seismic sections and aerogeophysical data (Reis and Alkmim 2015). In the São Francisco Craton western margin, the extent, direction, and location of the main rift-related faults is still unknown and vaguely debated in literature and rift inversion/uplift is also a subject not well explored.

Seeking to fill that void, this study makes use of gravimetric, structural and stratigraphic data to investigate the location and nature of the main faults involved in the installation and evolution of the Statherian rift represented by the Araí Group, its correlation to other regional and global Proterozoic rifting events and similarities with recent rift systems. A major unconformity marks the base of the sequence deposited as a sag-type basin that covers the Araí rift sediments. In this study, that unconformity has been mapped and associated with gravity and

other geologic data to better constraint the location of paleograbens and paleohorsts and the main faults that separate them.

Borehole and seismic data are not available for the studied area. Magnetic and gamaspectrometric surveys are available only at the northern part of the studied area and were not used. One seismic profile from an adjacent area further to SE is used for tectono-stratigraphic regional correlation. In areas where seismic and borehole data are not available, potential field studies can be useful to identify main structural features that affect the earth's crust at different depths. According to Bott and Hinze (2006), gravity data has been used to delineate hidden rifts, to determine the shape of ancient and modern rift troughs, including the nature of faulting, to investigate crustal and upper mantle structure in association with explosion and earthquake seismology, and to ascertain the isostatic state of recent and paleorifts.

Some methods require a priori knowledge of underground geology since they work with pre-defined models, others are more suitable to be used in regions where other types of subsurface data, such as wells and seismic profiles, are not available. The fundamentals of these methods reside in the capacity of gravimetric technics to detect abrupt variations in density of rocks. Those abrupt variations are most commonly attributed to faults, contacts with intrusive bodies or unconformities. Rift faults tend to be deep and generate significant fault throw, leading to planar density contrasts between the faulted blocks.

A variety of techniques can be employed to gravimetric data in order to delineate geological discontinuities that generate density contrast. Horizontal and vertical derivatives of potential field anomaly grids have been widely used, such as wavelet analysis (Fedi and Quata, 1998), analytical signal amplitude (Roest et al. 1992), normalized standard deviation (Cooper and Cowan 2008), tilt angle (Miller and Singh 1994, Salem et al. 2008), Euler deconvolution (Daniela et al. 2003, Salem and Ravat 2003) upward and downward continuations (Cella et al. 2009, Roy 2013). The peak value of the total horizontal derivative can be used to investigate the horizontal extension of the anomaly source bodies (Zhou et al. 2006) or to check how much interference these bodies might be causing.

The Euler deconvolution method is a well-established and widely used interpretative tool for locating three-dimensional anomalies in potential data (Alrefaee 2017, Brandmayr and Vlahovic 2016, Liu et al. 2016, Curto et al. 2015, Khalil et al. 2014, Chen et al. 2014, Dhaoui 2014, Jordan et al. 2013, Cooper and Liu 2011, Gout et al. 2010, Chenrai et al. 2010). Its great simplicity of implementation and use makes it very popular and extremely useful, especially in regional geology studies.

Geological data was compiled and grouped as to distinguish the spatial distribution of crystalline basement lithological heterogeneities and sedimentary sequences deposited during rift and sag phases. As the chief edge detector we use Euler deconvolution assisted geological mapping data from basement and covers. Secondary edge detectors are Total Horizontal Gradient and Analytic Signal Inclination. The edge detectors are used to delineate the Araí paleorift main faults that separate major horsts and grabens as well as to infer rift transfer faults. Upward continuation and Euler Voxel 3D model are used to investigate the source of low density anomaly beneath the Araí Paleorift.

2. Geological Setting

The intracratonic sequences studied are preserved in the Northern Brasília Belt, an orogenic belt formed during the Neoproterozoic Brasiliano collage in the western margin of the São Francisco Craton. The Brasília Belt external zone, where the study is focused (Figure 1), is a fold-and-thrust belt where rocks were locally submitted to low-greenschist facies metamorphism, but are generally anquimetamorphic (Dardenne 2000).

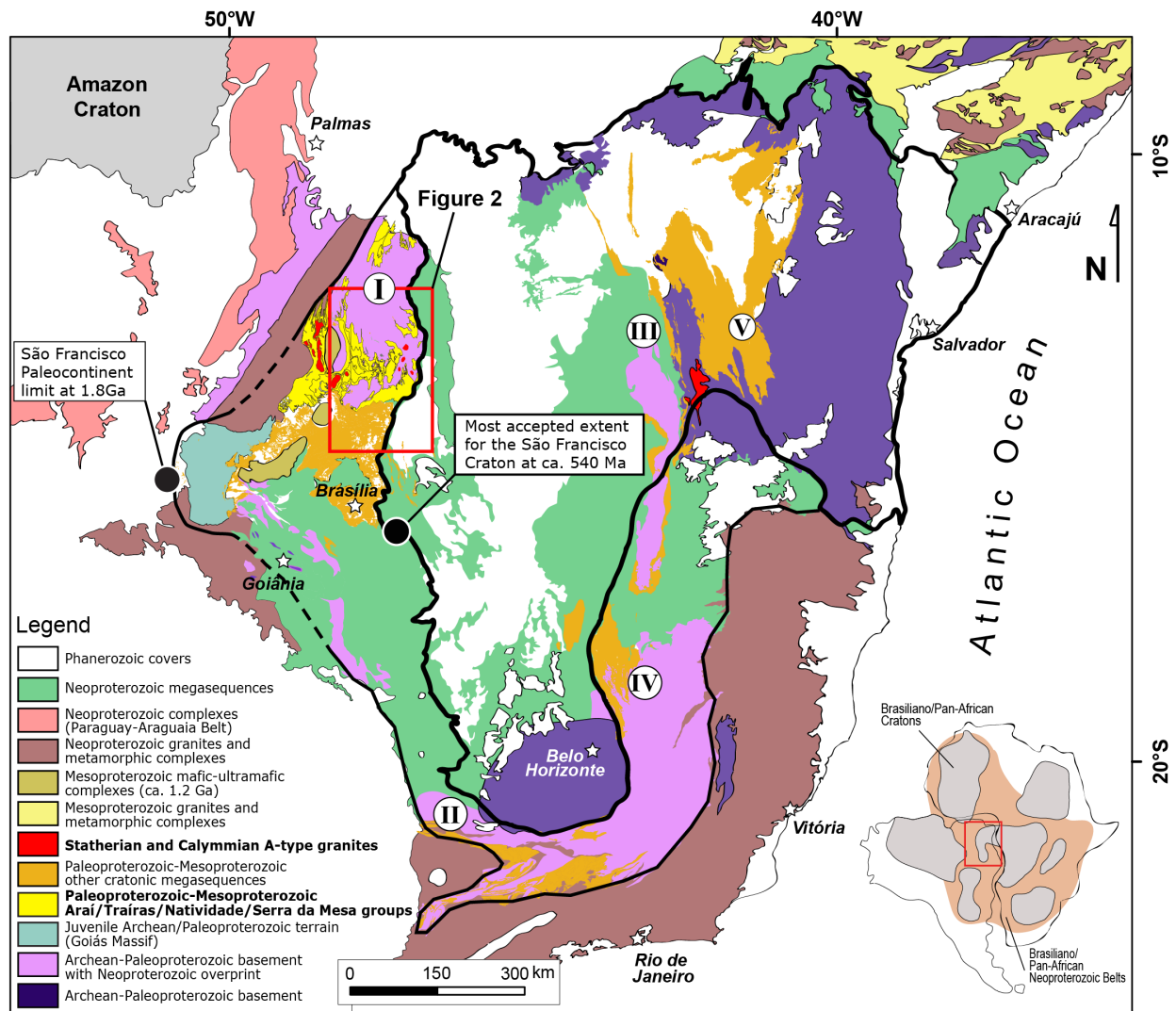


Figure 1 - Geologic map showing the distribution of the Paleoproterozoic rift-sag sequences in the São Francisco Craton interior and margins (orange color) formed by the Espinhaço and Veadeiros Supergroups. The black lines indicate the most accepted extent of the Craton after ca. 540 Ma and the probable extent of the paleocontinent at ca. 1.8 Ga as proposed by Martins Ferreira et al. (2017). I) Northern Brasília Belt; II) Southern Brasília Belt; III) Northern Espinhaço range; IV) Southern Espinhaço range and V) Diamantina Plateau (Chapada Diamantina); (modified from Guadagnin and Chemale, 2015).

Despite not much affected by metamorphism, the preserved sedimentary sequences are folded and faulted in different styles and intensities across the Brasília Belt. In the study area, isoclinal to gentle folds usually have curved axes and are occasionally cut by transpressional shear zones generally parallel to the fold axis.

The Rhyacian crystalline basement is mainly represented by the Aurumina Suite (Cuadros et al. 2017a), well exposed over a large area of hundreds of square km, allowing the mapping of its main structures and lithological heterogeneities (Figure 2). In the regional map, it can be promptly noticed that basement lithological variation seems to be controlled by tectonic contacts. That basement was affected by three main extensional pulses, that can be inferred by

the geological record as Statherian (Araí Group), Calymmian (Traíras and Serra da Mesa groups) and Ectasian-Stenian (Paranoá Group), forming distinct intracontinental basins whose sequences comprise the Veadeiros Supergroup (Martins-Ferreira et al. 2017). The geological aspects of these geologic units are described in more detail below.

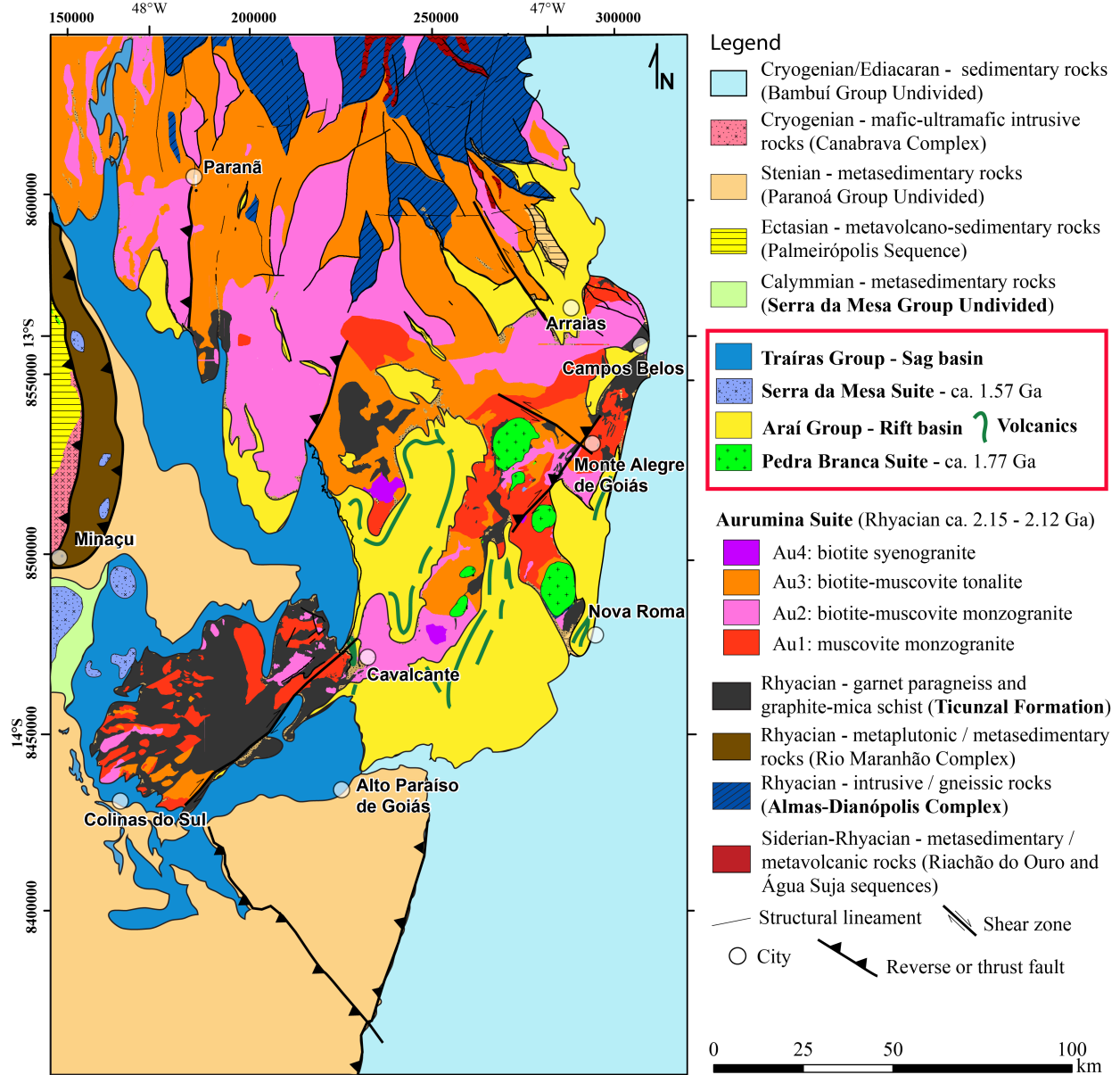


Figure 2 - Study area geological map showing the main cities. (Modified from Cuadros, 2017, that produced this map with geologic data compiled from geological maps SC.23, SD.22, SD.23, SC.22-Z-D, SD.22-X-D, SD.23-V-A, SC.23-Y-C, SD.23-V-C-III, SD.23-V-C-V and SD.23-V-C-VI by the Brazilian Geological Survey (CPRM; <http://geobank.cprm.gov.br/>). Modifications made to the map in this study were focused on the division of previously undivided geologic units, including the anorogenic granites (Pedra Branca and Serra da Mesa suites) and the supracrustal rift and sag sequences, respectively the Araí and Traíras groups (after Martins-Ferreira et al. 2017) using our own geologic mapping data and the data compiled from the final graduation detail geological mapping at University of Brasília in the years 1978, 1984, 1998, 1999, 2001, 2005, 2009 and 2010.

2.1. Rhyacian Crystalline Basement

In the western margin of the São Francisco Craton, the Rhyacian was a period of active margin development and terrain accretion. These orogenic events produced intense acid to intermediate magmatism known as the Aurumina suite. Botelho et al. (1999) originally described the rocks of this suite as medium- to coarse-grained foliated to isotropic granite, granodiorite and tonalite. It intruded and melted great portions of the previous basin sediments of the Ticunzal Formation (Cuadros et al. 2017b), leaving metric to km scale xenoliths and roof pendants composed of schist and paragneiss commonly exhibiting lit-par-lit injection. Initially, peraluminous granite facies were generated, which in turn contrast in composition with late tonalite and granodiorite facies. The rock types were labeled as Au1 to Au6 by Alvarenga et al. (2007) and comprise, from oldest to youngest facies: muscovite monzogranite (Au1), muscovite-biotite monzogranite (Au2), biotite tonalite (Au3), biotite syenogranite (Au4), tourmaline leucogranite and tourmaline pegmatite (Au6). Facies Au5 is rarely recorded and corresponds to migmatitic granites intimately associated with the Ticunzal Formation and will not be addressed. Facies Au6 will also not be represented on maps and charts for its low occurrence and scale reasons.

2.2. Statherian Rift and Calymmian Sag Sequences

The Araí Group was historically described as a Statherian rift-sag or rift-post-rift continuous sequence (Pimentel et al. 1991, Alvarenga 2007). It was latter divided in four tectono-sequences: pre-rift (Água Morna Formation), rift (Arraias Formation), transitional (Caldas Formation) and post-rift (Traíras Formation) (Tanizaki et al. 2015) based on their main lithological and environmental characteristics. The bimodal volcanism associated to rifting was dated ca. 1771 ± 2 Ma (Pimentel et al. 1991).

The sag sequence (Traíras Formation) that covers the characteristic rift sediments (Arraias Formation) was until then believed to be the Araí post-rift sequence. However, recent zircon provenance studies have shown that it in fact represents a much younger, Calymmian sequence, with maximum depositional age calculated ca. 1543 ± 31 Ma (Martins-Ferreira et al. 2017) and separated, thus, from the rift volcanism by a time span of at least ca. 230 Ma. The Traíras Formation was hence raised to group status and its members to formation status (Figure 3).

Both extensional pulses, Statherian Araí and Calymmian Traíras, have generated associated anorogenic magmatism. The Pedra Branca and Serra da Mesa suites, also known as

g1 and g2, respectively. The g1 suite is metaluminous and alkaline, while g2 suite is metaluminous to peraluminous, both enriched in F, Sn, Rb, Y, Th, Nb, Ga and REE (Botelho and Moura 1998) dated ca. 1.77 Ga and ca. 1.57 Ga, respectively (Pimentel et al. 1991). They are considered to be rapakivi granites (Lenharo et al. 2002). It is important to observe, though, that intrusions of the two suites do not overlap each other in spatial distribution, they form, otherwise, two different magmatic provinces, the Pedra Branca granites comprising the Tocantins Province to the east, related to the Araí Rift and the Serra da Mesa granites comprising the Parana Province to the west, associated to the Traíras sag, as can be seen on figure 2.

In the São Francisco Craton eastern margin, the Pedra Branca Suite is correlated to the Borrachudos and Lagoa Real suites of anorogenic granites (Pedrosa-Soares and Alkmim 2011, and references therein).

The Araí Group: Pre-rift and Rift Sequences

The *Araí Group* is divided in the *Água Morna* and *Arraias* formations (Martins-Ferreira et al. 2017), respectively representing the pre-rift and rift sequences. In this work, the term *pre-rift sequence* is understood as representing the sedimentary succession deposited previously to the beginning of rift faulting but already under the influence of crustal doming and/or subsidence associated the early stages rifting processes.

The *Água Morna Formation* is comprised mainly of braided river fluvial deposits interpreted as deposited in an intracontinental shallow basin generated by thermo-flexural subsidence previous to rift faulting. The occurrence or absence of this formation in the sedimentary record may reveal the dynamics of pre-rift as well as rift stages, once its deposition avoids the domed areas and its preservation is favored in the grabens. However, this sequence is thin and occurs at the base of the hills, being generally capped by colluvium and thereby difficult to be mapped.

The *Arraias Formation* is constituted exclusively by continental sedimentation of alluvial fans, lacustrine, fluvial and aeolian environments with great lateral variation of thickness and facies. Paraconglomerate interlayered with quartzite is the most abundant lithotype. Bimodal volcanism is evident and was dated ca. 1771 ± 2 Ma (Pimentel et al. 1991). Basaltic and rhyolitic lava flows are identified as well as pyroclastic volcanism. The aspects described above characterize the Arraias Formation as a classic representative of a rift type basin record. It has been subdivided in four members strongly interdigitated: Cubículo Member (aluvial fans); Prata Member (braided river); Mutum Member (aeolian); Ventura Member (lacustrine) and Buracão

Member (rhyodacite, rhyolite, ignimbrite, pyroclastic).

The Araí Paleorift volcanism is not significantly voluminous and is characterized by a bimodal association, where the felsic lavas are geochemically similar to the anorogenic A-type rapakivi granites from the Pedra Branca Suite, both showing intraplate magmatism signature and the mafic lavas present continental tholeiitic signature (Alvarenga et al. 2007).

The thickness of conglomerate deposits, layer tilt, the location of volcanic plugs and dikes are important features that aid to constraint the location of main rift structures. The Araí Group post-rift sequence is so far unknown; it might not have been deposited or eroded previously to the deposition of the Traíras Group.

The Traíras Group: Sag Sequence

The *Traíras Group* comprises a marine sequence formed essentially by siliciclastic and mixed platform sediments, frequently influenced by tides and represents a Mesoproterozoic sag basin record. It is subdivided in four formations, as follows: *Caldas Formation* (beach deposits representing transitional environment to the marine sequences that follow), *Boqueirão Formation* (mixed platform dominated by tides), *Rio Preto Formation* (siliciclastic platform dominated by tides) and *Rosário Formation* (mixed platform dominated by tides), and is interpreted as developed in a regime of flexural subsidence (Martins-Ferreira et al. 2017).

The Traíras Group is deposited over the Araí Group or directly over the crystalline basement separated by a major unconformity marked by coarse clast-supported conglomerates that rework the Araí Group strata as well as crystalline basement rocks. The clasts are mainly composed quartzite, resultant from the reworking of fluvial and aeolian quartzite from the Arraias Formation but granite and tonalite clasts are found as well. These conglomerates are positioned at the base of the Caldas Formation. The occurrence or absence of the Caldas Formation in the sedimentary record might also reveal important details on the Traíras basin installation.

The Natividade and Serra da Mesa Groups are considered correlatives of the Traíras Group, possibly the same wide sag-type basin system, respected the metamorphic differences and the lateral heterogeneities inherited from sediment source and paleo-relief variations (Martins-Ferreira et al. 2017).

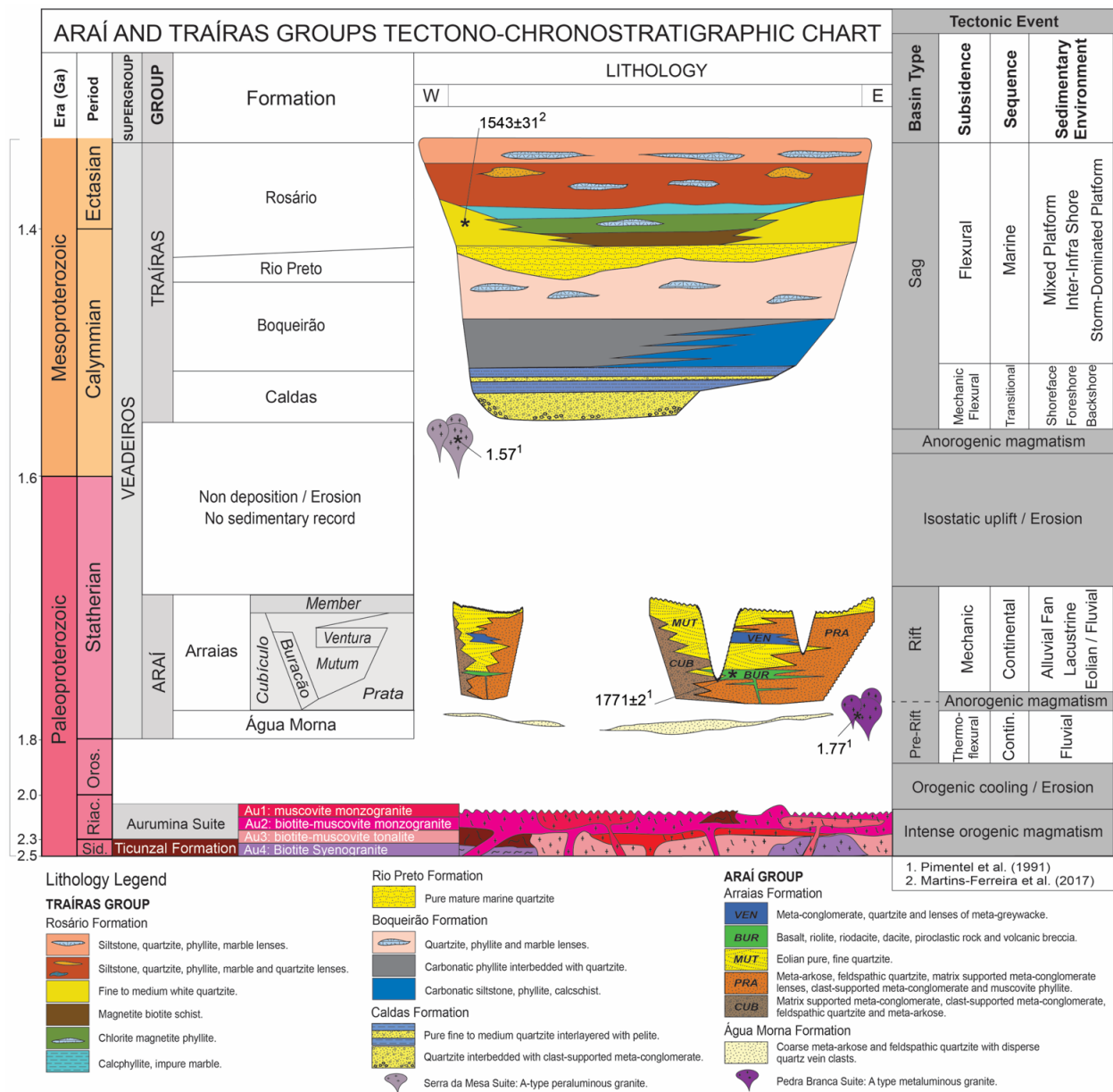


Figure 3 - Araí and Traíras groups tectono-stratigraphic chart with available geochronological data and interpreted basin tectonic events and settings (modified from Tanizaki et al. 2015).

2.3. Stenian to Tonian Intracontinental Sequences

The Paranoá Group covers the Araí Group, and is separated from it by an unconformity followed by the São Miguel Formation conglomerates. The group represents a shallow water marine succession composed mainly of sandstone and siltstone with subordinated limestone (Campos et al. 2013). Available geochronological data for the Paranoá Group consists in detrital zircon U-Pb ages determining a maximum depositional age of 1560 ± 10 Ma and diagenetic xenotime overgrowths U-Pb age suggesting end of deposition at 1042 ± 22 Ma (Matteini et al. 2012). The Canastra Group is considered its lateral correlative, being interpreted as deposited in a contemporaneous sub-basin (Dardenne 2000).

2.4. Cryogenian to Ediacaran Intracontinental Sequences

The Bambuí Group (Dardenne 2000) covers unconformably the Aurumina basement, Araí and Paranoá Groups rocks, indicating uplift and erosion prior to its deposition. In the studied area, it is comprised of basal glaciogenic distal deposits (Martins-Ferreira et al. 2013), followed mainly by pelitic and carbonatic rocks deposited in a marine intracontinental sequence that occurs in the Brasília Belt and continues throughout vast areas of the São Francisco Craton and its marginal orogenic belts. According to Dardenne (2000), the Bambuí sedimentary rocks comprise three regressive megacycles. Each of these megacycles begins with a fast marine transgression of regional amplitude, associated with a sudden subsidence of the basin evidenced by deep pelitic marine facies, passing to shallow-platform facies and tidal to supratidal facies. The subsidence is interpreted to be result from regional compressive tectonics associated to syn- to post Brasiliano orogeny. For that reason, it is not included in the Veadeiros Supergroup.

The accurate age of the Bambuí Group is still uncertain. Cap dolostones that overlie the Jequitaiá Formation diamictites have been dated at ca 740 Ma (Pb-Pb; Babinski et al. 2007), C and Sr isotopic data from the basal carbonates indicate Sturtian age for the Jequitaiá glaciation. Provenance studies, from the Sete Lagoas, Serra de Santa Helena, and Serra da Saudade formations encountered a great number of zircons dated ca. 600 Ma and a few ca. 550 Ma old (Rodrigues 2008; Pimentel et al. 2011), suggesting late Neoproterozoic to early Cambrian depositional age for this sequence (Pimentel 2016).

3. Previous Geophysical Studies

A seismic refraction cross-section in the study area indicates that the crust-mantle boundary is elevated under the region of Cavalcante and Nova Roma towns (Soares et al. 2006). Furthermore, the study admits the existence of a low-density body between ca. 30 and 35 km deep interpreted as remnant rocks resulting from partial melt of Aurumina Suite peraluminous tonalites (Figure 4). The mantle elevation as well as the low density body interpreted by the authors lies right under the Araí rift main locus of continental conglomerates and bimodal magmatism occurrence.

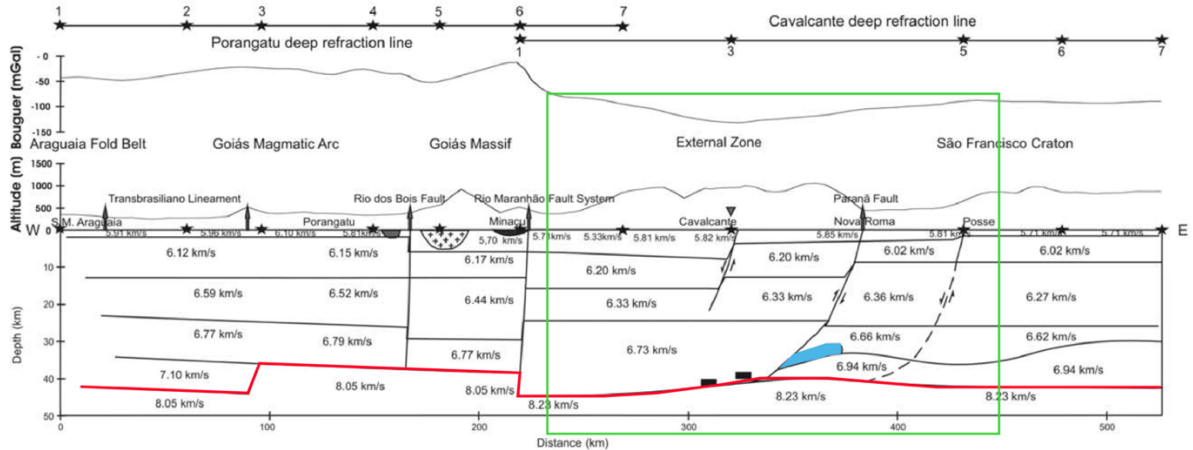


Figure 4 - 2D crust and mantle density model under the Porangatu and Cavalcante deep seismic refraction lines (Soares et al. 2006). The model was obtained by ray-tracing and receiver function modeling from the Cavalcante seismic station. This study area is outlined in green, the crust-mantle interface is outlined in red and the low-density body interpreted by the authors is marked in blue.

Trindade et al. (2014), giving continuity to the work of Soares et al. (2006) conclude that, according to their deep seismic reflection data, the gravimetric variation observed is not a function of crustal thickness, suggesting that regional isostatic equilibrium is actually controlled by mainly by variations in lithospheric physical properties.

4. Gravimetric Data Acquisition, Preparation and Processing

The gravimetric dataset used was composed by the sum of our own survey data and the data acquired by the Instituto Brasileiro de Geografia e Estatística (IBGE). Gravimetric field data from IBGE were acquired at gravimetric stations along roads and thus do not have regular spacing. Nonetheless, data spacing is sufficient for this study scale and objectives (Figure 5). Our specific survey data was acquired along four profiles (Figure 5, in grey) with 500 meters spacing between stations, aiming to fill a blank area in the IBGE dataset in a region where important features might occur.

Data preparation procedure involved the following corrections: instrumental drift, latitude correction, Bouguer correction, free-air correction and terrain correction. Preliminary processing included gridding the data by Kriging interpolation with 2500 m cell size in order to generate a smooth surface with values as close as possible to the original data (Keckler 1994; Geosoft 1995, 2001). A first approach was to grid without blanking distance in order to analyze the areas with absent data (Figure 5a) and later perform blanking distance with 150 points distance in order to fill data gaps (Figure 5b) and compare to the unblanked results. Very high frequencies were later removed by hanning filter.

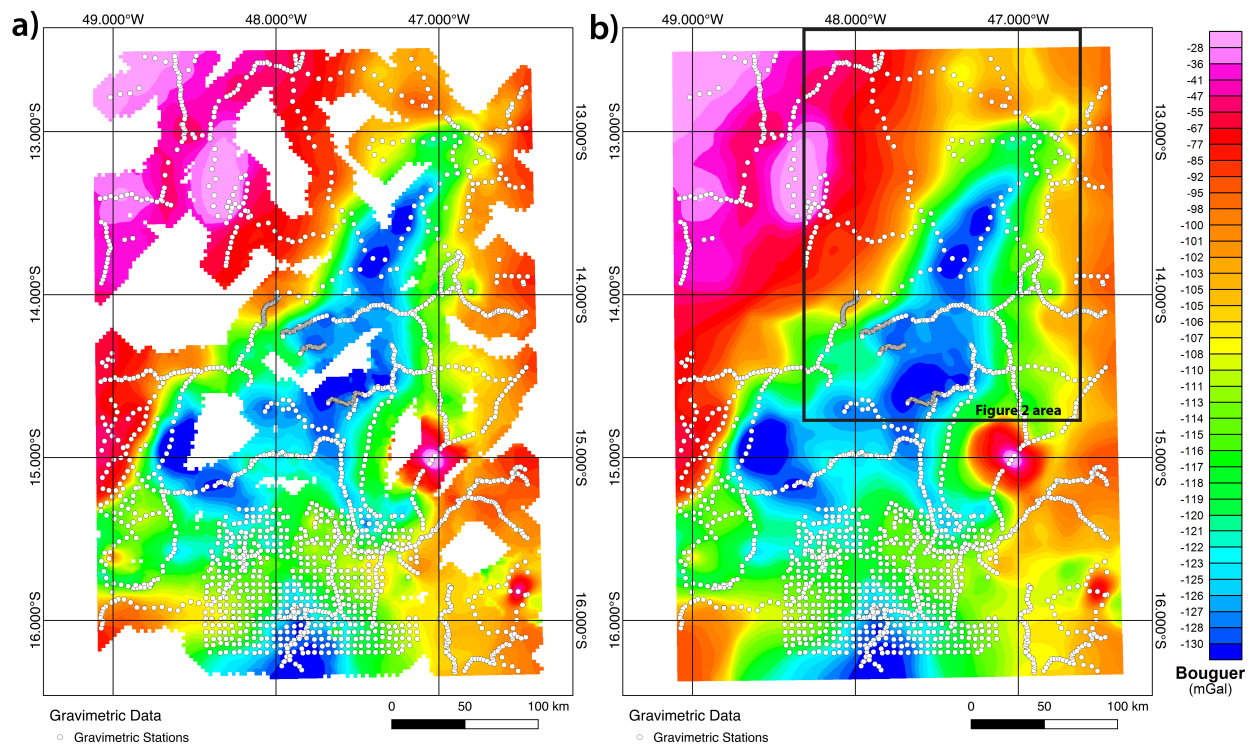


Figure 5 – Bouguer anomaly maps produced with the IBGE dataset with gravity stations superposed. Data was gridded a) without blanking distance and b) blanking distance was used with 150 m cell spacing. The stations surveyed by the authors are outlined in grey.

From the Bouguer gravity field map (BGF), filtering was applied for the generation of the following products: vertical (DZ) and horizontal (DX, DY) derivatives Bouguer, second vertical (2DZ) and horizontal (2DX, 2DY) derivatives Bouguer, analytic signal inclination or tilt angle (TA), total horizontal gradient of gravity data (THDR), analytic signal phase (ASP), as shown in the processing fluxogram (Figure 6). In addition, upward and downward continuation, Power Spectrum, Euler Solutions and 3D Voxel products were obtained. The intent on generating these products was to separate gravimetric wavelengths and thus be able to better isolate and enhance the anomalies related to the study objectives (e.g. Blakely 1996).

The vertical derivative products (DZ) were produced in order to estimate the rate of vertical change of the gravity field, aiming to detect compositional variations in the basement. According to Keller (2002), care should be taken when applying vertical derivatives to anything but very clean data sets because of the potential for amplifying noise. The total horizontal gradient maps high horizontal gradients, and was intended to be used to aid the delineation of faults and or abrupt geologic discontinuities, usually indicated by high continuous anomalies.

The upward continuation consists in a low pass filter process that simulates a higher elevation of data acquisition, allowing to capture the long wavelength anomalies. The process allows to extract the long wavelength anomalies captured from a higher altitude from the total

Bouguer anomaly map, resulting in short wavelength anomalies, called residual maps. Alternatively, the downward continuation works in the opposite way, it represents a high pass filter by which a lower elevation of data acquisition is simulated, enhancing anomalies that represent relatively shallow sources. Both upward and downward continuation were used to constraint the negative anomalies source properties as depth, spatial distribution and possible composition.

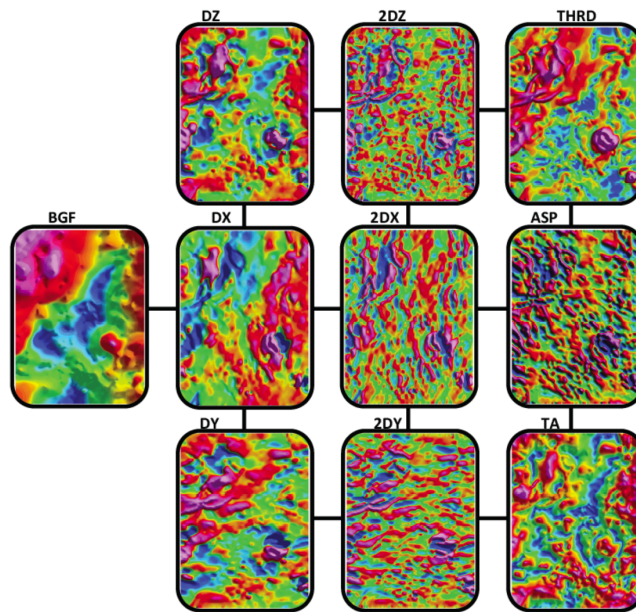


Figure 6 - Data processing fluxogram. THDR = Total horizontal gradient; ASI or TA = Analytic signal inclination (Tilt); BGF = Bouguer gravity fieldgrid.

The Bouguer anomaly and Vertical Derivative images were submitted to the Power Spectrum method. The method was developed by Spector and Grant (1970) for data spectrum analysis. With the use of the Fourier transform it was able to provide statistical estimates of source depths (Figure 7).

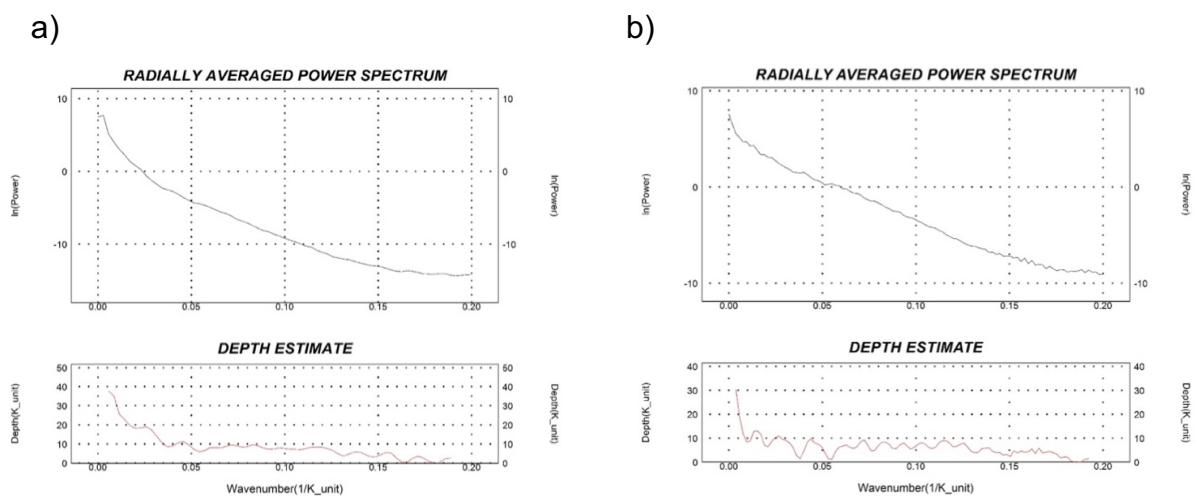


Figure 7 - Radially averaged power spectrum diagrams for a) Bouguer and b) vertical derivative.

Further data processing included the implementation of Euler Deconvolution method over the Bouguer field. The Euler method was developed by Thompson (1982) for profile data analysis and later extended for grid data by Reid et al. (1990). Since then, it has been adapted and improved by Keating (1998). Fairhead et al. (1994) have also improved the technique by including automated processes for eliminating statistically unviable solutions. Barbosa et al. (1999) has contributed significantly to the estimation of the structural indexes in which the basic form of the investigated bodies is restricted and Mushayandebvu et al. (2003) has implemented the ability to identify 2D bodies in a three-dimensional field data. Therefore, Euler deconvolution is both a boundary finder and depth estimation.

It should be emphasized though, that Euler's solutions do not represent the exact three-dimensional contour of the sources. They indicate at most an approximate place of origin. It is also important to remember that Euler solutions do not provide unique results and can be biased as any other reverse geophysical problem. Besides, Chen et al. (2014) estate that due to interference effects caused by nearby sources, especially with some deep-seated fault structures, the edges estimated by vertical and horizontal derivatives are somewhat shifted from the true position even for vertical-sided sources.

Taking into account the interference and bias problems, we have conducted Euler Deconvolution by testing different window and tolerance values, always using the 0 index, indicated for planar structures, since we are interested on mapping faults and contacts. The right parameters were chosen over the test results by several criteria presented in the discussion. Euler Solutions were plotted using a color scale representing gravity contrast source depth.

Later, the most representative index was chosen for a tridimensional representation by a Voxel 3D model. A voxel, short term for volume element, is an ordered pair whose first component is a location of a volume rectangular parallepiped and whose second component is the position and orientation of an edge running through the pixel and whose third component is the strength of the edge (Haralick and Shapiro 1991). It represents a widely used technic for three dimensional picture model construction from 2d sources that is frequently used with Euler data (Toriwaki and Yonekura 2002) but also applied to other types of data for different scientific applications.

5. Results and Discussion

5.1. Gravimetric Support to Rift-Related Research

Gravity modeling has been effectively used to assess crustal and upper mantle structure beneath present-day continental rift systems. The short wavelength effect of shallow structures such as sediment filled troughs can be removed, separated or incorporated to the model, according to the research goals. The Bouguer anomaly can be interpreted in terms of variation of Moho depth, density anomalies within the crust above the Moho and lateral variations in the density of the upper mantle. Characteristically, the Moho beneath a passive rift system shallows as a result of crustal stretching, there may be high and/or low density intrusions within the crust and a broad domal basin may form in preliminary stages of passive rifting. Alternatively, thermal uplift is usually associated with active continental rifts, which is commonly supported by a hot, low density region in the upper mantle usually associated with a plume head (Keen, 1985).

However, geophysical results on paleorift studies will differ from recent rifts because the heat associated to the ancient rifting process might be totally gone, so the possible source for density anomalies under paleorifts are reduced to crustal thickness, mantle composition and intrusions within the crust. Gravity studies have an important role in identifying and studying the crustal structure of ancient rifts. Bott and Hinze (2006) estate that the best approach to investigate ancient rifts is to use explosion seismology to define crustal thickness and its variation and to use the gravity anomaly to refine the crustal model and determine crust-mantle anomalies.

Several cases of low-velocity/low-density zones under recent and paleo-rifts have been reported. Maguire et al. (1994) report the presence of a wide, anomalously low-velocity (7.6-7.8 km/s-1) low-density body at the base of the crust identified along the axial line of the Kenya Rift and which it has been argued must include partial melt. Roy et al. (2005) also report a region of low upper mantle density beneath the Rio Grande rift and suggest it might reflect some combination of temperature variations, partial melt contents, and compositional variations related to Tertiary magmatic activity and/or thinned lithosphere. The relationship of this zone of modified upper mantle to extension in the Rio Grande rift remains enigmatic, however, particularly given the lack of spatial correlation between the trends of upper mantle structures and surficial extensional structures.

On the other hand, Yirgu et al. (2006) report high-density intrusive bodies within the Afar rift that are interpreted as mafic ($\leq 40\%$ gabbro) intrusions originating from magma generated by extension decompression melting of the upper mantle. Rifts that present positive gravity anomalies are those where mantle derived mafic intrusions were highly voluminous, as in the midcontinent rift, USA (1100 Ma old), which is not the case for the Araí rift.

Yu et al. (2015) report results from receiver function stacking and gravity modeling indicating that the crust/mantle boundary beneath the young Okavango Rift, Botswana is uplifted by 4 to 5 km, and the initiation of the rift is closely related to lithospheric stretching. Possible decompression melting of the sub-crustal lithosphere occurs beneath the rift, as evidenced by a relatively low upper mantle density based on the gravity modeling. There is also evidence in literature that low gravity anomalies beneath rifts are not only caused by rift-related modifications in the upper mantle, but also by regional pre-rift crustal sources associated with ancient sutures (Nyblade and Pollack 1992).

Keller et al. (1983) establish that ancient rifts are commonly characterized by a present day thickened crust probably as result of underplating during the rifting process, which results in a broad negative anomaly. This is not the case of the Araí Rift, where the negative anomaly is restricted under the main axis, forming a long and narrow anomaly, requiring a different explanation.

Seismic refraction profiles obtained by the Kenya Rift International Seismic Project show results for the upper mantle that present low Pn velocities of 7.5–7.8 km s⁻¹ under the rift axis, compared with 8.1 - 8.3 km s⁻¹ under the unrifted Mozambique Belt and Tanzania Craton. The change from low to high velocities is abrupt and coincides with the main rift border faults, indicating that thermal modification of the uppermost mantle in Kenya is confined to the rift (Dawson, 2008).

In northern Tanzania, an elongate Bouguer gravity low is almost continuous to the east and south side of the Sonjo-Eyasi fault and is interpreted as due to large thicknesses of low-density lavas that ponded against the fault escarpment (Fairhead 1976). Other large gravity lows occur in the Kitingiri Basin and in the Mponde graben (Ebinger et al. 1997).

5.2. Power Spectrum Results

The power spectrum diagram analysis allows the estimate of general approximate depths of sources. The Bouguer anomaly results (Figure 7a) shows the deepest anomalies between 30 and 38 km (base of local lower crust), intermediate sources between 15 and 30 km and shallow sources are detected from surface down to 15 km deep.

The vertical derivative results were obtained by the derivation of Bouguer magnetic field and they are expected to evidence shallower sources. The power spectrum diagram analysis for the vertical derivative (Figure 7b) shows the deepest sources from 20 to 30 km and the shallower sources are at a maximum depth of 12 km, nonetheless, most of the sources, represented by

gravity contrasts, are indicated to be located between 8 and 12 km.

5.3. Euler Deconvolution Results

The goal of using Euler Deconvolution in this study is the identification of planar structures as faults or major terrain boundaries in the crust. Euler deconvolution datasets were generated with structural index (i) equal to 0, ideal for planar structures. Depth tolerance to source (t) was tested with values 10 and 15, indicated for regional studies. Window source distances (w) values tested were 3,5,7,15. Frequency histograms were generated for the datasets of each combination (Figure 8) in order to analyze the data and discard unfit results.

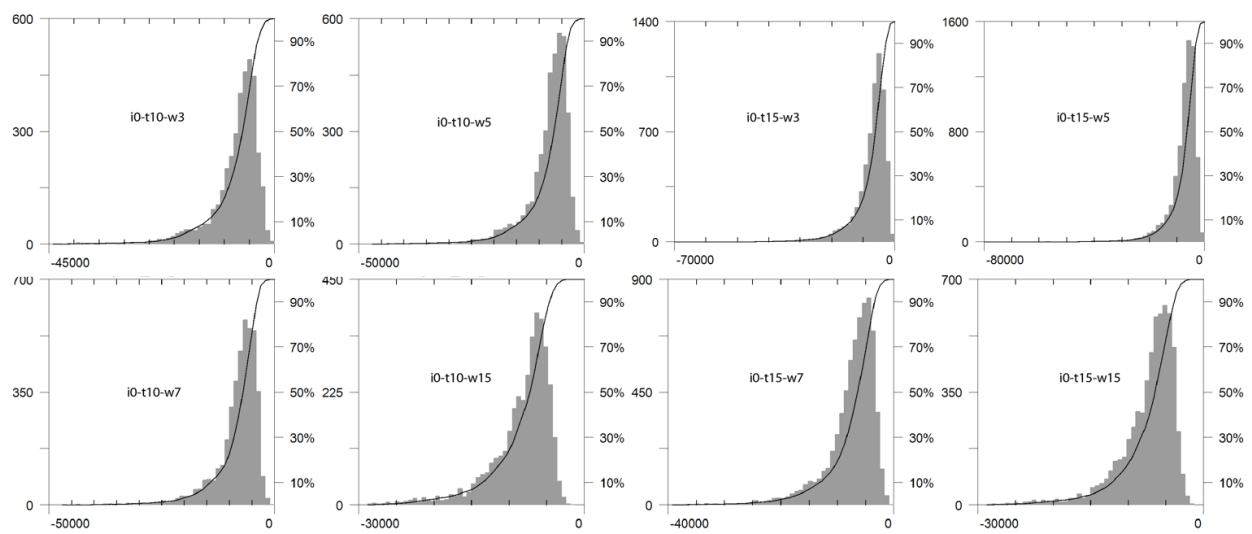


Figure 8 – Frequency histograms for the Euler deconvolution solutions datasets. i = index; t =time; w =window. X axis represents depth in km and Y axis represents the frequency of solutions per depth in number and percentage.

Only combinations $i0-t10-w15$, $i0-t15-w7$ and $i0-t15-w15$ did not extrapolate the mean 40km crustal depth, calculated by seismic refraction in the region (Soares et al. 2006). The other solution datasets were discarded once focus is set on crustal structures.

The frequency histograms of the selected solution datasets (Figure 9) were analyzed for depth intervals. The main criteria for interval distinction were: 1. Major vertical steps on histogram; 2. Frequency curve inflection; 3. Power spectrum source depth results; 4. Accepted depth intervals of brittle-ductile crust transition of 10-15 km for regular crustal temperatures (Sibson 1977) and 5. Mean depth of fault throw in recent similar rifts of 8-11 km (Bell et al. 2011).

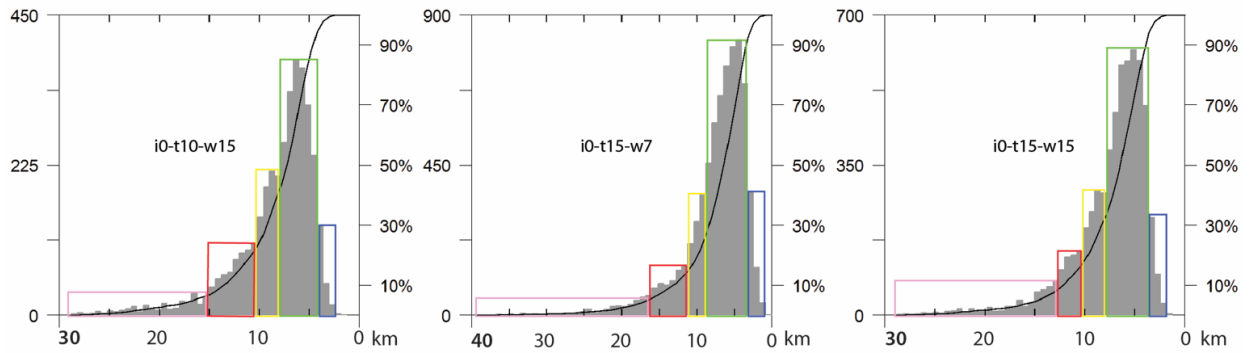


Figure 9 - Selected frequency histograms. Colored boxes represent the main depth intervals. Horizontal axis represents depth in km. Vertical axis represents the frequency of solutions per depth in number and percentage.

The selected frequency histograms yielded five main depth intervals. Since Euler solutions are approximations, the approximate mean minimum and maximum values of each interval from the three histograms were pondered and the depth intervals and their percentage of occurrence were established. In order not to lose data, two more intervals were created, the first one to cover the shallowest solutions and the last to cover the deepest, as follows: **1.** 0 to 2000 m (2%); **2.** 2000 to 4000 m (5%); **3.** 4000 to 8000 m (55%); **4.** 8000 to 11000 m (16%); **5.** 11000 to 15000 m (14%); **6.** 15000 to 20000 m (6%); **7.** > 20000 (2%).

The relative frequency diagram of Euler solutions in each interval (Figure 10a) were plotted over the Sibson (1977) diagram of major fault zones conceptual model (Figure 10b). The diagram reveals an abrupt variation in fault throw which can be interpreted as the Elastic-Frictional/Quasi-Plastic transition. It is important to keep in mind that these are present-day depths and by the time of rifting, mantle ascendance and magmatism were heating the crust, so there is chance that the transition was shallower. Depths by the time of rifting might have been different.

Another remark is that Sibson's diagram presents mean general depth values and depending on the type of faulting, depths may change, for example, for a given thermal condition, the EF-QP transition occurs significantly higher in thrusts than in normal faults (Sibson 1977). The mean depths of fault throw of 8-11 km measured by Bell et al. (2011) might correspond to the Araí Paleorift present-day 4 to 8 km interval of fault concentration, if accounted 3 km of denudation, which is not much considering that crystalline basement under basal rift sediments is exposed at an altitude of ca. 850m.

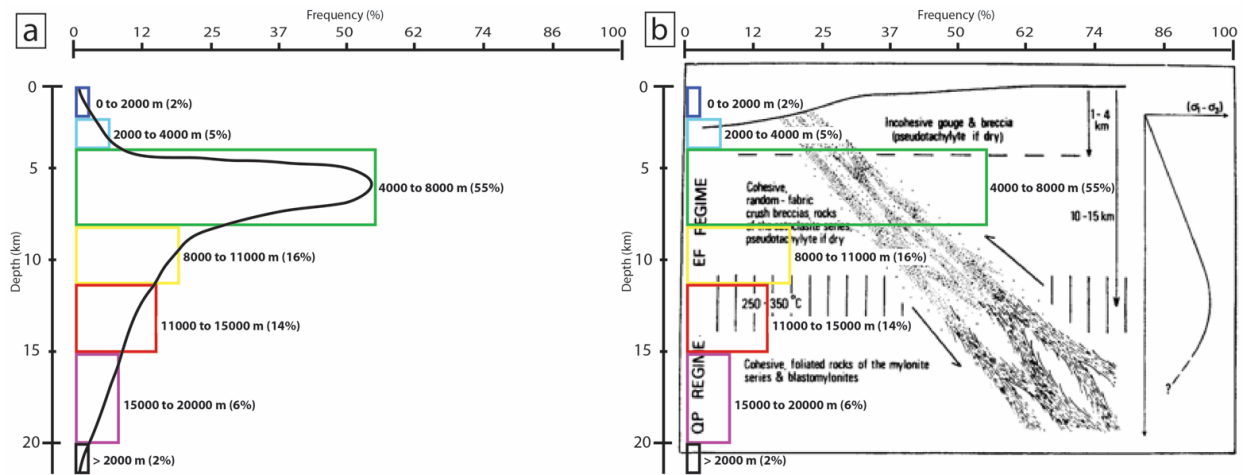


Figure 10 - Interpreted Euler fault depth intervals a) frequency histogram showing frequency curve and b) same histogram plotted over Sibson's conceptual model of a major fault zone (Sibson 1977).

5.4. Euler Solutions Interpretation

Preliminary analysis of the three Euler solution datasets has shown that the solutions contain complimentary information, and as such were plotted together, overlapping each other and summing information. In order to better understand the significance of these combined Euler solutions, the results were compared to Bouguer anomaly map and geologic contacts (Figure 11).

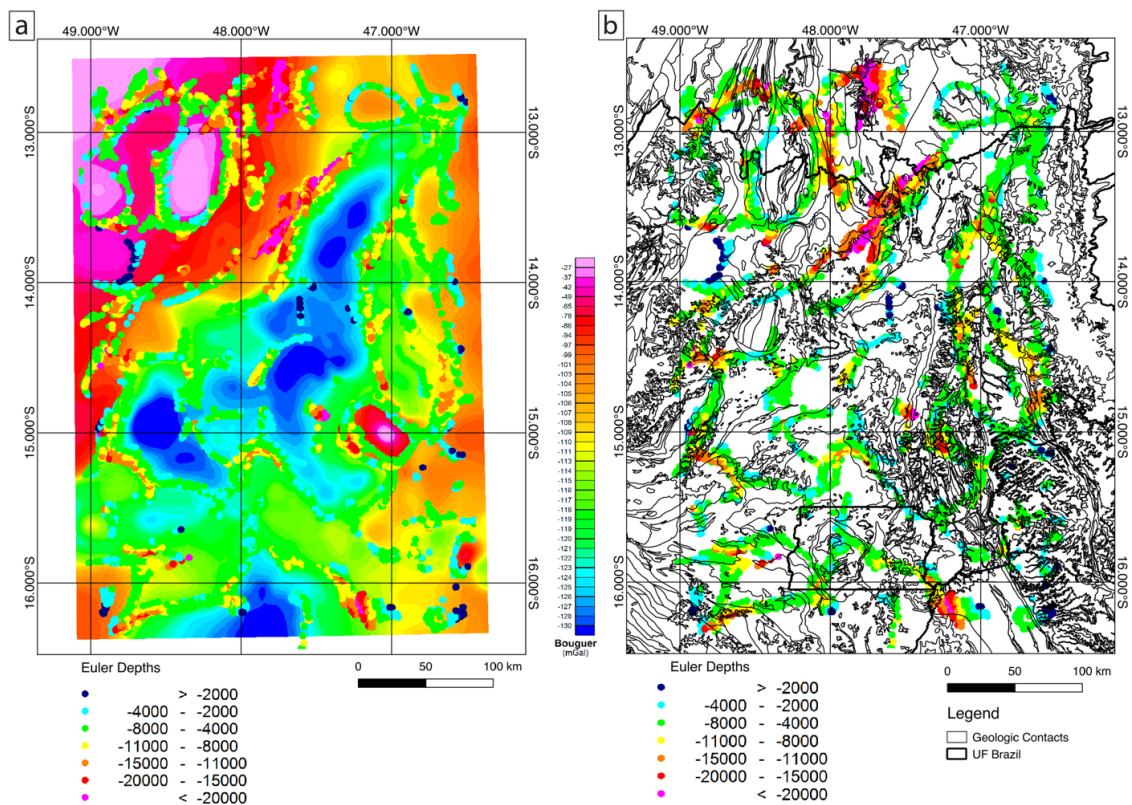


Figure 11 - Combined Euler solutions a) over the Bouguer map and b) under geologic contacts (CPRM 2008).

Many Euler Solutions do not have a definite surface expression (Figure 11b), including many aligned solutions at different depths and possible non outcropping mafic intrusions represented by some of the Euler solution rings and warm color ellipses/circles in the Bouguer map (Figure 11a). Nonetheless, some solutions do match surface geology, specially where crystalline basement is exposed.

Aligned solutions at different depths mark planar density boundaries and may represent: 1. Faults with sufficient throw to generate density contrast by leveling less dense shallow crust with denser composition deeper crust, and/or 2. long and narrow, low-density intrusive rocks, formed as a result of rifting processes. The Rio Maranhão Complex (mafic-ultramafic) shows a perfect Euler Solution match, circled by solutions indicating density contrast with metasedimentary country rock down to depths of 11 km.

According to Butler et al. (1997), the correlation of fault patterns within the upper crust with geophysical lineaments that reflect underlying basement heterogeneities or exposed structures may be the only indicators of deeper faults. Based on that, a procedure for the mapping of possible rift structures was established in the following order: 1. Euler lineament extraction of aligned solutions deeper than 4000 m (Figure 12a); 2. Euler lineaments superposition to geologic map; 3. Adjustment of lineaments according to basement geological heterogeneity and sedimentary cover stratigraphy (Figure 12b) and 4. Interpretation of paleorift general structure and final adjustments based on the Total Horizontal Gradient Amplitude map (Figure 13a) and the Analytic Signal Inclination map (Figure 13b). Finally, an interpreted rift map is proposed (Figure 14) where grabens and horsts are identified and named after the main local cities and rivers.

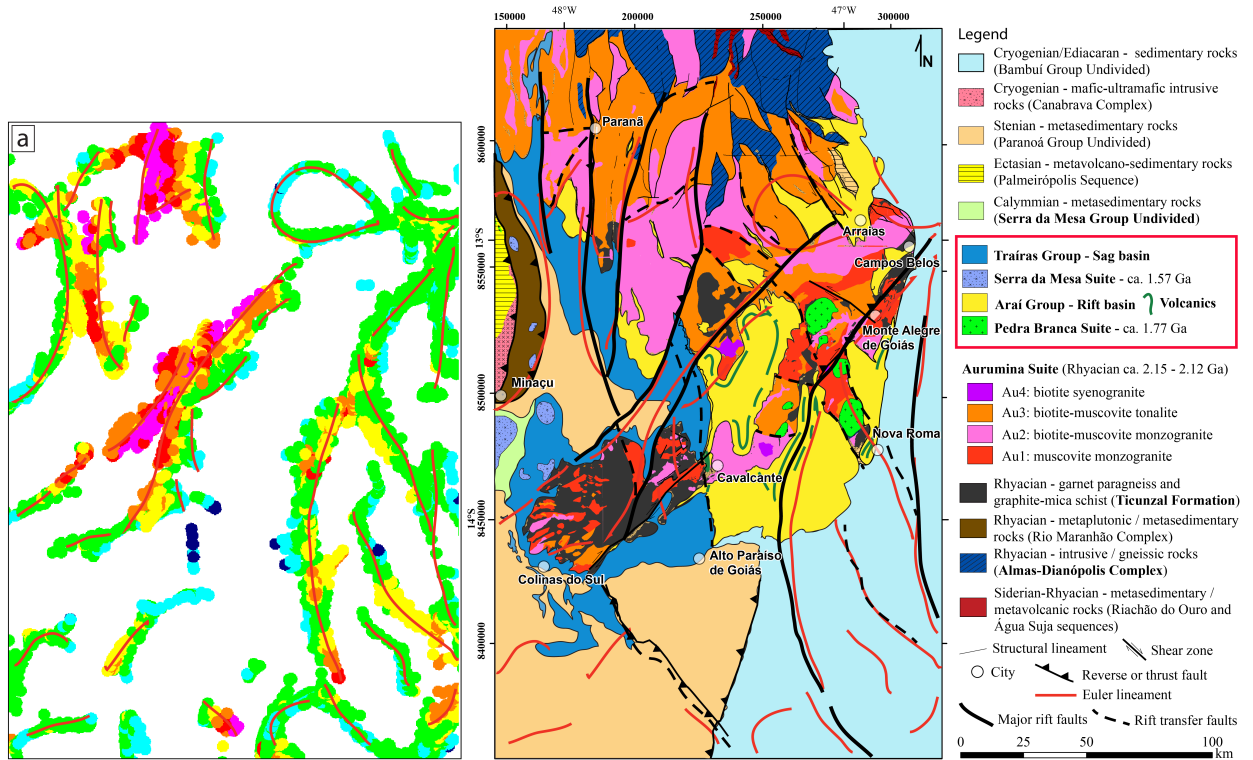


Figure 12 - a) Euler lineament extraction (red lines) plotted over Euler solutions; b) Geologic map with Euler lineaments (red lines) and the resulting adjusted and interpreted rift structures (solid black lines indicate interpreted main rift faults and dashed lines indicate interpreted rift transfer faults).

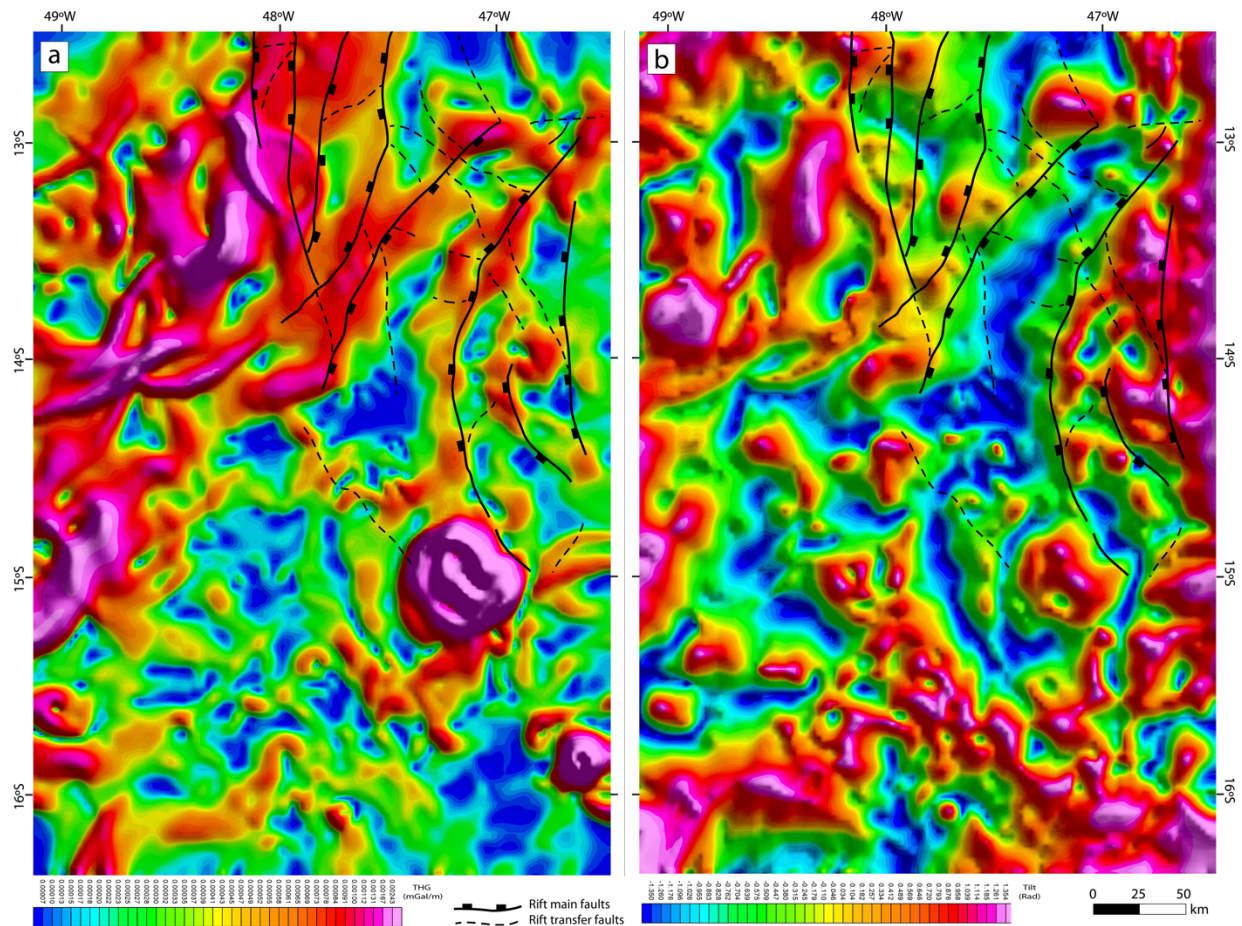


Figure 13 - Derivatives from the Bouguer image. a) Total horizontal gradient - THGR; b) Analytic signal inclination - Tilt from Bouguer image.

The method was successful for the region where basement and rift rocks are exposed, since geophysical data, rift stratigraphy and basement heterogeneity could be tied together, leaving little error margin for the proposed work scale. Nonetheless, future refinement work in more detailed scale is recommended. The continuity of hidden rift structures to the south, where rift sediments are covered, can also be inferred in future works by comparing geophysical patterns to the ones achieved in this work. One significant high density anomaly is present at approximately 15°S; 47.3°W which may represent mafic intrusions not cropping out. The intrusion position relative to the interpreted rift faults leads to the interpretation as a non-outcropping rift-related mafic intrusion, related to the mafic volcanism observed to the north. Another possible non outcropping mafic intrusion is detected at 12.8°S; 47.2°W, near Arraias town.

5.5. Description of Main Arai Paleorift Structures

The investigation reported in this work resulted in the positioning of principal paleorift structures, relative to the local geologic map, constrained by our interpretation of geophysical, geologic and stratigraphic data (Figure 14) and is described below.

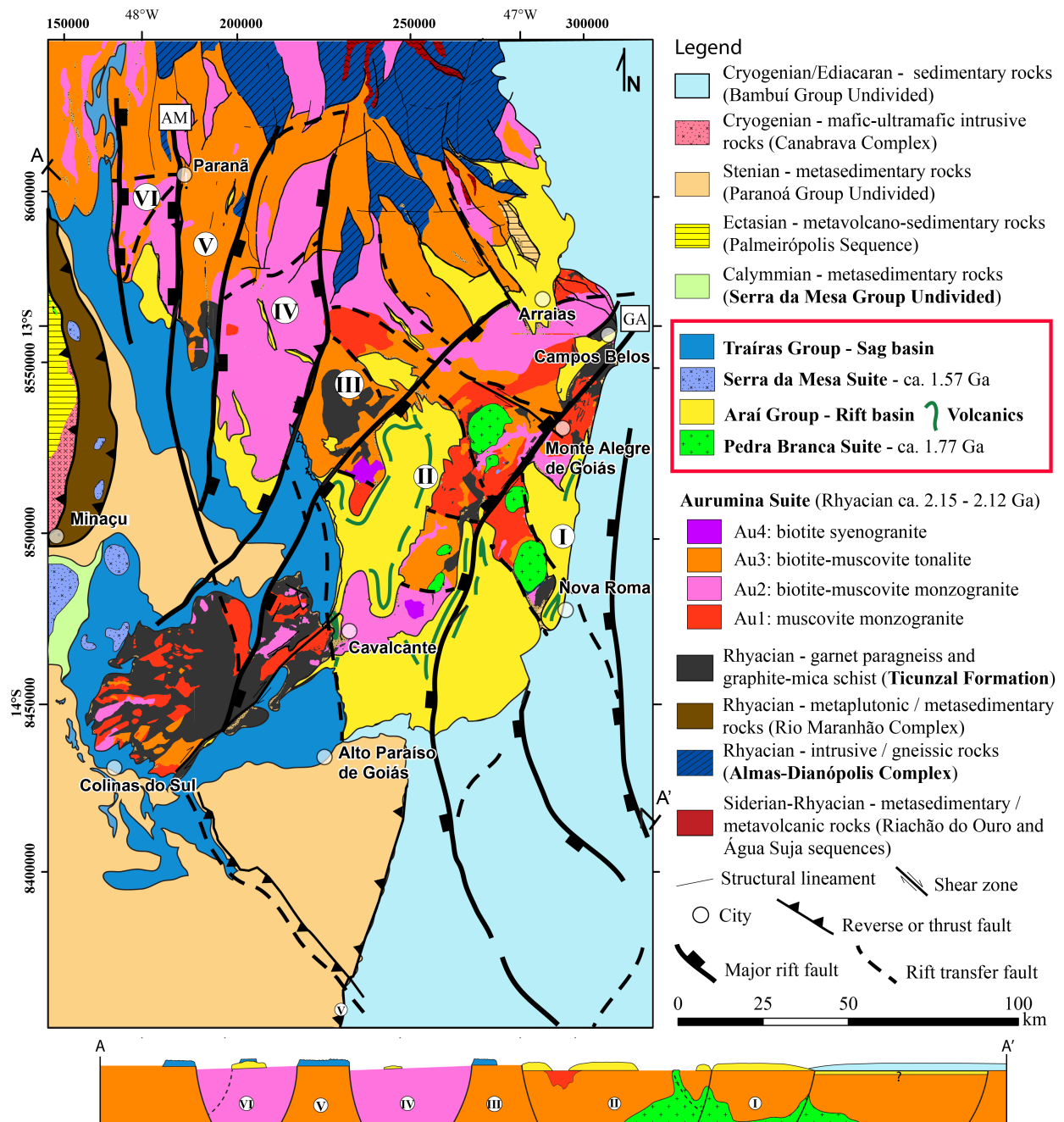


Figure 14 – Geologic map showing interpreted results of main rift faults and transfer faults. I = Nova Roma Graben; II = Arraias-Cavalcante Graben; III = Bezerra Horst; IV = Prata Graben; V = Paranã Horst; VI = Paranã Graben; AM = Água Morna Fault; GA = Goiana-Aboboreira Fault. The location of A-A' profile is indicated.

The crustal blocks where the Traíras Group is deposited directly over the crystalline basement were interpreted as horsts. The horst blocks are predominantly composed of Au3 and show greater occurrence of Ticunzal Formation rocks, which are remnants of Aurumina Suite country rock that probably represent roof pendants and mega xenoliths. The blocks where greater thickness of rift sediments were deposited, especially where clast supported conglomerates and bimodal magmatism occur, were interpreted as grabens. Between the mapped graben and horst

blocks, abrupt variations of basement lithology occur, and Euler Solutions indicate that most of these variations result in density contrasts that continue deep in the crust. These deep abrupt lithologic variation were interpreted as controlled by rift major faults. However, inside these grabens and horsts, abrupt variations of basement lithology also occur, and were interpreted as controlled by transfer faults, common features in any type of rift system that may produce relative topographic highs in grabens and relative topographic lows in horsts. It is possible that some of these mapped grabens were in fact hemigrabens, but the work scale and the presently available data do not allow that degree of detail in our interpretation.

The main loci of volcanic dikes and plugs occurrence lie along the interpreted fault that divide the Nova Roma and Arraias-Cavalcante grabens, in this study called Goiana-Aboboreira (GA) Fault (Figure 14). In surface, that fault matches, to the north, the Serra da Goiana hill, a long and linear elevation ca. 50 m formed by quartz-sericite mylonite, previously interpreted as of orogenic origin for cutting the Araí rocks and indicating shear movement (Alvarenga et al. 2007). These same mylonites are found at the Aboboreira and Boqueirão hills, accompanied by the greatest thickness of rift-related rhyolites (Alvarenga et al. 2007).

Considering that these linear hills also coincide with the interpreted fault trace to the south, and rhyolitic lavas do not travel long distances, the rhyolite occurrence lead to the interpretation that this is one of the main rift structures and mylonite occurrence suggests that rift faults were likely reactivated in the Neoproterozoic Brasiliano orogeny and reworked as shear zones. Furthermore, these considerations lead to the conclusion that the Goiana-Aboboreira Fault marks the Araí Paleorift volcanic center.

The Nova Roma Graben (I) is partially covered by Ediacaran Bambuí Group rocks. Euler solutions suggest that its eastern fault is not deep nor continuous, indicating that it might be a tilted-to-the-west hemigrabem. It is comprised mainly of Au1 and Au2 monzogranite with several occurrences of preserved Ticunzal rocks, all intruded by the Pedra Branca Suite anorogenic granites. Great occurrences of volcanic rhyolite and pyroclastic deposits occur in this graben. Arraias Formation rift sediments are thick and clast supported conglomerates abound.

The Arraias-Cavalcante Graben (II) is wider and longer than the other arms, presents a curved shape and hosts the thickest rift sequence sediment including rhyolitic lava flow levels up to 100 m thick as well as local pyroclastic deposits and minor mafic volcanic occurrences. The basement alternates between Au1, Au2 and Au3 sometimes by abrupt change, apparently controlled by faults. Ticunzal Formation rocks are widely present bordering the Goiana-Aboboreira Fault. Together, the Arraias-Cavalcante and Nova Roma grabens form the Araí

Paleorift main feature, where all the known anorogenic rift-related intrusives occur. These plutonic bodies that are now exposed at the surface seem to be controlled by rift main faults and transform faults, as are the related extrusive rhyolitic rocks, that occur interlayered to rift sediments, suggesting sin-rift magmatic activity. Under the Arraias-Cavalcante graben, an NNE-SSW elongated low-density anomaly occurs, as can be seen in the Bouguer anomaly map (Figure 5) and its derivatives (Figure 13).

The other Araí Paleorift grabens and horsts represent minor features, with much less volcanism, no record of rift-related anorogenic plutonism and might represent late structures generated to accommodate the progression of the extensional process. Together, the three grabens form a fan-like structure, as if rifting was spreading. These minor features are described below.

The Bezerra Horst (III) widens to the northwards and subside through transfer or subsidiary faults, which seem to control basement abrupt lithologic variation, apparently becoming a wide paleogaben where rift Arraias Formation sediments are sparsely preserved over the basement, forming topographic elevations.

The Prata Graben (IV) is the one with less available information, lacking more detailed geologic mapping. Nonetheless, it is predominantly overlain by Au2 basement rocks and seems to narrow northwards, where a transfer fault, well marked in the Euler Solutions, seems to be responsible to expose Au3 rocks. Its rift sequence is well preserved in the south, under the Traíras Group, that is directly deposited on the crystalline basement on both east and west graben shoulders (or adjacent horsts).

The Água Morna Fault (AM) separates the Paranã Graben (VI) from the Paranã Horst (V). Euler Solutions indicate that this fault can be as deep as 20 km. Along its trace a silicified long and linear N-S hill outstands ca. 150 m in the flat basement and occasional warm water springs occur, suggesting that the fault discontinuity is locally spaced allowing meteoric water to flow deep into the crust and return heated by.

The Paranã Horst (V) is comprised mainly of Au3 tonalite and contain great areas of preserved Ticunzal Formation rocks. The Traíras Calymmian sag sequence is here deposited directly over crystalline basement and Ticunzal rocks

The Paranã Graben (VI) is similar to the Prata Graben, but narrower, it accommodates little volcanics, but a thick continental sedimentary sequence, full of internal unconformities suggesting a deep and narrow paleogaben with a polycyclic subsidence history. As in the Prata Graben, the rift Arraias sequence is partially eroded and covered by the Traíras Formation,

which is deposited directly over the crystalline basement on both east and west graben shoulders (or adjacent horsts).

The erosive unconformity between rift Araí Group and sag Traíras Formation indicates uplift and erosion between 1.75 and 1.55 Ga, but no record of compressive tectonics in that age span was ever found, suggesting some mechanism of isostatic uplift, possibly related to the expressive low-density anomaly under the Araí paleorift.

5.6. Araí Paleorift Tectonic Model

Once the correlation of fault patterns within the upper crust with geophysical lineaments has been established, a rift tectonic model analysis is attempted by analyzing the integrated data and comparing the Araí Paleorift to other recent and ancient global examples.

If compared to other examples of paleorifts, volcanic activity is scarce in the Araí, lacking expressive magma volumes as reported in other Statherian rifts as in Siberia (Guryanov and Peskov 2017) and North China (Lu et al. 2002). In the Araí Paleorift, pre-rift volcanism is not recorded and maximum thickness for rhyolites in the Araí rift rarely surpasses 100 meters (Alvarenga et al. 2007) and much less for basalts.

Passive rifting is a result of tensional intraplate far-field forces transmitted within the lithospheric plate (Fleitout et al. 1986), lithospheric stretching induces mantle upwelling (McKenzie 1978), which in turn produce melting responsible for anorogenic volcanism. Alternatively, active rifting or plume rifting, occurs as a result of dynamic stress caused by mantle diapirs or sheet-like mantle upwelling (Turcotte and Emerman 1983), that advect great amounts of heat, sufficient to produce large amounts of surface volcanism (Sengör and Burke, 1978). In general terms, passive rifts are magma poor and active rifts are magma rich (Sengör and Natal'in 2001).

The passive rift model for the Araí Paleorift seems to be best fit. Active or even mixed plume-craton interaction models (Koptev et al. 2016) would require volcanism-rich chronocorrelated rifts nearby, but low volume of volcanism is also reported in the Lower Espinhaço sequence, that represent the Araí eastern counterpart in the São Francisco Craton (Figure 17b), and are also interpreted as fitted on the passive rift model (Danderfer Filho et al. 2015). The Araí Group pre-rift sequence, Água Morna Formation, is also evidence of broad subtle subsidence, characteristic of passive rifting mechanism.

5.7. Origins of Anorogenic Magmatism

According to Kearey et al. (2009), there are three ways in which the mantle may melt to

produce basaltic liquids beneath rifts. 1. By heating the mantle above the normal geotherm in perturbations related to deep mantle plume underplating; 2. by lowering the ambient pressure via ascent of hot mantle during lithospheric stretching or also by the rise of a mantle that leads to decompression melting at a variety of depths and 3. by the addition of volatiles, which has the effect of lowering the solidus temperature. A combination of all these three mechanisms probably contribute to generation of basaltic melts beneath continental rifts.

Once formed, the composition of mafic magmas may be affected by partial melting. This process results in the separation of a liquid from a solid residue, which can produce a variety of melt compositions from a single mantle source. Primary mafic melts also tend to fractionate, whereby crystals are physically removed from melts over a wide range of crustal pressures, resulting in suites of compositionally distinctive rocks. Current models generally favor fractional crystallization of basaltic melts in shallow magma chambers as the dominant process that generates rhyolite. Compositional variability also reflects the assimilation of crustal components and magma mixing. The bimodal basalt-rhyolite eruptions are thought to reflect combinations of mantle and silica-rich crustal melts.

As previously mentioned, the Pedra Branca Suite varies from metaluminous to alkaline and is enriched in incompatible elements (Botelho and Moura 1998), being thus less dense than their country rock, represented in the superior crust mainly by the Aurumina Suite tonalite and biotite granite. Dated ca. 1.77 Ga (Pimentel et al. 1991), the Pedra Branca Suite is considered to be associated to rapakivi granites (Bilal et al. 1997; Botelho and Moura 1998; Lenharo et al. 2002). Coincidentally, Rämö and Haapala (1995) conclude that most of the rapakivi granites are Proterozoic (mainly 1.0 to 1.7 Ga) and their magmatic association is bimodal, showing characteristics of A-type granites. They state that these anorogenic granites may occur as very large (up to 40,000 km²) batholiths and some complexes host important Sn-polymetallic deposits, which is the case of Pedra Branca Suite.

The rapakivi granites generally crystallize from relatively hot, restite-poor magmas at low pressure, in epizonal-subvolcanic zones, but show isotopic composition compatible with a lower crustal protolith, moreover, pressures of 7 to 10 kb are required to explain the low H₂O deficiency of these magmas, indicating a protolith depth of 25 to 36 km (Anderson and Cullers 1978; Rämö 1991). It is consensus that anorogenic magmas form at the lower crust and ascend diapirically to the upper crust, eventually reaching the surface.

As mentioned before, a low-density body was already interpreted over seismic data by Soares et al. (2006) and allocated at 30-35 km in the crust (Figure 4). The gravity residual maps

generated by upward continuation (Figure 15a) as well as the Euler Solutions, suggest that this body has greater dimensions than previously proposed and might occur as an expressive granite body from surface down to depths of 19 km, as indicated by the Voxel 3d map (Figure 15b). These results show that the low-density anomaly under the Araí Rift is continuous to depths of 19 km and coincides with the anorogenic Pedra Branca granites and their rhyolitic volcanic phase at surface.

The density contrast of this low density body with the country rock becomes greater with depth crust which might explain the well-defined anomaly up to 19 km deep. This deep occurrence of low-density rocks identified in this work, generally crystallized in much shallower depths than 19 km, might be explained by the great decompression generated by the Araí Paleorift recurrent faulting. As can be seen on figures 6, 10 and 11, the Araí Paleorift faults were detected in this study to a maximum depth of ca. 38 km but generating significant fault throw (and consequent pressure release) up to 20 km deep. After that depth the faults die out, become sparse and do not generate great density contrast, meaning that fault throw is not significant.

The melt source of these granites is inferred to be at greater depths, near inferior zones of the lower crust, below 30 km, as modeled by Soares et al. (2006) and indicated as source zone for anorogenic granites by Anderson and Cullers (1978) and Rämö (1991). At the source, restite of fusion might have been left to generate the low-velocity body shown on seismic data. These rapakivi granites do not need a peraluminous source as suggested by Soares et al. (2006), being formed from basaltic magmas in the lower crust (Kearey et al. 2009; Anderson and Cullers 1978; Rämö 1991). The Aurumina Suite might have contributed in part to the formation of anorogenic magmas by shallow contamination and might even be one of the possible explanations for the predominance of rhyolites over basalts and granites over mafic intrusions, however, Aurumina orogenic granites are very unlikely to be found below 30 km, where the source for the rapakivi granites is most probable.

The presented data strongly indicate that the source for the deep portion of the negative Bouguer anomaly under Araí Rift must be the Pedra Branca Suite magmas that were impeded to ascend from the inferior crust once the rift was aborted. At ca. 1.77 Ga the Aurumina Suite was exposed at the surface, extension lead to rifting and generated fusion at the lower crust, during the ascent, felsic magma might have left behind voluminous bodies of low density material from its birth place up to the surface. With limited access to the surface, the partial melt might have partially crystallized in depth, being up to present day lodged in the crust under the Araí Paleorift.

The best-fit interpretation for anorogenic magma generation in the Araí Paleorift is partial

melting as a result of crustal extension decompression and mantle upwelling, whereas the plume underplating rift model is discarded once it would require much larger volumes of volcanism, including basaltic volcanism, which are restrained.

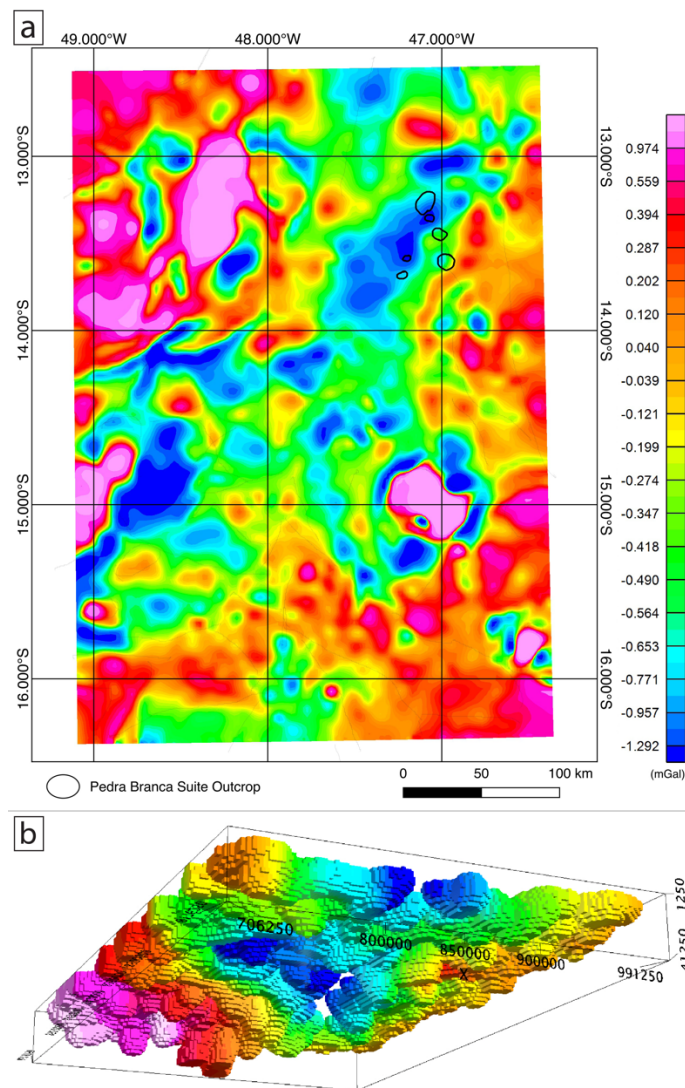


Figure 15 – a) gravity residual map for upward continuation at 1000 m elevation indicating the outcrop location of anorogenic granites from Pedra Branca Suite and b) voxel 3d for the same area showing maximum depth of anomalies.

5.8. Present-day Local Topography

The region where the Araí rift rocks and crystalline basement are exposed is being submitted to uplift and erosion. The highest crystalline basement altitudes reach 850 m and are coincident to the area where a long and narrow low-density Bouguer anomaly occurs. Not influenced by the studied low-density anomaly, crystalline basement altitudes rapidly fall to 500 m to the east, 400 m to the north and down to near sea level to the south, this last altitude inferred by the thickness of the sedimentary covers, once basement is not exposed. Seismic data indicate that this Bouguer anomaly is produced by a low density body located in the lower crust.

Gravity data indicate that this low-density anomaly is also present in the upper crust and stratigraphic data show that this anomaly is coincident to the main Araí rift axis, a strong evidence that the rifting process that started ca. 1.8 Ga ago has produced a low density body in the upper and lower crusts that persists to this day. Present-day high topography is most probably supported by crustal buoyancy, as an isostatic response to these substantial volumes of low-density material within the crust and upper mantle revealed by gravity maps and seismic refraction models.

This Cenozoic uplift is being investigated by zircon and apatite thermochronology by the authors. Preliminary data indicate that after the Brasiliano orogeny, when the thermochronologic clocks were reset, the region has been submitted to practically constant uplift, exposing the crystalline basement that was buried down to estimated depths of between 3 to 5 km in the Cryogenian. Recent uplift has generated great valleys with topographic unevenness locally as high as 600 m with plenty of water springs and waterfalls on sharp scarps, giving shape to the famous landscapes of Chapada dos Veadeiros region. The region is inserted in the Brazilian Central Plateau and its highest altitude, 1676 m, is also positioned over the Araí Paleorift low-density anomaly. That altitude is considerably high for an intracontinental region that was uplifted over half a billion years ago and has not seen great tectonic events ever since, so that the subject demands a more detailed investigation.

5.9. Global Correlation to Other Statherian Rifts

Guryanov and Peskov (2017) have studied a paleorift in the Siberian Platform that presents similar age, structural geometry and mineralogic characteristics to the Araí paleorift. They describe the Statherian Ulkan paleorift as having a three-armed geometry and records of intense intraplate alkaline-granitic magmatism during late Paleoproterozoic time and concentration of ore deposits aged between 1.72 and 1.67 Ga, (Be, Ta, Nb, Zr, TRy, TRce, U, Au, Li, Sn, W, Mo, Ti and (Pt)). Rare-metal and rare earth deposits are in essence the product of rare metal-alkaline-granitic magmatism and associated with the evolution of the mantle source related to plume activity. due to hotspot activity. They interpret the volcanics and granitoids of the Ulkan Complex as formed in an extensional intracontinental environment accompanied by rising of the mantle plume. The Ulkan alkali-leucogranite massif is interpreted as a hotspot-type diapir. The area of convergence of basite dyke swarms coincides with the area of development of alkali-granite magmatism and alkali metasomatism at the center of the massif was a likely epicenter of the hotspot.

In the interval between 1.77 and 1.70 Ga, rift-related anorogenic magmatism is also

recorded along the southeastern and southwestern flanking of the Siberian platform of the North Asian Craton (Yu et al. 1991), in the North American platform (Rämö and Haapala 1995 - Nueltin, M'Clure Bay and Cap Farewell massifs) and in northern China (Lu et al. 2002).

5.10. Regional Correlation in the São Francisco Craton

The aborted Arai rift could be interpreted as the NNE arm of a triple junction expected for a complete rift system. The Pirapora aulacogen would be a strong candidate for the SE arm of this system, since, according to Alkmim et al. (2012), seismic profiles show that the aulacogen is filled by an unknown sequence, previous to the Stenian/Tonian Paranoá Group, and is separated from it by an unconformity. Other rift segments may occur under the Brasília Belt Meso and Neoproterozoic sedimentary covers. In a typical triple junction model, with arms at 120 degrees, a third arm would be expected in the approximate direction EW and the triple junction would be east of the Federal District, where a structural pattern of structural domes, brachi-synclines and brachi-anticlines is well defined.

This hypothesis could be supported by the deep gravimetric discontinuities shown by the Euler solutions in the Federal District area and also by the Paranoá Group paleo-current directions, which accurately coincide with the anomaly direction in the region of Alto Paraíso de Goiás and the direction of the Pirapora aulacogen in the Cristalina town region, whereas in the Federal District area the paleocurrent direction is diffuse, coming from all directions. That scenario leads to the conclusion that the Paranoá basin subsidence was controlled by the Arai Rift structures reactivation, which in any case could be right. However, the triple junction model is best suited for large and successful rifts that occur at continental ocean opening scales. This is not the case with the Arai rift, which is characterized by being a narrow rift, aborted before generating oceanic crust.

By studying global examples of aborted narrow rifts, we see that these occur as a complex system of segments, not obeying the triple junction model and generally favored by older weak crustal zones. The current example that best matches the São Francisco Craton Paleoproterozoic rift system is the East African rift, both in scale and in its main features. The East African Rift is formed by a complex system of rift segments of varying directions that generally surround the Tanzanian Craton (Figure 16a). In a similar way, the Arai / Pirapora / Paramirim aborted rift systems (Lower Veadeiros / Lower Espinhaço sequences) seem to surround the São Francisco Craton (Figure 16b).

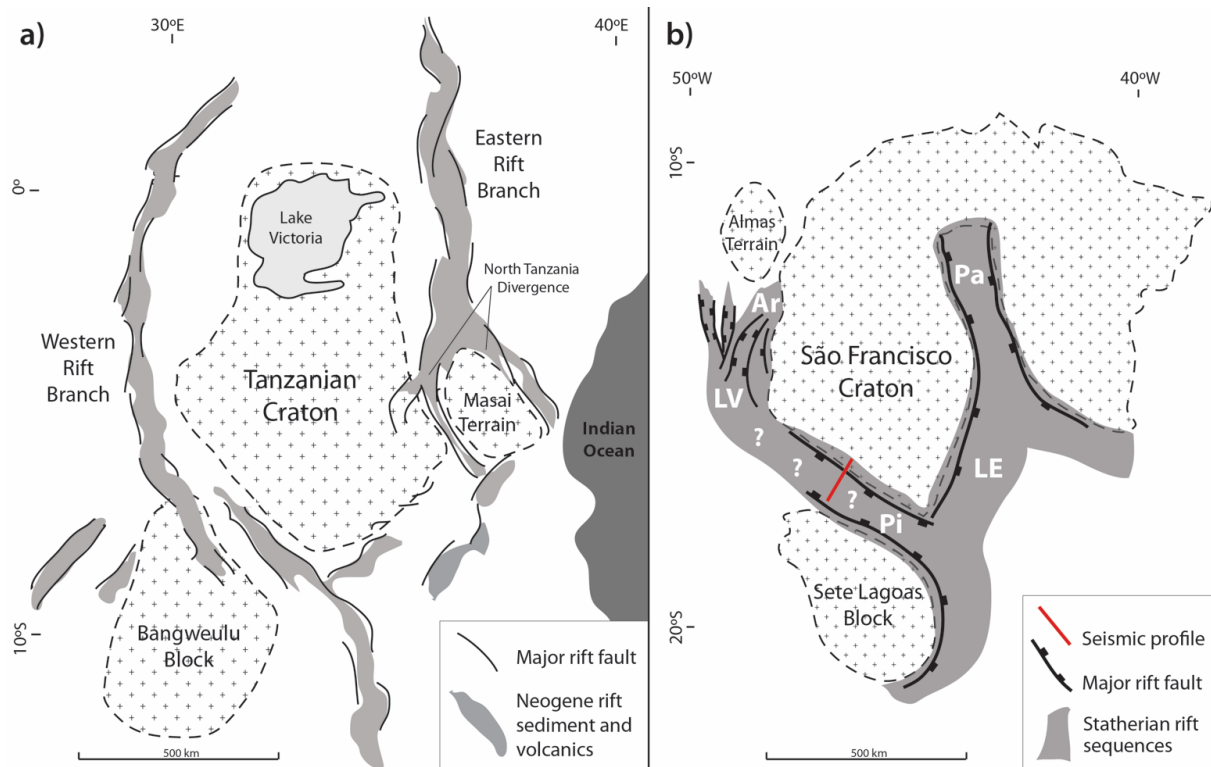


Figure 16 - Analogy between the (a) East Africa Rift system (adapted from Fontijn et al. 2010) and (b) the São Francisco Craton Paleoproterozoic rift system: Ar) Araí rift; Pi) Pirapora Aulacogen; Pa) Paramirim Aulacogen; LV) Lower Veadeiros and LE) Lower Espinhaço (adapted from Reis et al. 2017) the red line marks the location of the seismic section shown in figure 19a, that crosses the Pirapora Aulacogen.

Reis et al. (2017) affirm that the first tectonic stage so far identified in the São Francisco basin corresponds to the opening of the NW-trending Pirapora aulacogen. Cutting across the central portion of the São Francisco basin, this structure hosts the Paranoá-Upper Espinhaço sequence, which unconformably overlies an apparently older unknown succession. This succession is a potential correlative of the lower Espinhaço Supergroup exposed in the Araçuaí belt and Paramirim aulacogen.

In the northern Brasília belt, the Paleoproterozoic metasedimentary successions and associated rocks of the Araí Group are also potential correlatives of these unknown strata. Nevertheless, additional studies are needed to better evaluate the age and tectono-stratigraphic significance of this succession.

The interpretation of the Pirapora seismic line profile (location on figure 16a) allows three main interpretations (Figure 17a): The aulacogen is filled by a sin-rift sequence representing the Arraias Formation (Araí Group), followed by the Paranoá Group on a sag type basin facilitated by rift reactivation; (Figure 17b): The aulacogen is filled by a sin-rift sequence representing the Arraias Formation (Araí Group), followed by the Traíras Group (Martins-Ferreira et al. 2017) on a sag type basin facilitated by rift reactivation and later covered by a

thinner sequence of Paranoá Group.

Strong carbonate reflectors under the base of Bambuí Group could represent either carbonates from the top Paranoá Group or Traíras Formation. A third interpretation is that a lower Paranoá as a local rift sequence and an upper Paranoá as a sag sequence fill the Pirapora Aulacogen (Figure 17c). Opposed to that hypothesis is the fact that rift-related sediments and volcanism were never found in the Paranoá Group, anywhere in the Brasília Belt. Besides, the Statherian extension was the strongest and most widespread rifting event recorded in the São Francisco Craton, showing interconnection evidence to other Statherian rifts, as the Araí and the Espinhaço systems, being thus the most probable to be filling the basal sequence in the aulacogen. By all the above stated, the interpretation represented in figure 17b is the most likely to be closer to reality and matches the proposition described by Guadagnin and Chemale (2015).

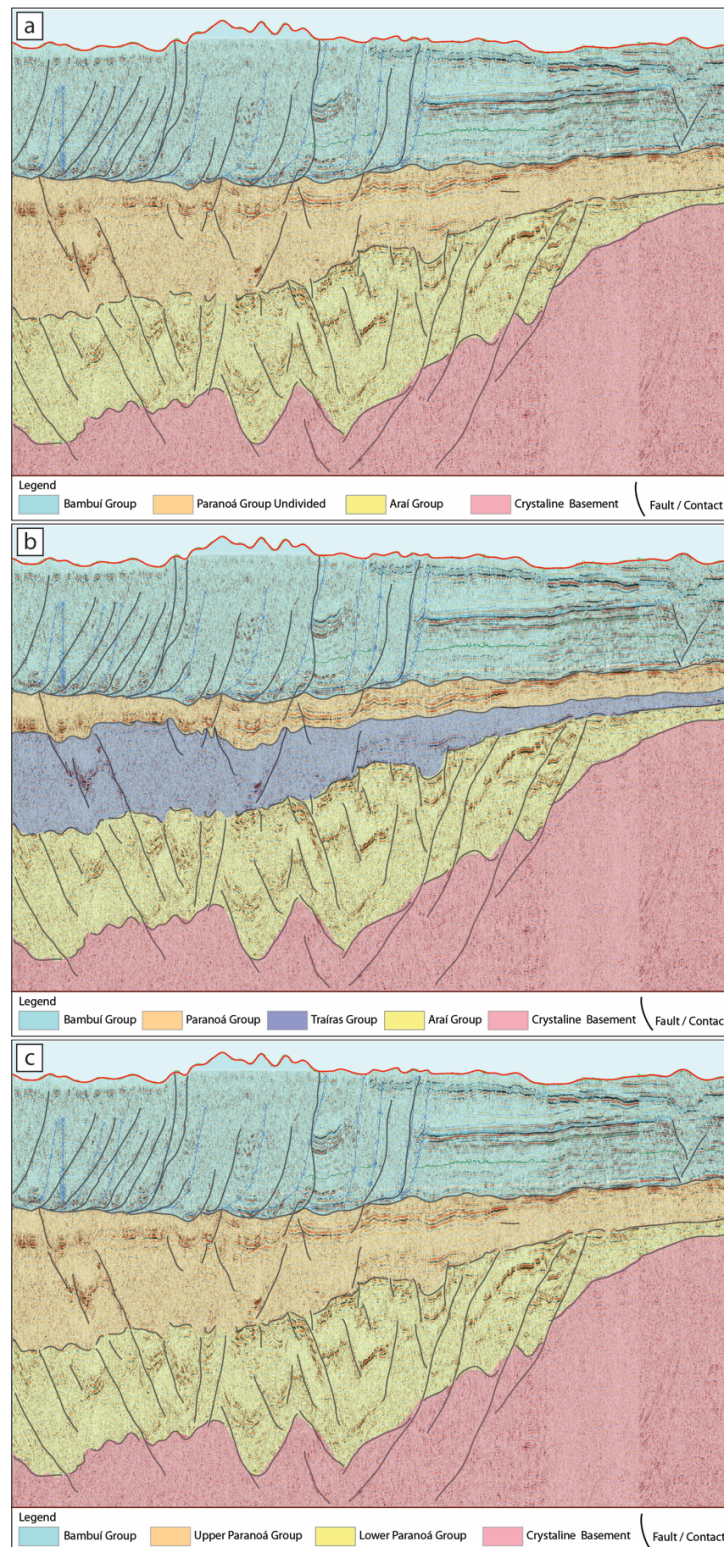


Figure 17 – Interpreted Pirapora aulacogen seismic profile (see profile line location on figure 16b). a) Filled by undivided Araí and thick undivided Paranoá sequence; b) filled by Araí Group rift sediments followed by sag sequence Traíras Group and a thinner Paranoá sequence. c) Upper and Lower Paranoá sediments filling the aulacogen, separated by an unconformity.

The East African rift system consists of narrow and segmented basins, which locally open out in fan form or form multiple parallel rifts in broader rift zones reaching up to 400 km in

width. The largest extensions occur in the northern region, Afar and Ethiopia, but are reduced to the south, where the rift surrounds the Tanzania Craton. In northern Tanzania, the eastern branch of the East African Rift becomes broader and splits into three branches, forming the North Tanzania divergence zone (Figure 16a). One of these branches enters the craton of Tanzania Craton reworking previous orogenic structures (Dawson, 2008). Delcamp et al. (2016), account that the precise origin of the divergence is still debated, but it is thought to be linked to interactions between rifting processes and the presence of the ancient, thick, and cold Archean Tanzanian craton and the Massai Terrain together with the weakness offered by the Mozambique Belt.

The distribution of the principal faults mapped in the Araí rift shows that it also opens in three branches in a fan forming system of grabens and horsts whose volcanic center is located at the main branch and coincides with the eastern edge of the low gravimetric anomaly of Arraias-Alto Paraíso. The main arm of the Araí rift is curved towards the São Francisco craton and disappears under the sediments of the Bambuí (Neoproterozoic) and Urucuaia (Cretaceous) groups, suggesting that there is a possibility of this arm entering the craton below the Cretaceous sequence. The generation of the other two branches might have been induced by preexisting structural or compositional weaknesses generated during the Rhyacian Orogeny, also known as Transamazonian Orogeny.

Global examples have shown that the rheology of the lithosphere dictates rift structure. Old, cold, thick and strong lithosphere as those found in cratonic interiors, produce narrow rifts with steep faults, e.g. East Africa Rift. Alternatively, warm, young thin and weak lithosphere tend to develop broad rifts with distributed listric faults e.g. Basin and Range, Aegean Sea.

We interpret the Araí Paleorift as transitional between narrow and broad rifts, but tending to the narrow type in a divergent setting. It was probably formed as a continuation of the Pirapora Aulacogen, coming as a narrow rift with steep faults (Figure 17) when crossing the São Francisco Craton (old, cold and strong lithosphere) and transitioning to a broader and more distributed rift, characterizing a rift divergence as it enters the craton western margin. The São Francisco cratonic margins were reworked during late Rhyacian orogeny and were thus relatively young and weak lithosphere in early Statherian (only ca. 200 Ma old), and much thinner than the craton. The North Tanzania divergence formation is suspected to be related to the existence of the Massai Terrain, analogously, the Araí divergence development can be related to the existence of the Almas Terrain, that may have offered resistance relatively to the Rhyacian orogenic belts, inducing the rift to diverge (Figure 16b).

6. Final Discussion and Conclusions

Understanding the controls for basin installation, evolution and cyclic reactivation can be challenging in a thrust belt environment due to deformation and metamorphism. Nonetheless, the results obtained exceeded the expectations and the approach chosen for this study have shown to be a successful method for the investigation of Paleorifts. Rifts tend to develop as a complex system of faults and the structures identified in this work represent the most basic framework of the studied Paleorift. Detailed surveys will certainly reveal a much more complex array of structures, however, this first basic approach presents important advances that may serve as basis for future research involving the Araí Paleorift. Finally, the method developed for this study may be adapted for the investigation of other ancient rifts.

The main conclusions and contributions to the local and regional geological knowledge are listed below:

- The mapping of the Araí Paleorift structures led to the identification of its main features: Nova Roma Graben; Arraias-Cavalcante Graben; Bezerra Horst; Prata Graben; Paranã Horst; Paranã Graben; Água Morna Fault; Goiana-Aboboreira Fault. Other main faults were inferred separating grabens and horsts but could not be correlated so far to known surface features.
- The Goiana-Aboboreira Fault was identified to be the principal volcanic center of the Araí Paleorift.
- The Araí Paleorift can be classified as a passive, three armed failed rift, narrow to divergent type, that produced preferably anorogenic rapakivi-related magmas and subsidiary mafic magmas.
- The Araí rift faults occur to a maximum depth of ca. 38 km, but major fault throw occurs from 4 to 8 km deep and are attenuated at the 8-12 km interval, probably the brittle-ductile transition zone at the time of rifting and practically disappear after 20 km.
- The low-density anomaly under the main rift segment is most likely produced by a large subsurface volume of the rapakivi-type anorogenic granite, lodged in the crust as far as 19 km deep, representing the Serra da Mesa Suite in surface, whereas the orogenic Aurumina granite sources are shallow by nature and cannot have been the source of the anorogenic granites.
- The deep occurrence of low-density magmas was likely caused by decompression generated by rift faulting, which generated pressure release as far as 20 km deep, maximum depth for minimum significant fault throw detected.

- The active rift model by plume underplating is discarded for the lack of more expressive volcanism, especially the mafic type. Instead, far-field passive rift model is adopted.
- The erosive unconformity between rift-related Araí Group and sag-basin Traíras Formation indicates uplift and erosion between 1.75 and 1.55 Ga, but no record of compressive tectonics in that age span was ever found, suggesting some mechanism of isostatic uplift, possibly related to the expressive low-density rocks under the Araí Paleorift.
- Present-day high topography is most probably supported by crustal buoyancy, as an isostatic response to the substantial volumes of low-density rocks within the crust.
- Together, the Araí, Pirapora and Paramirim aborted rift systems seem to form a single rift system that surrounds the São Francisco Craton, similar to the East African Rift system, that surrounds the Tanzanian Craton.
- The overall similarity of the São Francisco Rift system with the East Africa Rift system show that rift-related crustal dynamics in the Paleoproterozoic was not very different from present day processes.

Acknowledgements

The authors are grateful to the University of Brasília (UnB) and the *Observatório Sismológico de Brasília* (ObsSIS) for the laboratorial and financial support, to the *Instituto Brasileiro de Geografia e Estatística* (IBGE) for the gravity data used in this study.

References

- Alkmim, F.F. and Martins-Neto, M.A., 2012. Proterozoic first-order sedimentary sequences of the São Francisco craton, eastern Brazil. *Marine and Petroleum Geology*, 33, 127–139.
- Alrefaee, H. A. 2017. Crustal modeling of the central part of the Northern Western Desert, Egypt using gravity data. *Journal of African Earth Sciences*, 129, 72-81.
- Alvarenga, C.J.S.; Dardenne, M.A.; Botelho, N.F.; Lima, O.N.B.; Machado, M.A.; Almeida, T. 2007. Nota Explicativa das folhas SD.23-V-C-III (Monte alegre de Goiás), SD.23-V-C-V (Cavalcante), SD.23-V-C-VI (Nova Roma). CPRM, Brasília, p. 65.
- Anderson, J.L.; Cullers, R.L. 1978. Geochemistry and evolution of the Wolf River batholith, a late Precambrian rapakivi massif in North Wisconsin, U.S.A. *Precambrian Res* 7: 287-324.
- Babinski, M.; Vieira, L.C.; & Trindade, R.I. 2007. Direct dating of the Sete Lagoas cap carbonate (Bambuí Group, Brazil) and implications for the Neoproterozoic glacial events. *Terra Nova*, 19(6), 401-406.

- Barbosa, V.C.F.; Silva, J.B.C.; Medeiros, W.E. 1999. Stability analysis and improvement of structural index estimation in Euler deconvolution. *Geophysics*, 64: 48–60.
- Bell, R.E., McNeill, L.C., Henstock, T.J., Bull, J.M. 2011. Comparing extension on multiple time and depth scales in the Corinth Rift, Central Greece. *Geophys J Int*; 186 (2): 463-470. doi: 10.1111/j.1365-246X.2011.05077.x
- Bilal, E., Moutte, J., Botelho, N.F., Marini, O.J. and Andrade, G., 1997. Geochemistry of two Proterozoic A-type granites of Goiás state, Brazil: possible links with rapakivi series. *Anais da Academia Brasileira de Ciências*, 69(3), pp.349-365.
- Blakely, R. J., *Potential theory in gravity and magnetic applications*. Cambridge University Press, Cambridge, 1996.
- Botelho, N.F.; Alvarenga, C.J.S.; Meneses, P.R.; D'el-Rey Silva, L.J.H.; 1999. Suíte Aurumina: Uma suíte de granitos Paleoproterozoicos, peraluminosos e sin-tectônicos na Faixa Brasília. In: SBG, Simp. Geol. Centro-Oeste, 7, Brasília, Anais pp.17.
- Botelho, N.F. and Moura, M.A. 1998. Granite-ore deposit relationships in Central Brazil. *Journal of South American Earth Sciences*, 11(5), 427-438.
- Bott, M.H.P.; Hinze, W.J. 2006. Potential field methods. *Developments in Geotectonics, in: Continental Rifts: Evolution, Structure, Tectonics*, Chapter 3C. Olsen, K.H. (Ed) 25, 93-98.
- Brandmayr, E. and Vlahovic, G. 2016. The upper crust of the Eastern Tennessee Seismic Zone: Insights from potential fields inversion. *Tectonophysics*, 685, 1-7.
- Butler, R.W.H., Holdsworth, R.E. & Lloyd, G.E. 1997. The role of basement reactivation during continental deformation. *Journal of the Geological Society, London*, 154, 69-71.
- Campos, J.E.G., Dardenne, M.A., Freitas-Silva, F.H., Martins-Ferreira, M.A.C., 2013. Geologia do Grupo Paranoá na porção externa da Faixa Brasília. *Brazilian Journal of Geology* 43, 461-475.
- Cella, F.; Fedi, M.; Florio, G. 2009. Toward a full multiscale approach to interpret potential fields. *Geophysical Prospecting*, 57(4), 543-557.
- Cooper, G.R.J., and Cowan, D.R., 2008. Edge enhancement of potential-field data using normalized statistics. *Geophysics* 73(3), H1-H4.
- Cooper, S.M.; and Liu, T. 2011. A magnetic and gravity investigation of the Liberia Basin, West Africa. *Journal of African Earth Sciences*, 59(2), 159-167.
- Chekunov, A.V.; Kaluzhnaya, L.T.; Ryabchun, L.I. 1993. The Dnieper-Donets paleorift, Ukraine, deep structures and hydrocarbon accumulations. *Journal of Petroleum Geology*, v.16, 183-196.
- Chen, Q.; Dong, Y.; Cheng, S.; Han, L.; Xu, H.; Chen, H. 2014. Interpretation of fault system in the Tana Sag, Kenya, using edge recognition techniques and Euler deconvolution, *Journal of Applied Geophysics*. doi: 10.1016/j.jappgeo.2014.07.020
- Chenrai, P.; Meyers, J.; Charusiri, P. 2010. Euler deconvolution technique for gravity survey. *Journal of Applied Sciences Research*, 6(11), 1891-1897.

- Cuadros, F.A. 2017. Relações petrogenéticas entre a suíte granítica Aurumina e sua encaixante, a Formação Ticunzal, no setor setentrional da zona externa da Faixa Brasília. Doctorate Thesis. n.136. Instituto de Geociências, Universidade de Brasília. 101 pgs.
- Cuadros, F.A.; Botelho, N.F.; Fuck, R.A.; Dantas, E.L. 2017a. The peraluminous Aurumina Granite Suite in central Brazil: An example of mantle-continental crust interaction in a Paleoproterozoic cordilleran hinterland setting?. *Precambrian Research*, 299, 75-100.
- Cuadros, F.A.; Botelho, N.F.; Fuck, R.A.; Dantas, E.L. 2017b. The Ticunzal Formation in central Brazil: Record of Rhyacian sedimentation and metamorphism in the western border of the São Francisco Craton. *Journal of South American Earth Sciences*, 79, 307-325.
- Curto, J.B.; Vidotti, R.M.; Blakely, R.J.; & Fuck, R.A. 2015. Crustal framework of the northwest Paraná Basin, Brazil: Insights from joint modeling of magnetic and gravity data. *Tectonophysics*, 655, 58-72.
- Danderfer Filho, A.; Lana C.C.; Nalini Júnior H.A.; Costa A.F.O. 2015. Constraints on the Statherian evolution of the intraplate rifting in a Paleo-Mesoproterozoic paleocontinent: New stratigraphic and geochronologic record from the eastern São Francisco craton. *Gondwana Research*, <http://dx.doi.org/10.1016/j.gr.2014.06.012>
- Daniela, G.; Marcos, J.; Araújo, B. 2003. Automatic interpretation of magnetic data based on Euler deconvolution with unprescribed structural index. *Computers & Geosciences* 29, 949-960.
- Dardenne, M.A. 2000. The Brasília Fold Belt. In: U.G. Cordani, E.J. Milani, A. Thomaz Filho & D.A. Campos. 2000. Tectonic evolution of South America. 31 st International Geological Congress, Rio de Janeiro, SBG. p. 231-263.
- Dawson, J.B. 2008. The Gregory Rift Valley and Neogene–Recent Volcanoes of Northern Tanzania. *Geological Society, London, Memoirs*, 33, 13–20. The Geological Society of London 2008. DOI: 10.1144/M33.4
- Delcamp, A.; Kervyn, M.; Benbakkar, M.; Kwelwa, S.; Peter, D. 2016. Large volcanic landslide and debris avalanche deposit at Meru, Tanzania. *Landslides* (2016). doi:10.1007/s10346-016-0757-8.
- Dhaoui, M.; Gabtni, H.; Jallouli, C.; Jleilia, A.; Mickus, K.L.; & Turki, M.M. 2014. Gravity analysis of the Precambrian basement topography associated with the northern boundary of Ghadames Basin (southern Tunisia). *Journal of Applied Geophysics*, 111, 299-311.
- Ebinger, C.; Poudjom Djomani, Y.; Mbede, E.; Foster, A.; Dawson, J. B. 1997. Rifting Archaean lithosphere: the Eyasi– Manyara – Natron rifts, East Africa. *Journal of the Geological Society, London*, 154, 947–960.
- Fadel, I.; van der Meijde, M.; Kerle, N.; Lauritsen, N. 2015. 3D object-oriented image analysis in 3D geophysical modelling: Analysing the central part of the East African Rift System, *International Journal of Applied Earth Observation and Geoinformation*, Volume 35, 2015, Pages 44-53, ISSN 0303-2434, <http://dx.doi.org/10.1016/j.jag.2013.11.004>.
- Fairhead, J.D.; Bennett, K.J.; Gordon, D.R.H.; Huang, D. 1994. Euler: beyond the “black box”. In 64th

- Annual International Meeting. Los Angeles, CA: Society of Exploration Geophysicists, Expanded Abstracts, pp. 422–424.
- Fairhead, J.D. 1976. The structure of the lithosphere beneath the eastern rift, East Africa, deduced from gravity studies. *Tectonophysics*, 30, 269–298.
- Fedi, M., Quata, T., 1998. Wavelet Analysis for the regional-residual and local separation at potential field anomalies. *Geophysical Prospecting*, 46, 507-525.
- Fleitout, L.; Froidevaux, C.; Yuen, D. 1986. Active lithospheric thinning. *Tectonophysics*, 132(1-3), 271-278.
- Fontijn, K.; Delvaux, D.; Ernst, G.G.J.; Kervyn, M.; Mbede, E.; Jacobs, P. 2010. Tectonic control over active volcanism at a range of scales: Case of the Rungwe Volcanic Province, SW Tanzania; and hazard implications, *Journal of African Earth Sciences*, Volume 58, Issue 5, Pages 764-777, ISSN 1464-343X, <http://dx.doi.org/10.1016/j.jafrearsci.2009.11.011>.
- Geosoft 1995. Geosoft mapping and processing system. Toronto, Geosoft Inc., 50p.
- Geosoft 2001. Geosoft technical note: microlevelling using bi-directional gridding. In: www.geosoft.com/news.
- Gout, R.E.; Khattach, D.; Houari, M.R.; Kaufmann, O.; Aqil, H. 2010. Main structural lineaments of north-eastern Morocco derived from gravity and aeromagnetic data. *Journal of African Earth Sciences*, 58(2), 255-271.
- Guadagnin, F.; Chemale, F. 2015. Detrital zircon record of the Paleoproterozoic to Mesoproterozoic cratonic basins in the São Francisco Craton, *Journal of South American Earth Sciences*, Volume 60, July 2015, Pages 104-116.
- Guryanov, V.A.; Peskov, A.Y. 2017. Ulkan Paleorift Structure in the South-Eastern Environs of the Siberian Platform: Age, Conditions, Sources, and Geodynamic Setting. *Geosciences Research*, Volume 2; doi:10.22606/gr.2017.22001
- Haralick, R.M.; Shapiro, L.G. 1991. Glossary of computer vision terms. *Pattern recognition*, 24(1), 69-93.
- Jordan, T.A.; Ferraccioli, F.; Ross, N.; Corr, H.F.; Leat, P.T.; Bingham, R.G.; Rippin, D.M.; le Brocq, A.; Siegert, M.J., 2013. Inland extent of the Weddell Sea Rift imaged by new aerogeophysical data. *Tectonophysics*, 585, pp.137-160.
- Kearey, P.; Klepeis, K.A.; Vine, F.J. 2009. *Global Tectonics*, Third edition, Wiley-Blackwell Publication.
- Keating, P.B. 1998. Weighted Euler deconvolution of gravity data. *Geophysics*, 63(5), 1595-1603.
- Keller, G.R.; Lidiak, E.G.; Hinze, W. J.; Braile, L.W. 1983. The role of rifting in the tectonic development of the Midcontinent, U.S.A., *Tectonophysics*, v. 94, p. 391-412.
- Keller, R.G. 2002. *Getting Started with Gravity and Magnetic Techniques. A Tutorial for Using and Understanding Gravity Data*. Department of Geological Sciences Pan American Centre for Earth and Environmental Studies. University of Texas at El Paso.
- Keckler, D. 1994. SURFER for Windows User's Guide. Golden Software Inc.: Golden, Colorado.

- Keen, C.E. 1985. The dynamics of rifting: deformation of the lithosphere by active and passive driving forces. *Geophysical Journal International*, 80(1), 95-120.
- Khalil, M. A.; Santos, F. M.; Farzamian, M. 2014. 3D gravity inversion and Euler deconvolution to delineate the hydro-tectonic regime in El-Arish area, northern Sinai Peninsula. *Journal of Applied Geophysics*, 103, 104-113.
- Koptev, A.; Burov, E.; Calais, E.; Leroy, S.; Gerya, T.; Guillou-Frottier, L.; Cloetingh, S. 2016. Contrasted continental rifting via plume-craton interaction: applications to Central East African rift. *Geoscience Frontiers*, 7(2), pp.221-236.
- Lenharo, S.L.; Moura, M.A.; Botelho, N.F. 2002. Petrogenetic and mineralization processes in Paleo-to Mesoproterozoic rapakivi granites: examples from Pitinga and Goiás, Brazil. *Precambrian Research*, 119(1), 277-299.
- Liu, F.; Wang, Y.; Liang, W. 2016. Field source characteristic of gravity variation in Hexi region before Menyuan Ms6. 4 earthquake based on the Euler deconvolution. *Geodesy and Geodynamics*, 7(5), 317-322.
- Lu, S. N., Yang, C. L., Li, H. K., and Li, H. M. 2002. A group of rifting events in the terminal Paleoproterozoic in the North China Craton: *Gondwana Research*, v.5, n.1, p.123–131, [http://dx.doi.org/10.1016/S1342-937X\(05\)70896-0](http://dx.doi.org/10.1016/S1342-937X(05)70896-0).
- McKenzie, D. 1978. Some remarks on the development of sedimentary basins. *Earth and Planetary science letters*, 40(1), 25-32.
- Maguire, P.K.H.; Swain, C.J.; Masotti, R.; Khan, M.A. 1994. A crustal and uppermost mantle cross-sectional model of the Kenya Rift derived from seismic and gravity data. *Tectonophysics*, 236(1-4), 217-249.
- Martins-Ferreira, M.A.; Campos, J.E.G.; Alvarenga, C.J.S. 2013. The Jequitaiá Formation at the Boa Vista Ridge, town of Vila Boa, GO: example of sedimentation by Neoproterozoic terminal glaciers. *Brazilian Journal of Geology* 43 (2), 373–384.
- Martins-Ferreira, M.A.; Chemale, F.; Dias, A.N.C.; Campos, J.E.G. 2017. Proterozoic intracontinental basin succession in the São Francisco Craton western margin constrained by detrital zircon geochronology. *Journal of South American Earth Sciences*. *Submitted for Publication on the Journal of South American Earth Sciences*.
- Matteini, M.; Dantas, E.L.; Pimentel, M.M.; Alvarenga, C.J.S.; Dardenne, M.A. 2012. U-Pb and Hf isotope study on detrital zircons from the Paranoá Group, Brasília Belt Brazil: constraints on depositional age at Mesoproterozoic-Neoproterozoic transition and tectono-magmatic events in the São Francisco Craton. *Precambrian Res.* 206-207, 168-181.
- Miller, H.G.; Singh, V. 1994. Potential field tilt - a new concept for location of potential field sources. *Journal of Applied Geophysics* 32, 213-217.
- Mushayandebvu, M.F.; Lesur, V.; Reid, A.B.; and Fairhead, J.D. 2003. Grid Euler deconvolution with

- constraints for two-dimensional structures. *Geophysics*, 69: 489–496.
- Nyblade, A., and Pollack, H., 1992, A gravity model for the lithosphere in western Kenya and northeastern Tanzania: *Tectonophysics*, v. 212, p. 257–267.
- Pedrosa-Soares, A.C.; Alkmim, F.F. (2011). How many rifting events preceded the development of the Araçuai-West Congo orogen? *Revista Geonomos*, 19(2).
- Pimentel, M.M.; Heaman, L.; Fuck, R.A.; Marini, O.J. 1991. U-Pb zircon geochronology of Precambrian tin-bearing continental-type acid magmatism in central Brazil. *Precambrian Research*, 52(3), 321-335.
- Pimentel, M.M.; Rodrigues, J.B.; DellaGiustina, M.E.S.; Junges, S.; Matteini, M.; Armstrong, R. 2011. The tectonic evolution of the Neoproterozoic Brasília Belt, central Brazil, based on SHRIMP and LA-ICPMS U–Pb sedimentary provenance data: a review. *Journal of South American Earth Sciences*, 31(4), 345-357.
- Pimentel, M.M. 2016. The tectonic evolution of the Neoproterozoic Brasília Belt, central Brazil: a geochronological and isotopic approach. *Brazilian Journal of Geology*, 46, 67-82.
- Ramberg, I.B. and Morgan, P. 1984. Physical characteristics and evolutionary trends of continental rifts: Proceedings, International Geological Congress, 27th, Moscow: Utrecht, Netherlands. VNU Science Press, v.7, p. 165-216.
- Rämö, O.T. 1991. Petrogenesis of the Proterozoic rapakivi granites and related basic rocks of southeastern Fennoscandia: Nd and Pb isotopic and general geochemical constraints. *Geol Surv Finland Bull* 355, 161 pp
- Rämö, O.T.; Haapala, I. 1995. One hundred years of Rapakivi Granite, *Mineral. Petrol*, vol. 52, pp. 129–185.
- Reid, A.B.; Allsop, J.M.; Granser, H.; Millett, A.J.; and Somerton, I.W.; 1990. Magnetic interpretation in three dimensions using Euler deconvolution. *Geophysics*, 55: 80–91.
- Reis, H. L. and Alkmim, F. F. 2015. Anatomy of a basin-controlled foreland fold-thrust belt curve: The Três Marias salient, São Francisco basin, Brazil. *Marine and Petroleum Geology*, 66, 711-731.
- Reis, H.L.S.; Alkmim, F.F.; Fonseca, R.C.S.; Nascimento, T.C.; Suss, J.F.; Prevatti, L.D. 2017. The São Francisco Basin. *In: São Francisco Craton, Eastern Brazil. Tectonic Genealogy of a Miniature Continent*. Pgs 205-220. Editors: Heilbron, Monica, Cordani, Umberto G., Flecha de Alkmim, Fernando. *Regional Geology Reviews*. Springer, 1ed. DOI:10.1007/978-3-319-01715-0.
- Rodrigues J.B. 2008. Proveniência de sedimentos dos grupos Canastra, Ibiá, Vazante e Bambuí - Um estudo de zircões detríticos e Idades Modelo Sm-Nd. Doctorate thesis, Universidade de Brasília.
- Roest, W. R., Verhoef, J., and Pilkington, M., 1992. Magnetic interpretation using the 3-D analytic signal. *Geophysics* 57(1), 116-125.
- Roy, M.; MacCarthy, J.K.; Selverstone, J. 2005. Upper mantle structure beneath the eastern Colorado plateau and Rio Grande rift revealed by Bouguer gravity, seismic velocities, and xenolith data,

- Geochem. Geophys. Geosyst. 6, Q10007, doi: 10.1029/2005GC001008.
- Roy, I.G., 2013. Tilt angle interpretation of dipping fault model. *Journal of Applied Geophysics* 98, 33-43.
- Salem, A., and Ravat, D., 2003. A combined analytic signal and Euler method (AN-EUL) for automatic interpretation of magnetic data. *Geophysics*, 68(6): 1952-1961.
- Salem, A.; Williams, S.; Fairhead, D.; Smith, R.; Ravat, D. 2008. Interpretation of magnetic data using tilt-angle derivatives. *Geophysics* 73, L1-L10.
- Sengör, A.M.C., Natal'in, B.A., 2001. In: Ernst, R.E., Buchana, K.L. (Eds.), *Mantle Plumes: Their Identification Through Time*. Colorado, Geological Society of America Special Paper, vol. 352, pp. 389–482.
- Sibson, R. H. 1977. Fault Rocks and Fault Mechanisms. *Journal of the Geological Society* 133, no. 3 (1977): 191-213.
- Soares, J. E., Berrocal, J., Fuck, R. A., Mooney, W. D., & Ventura, D. B. (2006). Seismic characteristics of central Brazil crust and upper mantle: a deep seismic refraction study. *Journal of Geophysical Research: Solid Earth*, 111(B12).
- Spector, A. and Grant, F.S. 1970. Statistical models for interpreting aeromagnetic data. *Geophysics*, 35(2), 293-302.
- Tanizaki, M.L.N.; Campos, J.E.G.; Dardenne, M.A. 2015. Stratigraphy of the Araí Group: record of Paleoproterozoic rifting in Central Brazil. *Brazilian Journal of Geology*, 45(1), 95-108.
- Thompson, D.T., 1982. EULDPH—a new technique for making computer-assisted depth estimates from magnetic data. *Geophysics*, 47: 31–37.
- Toriwaki, J. and Yonekura, T. 2002. Euler Number and Connectivity Indexes of a Three Dimensional Digital Picture. *Forma*, 17, 183-209.
- Trindade, C.R.; Soares, J.E.; Fuck, R.A.; Carmelo, A.C.; Peixoto, C.L.O. 2014. Estrutura Crustal do Brasil Central. *Laboratório Nacional de Energia e Geologia*. Porto, Portugal. *Comunicações Geológicas, Especial I*, 101, 339-343.
- Turcotte, D. L.; Emerman, S. H. 1983. Mechanisms of active and passive rifting. *Tectonophysics*, 94(1-4), 39-50.
- Watts, A.B. 2001. *Isostasy and flexure of the lithosphere*, Cambridge University Press.
- Yirgu, G.; Ebinger, C.J.; Maguire, P.K.H.; (Eds.) 2006. The structure and evolution of the East African Rift System in the Afar Volcanic Province, *Geol. Soc. Spec. Publ.*, 259, 1–327.
- Yu, J.H.; Fu, H.Q.; Zhang, F.; Guan, M. 1991. Geochemistry of a Rapakivi Granite Suite in the Proterozoic Rift Trough in Beijing and its Vicinity, *Acta Geol Sinica*, vol. 4, no. 2, pp. 169–186, 1991.
- Yu, Y., Liu, K.H., Reed, C.A., Moidaki, M., Mickus, K., Atekwana, E.A., Gao, S.S. 2015, A joint receiver function and gravity study of crustal structure beneath the incipient Okavango Rift,

Botswana: *Geophysical Research Letters*, v. 42, p. 8398–8405, doi:10.1002/2015GL065811.

Zhou, D., Wang, W. Y., Wang, J. L., Pang, X., Cai, D. S., and Sun, Z., 2006. Mesozoic subduction-accretion zone in northeastern South China Sea inferred from geophysical interpretations. *Science in China: Science in China Series D: earth Sciences* 49(5), 471-482.

CAPÍTULO 5

Artigo 4:

**Paranoá basin evolution constrained by stratigraphic and gravimetric data
and its correlation to regional basins**

Proterozoic Paranoá basin evolution constrained by stratigraphic and gravimetric data and correlation to regional sequences in the São Francisco Craton

Marco Antonio Caçador Martins-Ferreira¹ marcoc@unb.br (Corresponding Author); José Eloi Guimarães Campos¹; Monica Giannoccaro Von Huelsen^{1,2}

3. Instituto de Geociências, Universidade de Brasília, Campus Univ. Darcy Ribeiro, Asa Norte, 70910900, Brasília, DF, Brazil.

4. Observatório Sismológico. Universidade de Brasília. Campus Univ. Darcy Ribeiro, Asa Norte, 70910900, Brasília, DF, Brazil.

Abstract

Gravimetric and stratigraphic data were used to constraint the mechanisms responsible for the installation of the Stenian Paranoá Basin, not yet studied in detail. Gravimetric modeling results confirmed the existence of paleorift structures beneath the Paranoá sequence at the study area. Both stratigraphic and gravity results show evidences of mechanic subsidence by faulting during basin installation generating topographic unevenness of approximately 700 m. Basin installation faults coincide with deeper paleorift faults and are thus interpreted as reactivations. Despite faulting, the exposed sedimentary record that follows basal conglomerates does not contain evidences indicating strict rift environment. Besides, the great lateral continuity and relatively constant thickness of facies indicate that initial mechanic subsidence rapidly gave place to flexural subsidence at the following stages of basin evolution. Furthermore, a review of basin types and nomenclature is presented in order to compare worldwide cases with the Paranoá Group. Main basin characteristics indicate that the Paranoá Group was deposited as different basin types, intracratonic sag, continental platform and cratonic margin, according to the tectonic locality and basement characteristics and was possibly connected to a passive margin basin at times of sea level rise, or evolved to a passive margin later in time, most probably related to the Canastra and Vazante groups, Lower Araxá Group and Lower Ibiá Group (Cubatão Formation). Finally, regional and global correlations are proposed. Regional correlation to other São Francisco craton sequences is established based on a comprehensive geochronologic and stratigraphic data compilation. The Paranoá Group is correlated to the intracontinental sequences of the Upper Espinhaço first-order sequence and the Canastra Group is interpreted as a younger unit, suggested to be transitional or correlated to the passive margin sequences of the Macaúbas first-order sequence. Temporal and spatial correlation of these sequences with glaciation and rifting events suggest that the passive margin was possibly formed by rifting during the ca. 1.0-0.8 Ga. interval, known as Tonian Taphrogenesis, and affected by an early Sturtian glaciation. An attempt to establish a connection of rift-related magmatism and basin dynamics to global events has shown that the main extensional and compressional events in the craton can be correlated to the assembly and break-up of Columbia, Rodinia and Gondwana, the data and analysis presented in this paper might specifically contribute for the puzzling participation of São Francisco-Congo craton in Rodinia.

Keywords: Paranoá Group, São Francisco Craton, Rodinia, cratonic margin basin, passive margin, gravimetric modeling

1. Introduction

Sedimentary basins may form as one type of basin and evolve through time into another type of basin. Cratonic basins may have their origin as rift basins. Foreland basins may have much earlier phases as rift basins that evolve to passive margin basins and so on (Coleman and Cahan 2012).

During the Proterozoic Eon, the São Francisco-Congo craton hosted a series of poly-historic basins filled by a series of unconformity-bounded units that record distinct subsidence pulses, distributed over relatively long time periods of time (Uhlein et al. 1995, Chemale et al. 2012, Alkmim et al. 2012). These pulses, despite isolated in time, seem to be connected to each other, either by the result of long term isostatic reflexes and/or by reactivation of previous extensional structures during new extensional events.

Supercontinent reconstruction studies use several geological, biological and physical parameters to constraint the position of cratons in each time period. Initially, the reconstructions were primarily based on the linkage of truncated orogenic belts of same age (Dalziel 1991, Hoffman 1991, Moores 1991). Recent works use practically all available data, being paleomagnetic data and cratonic border tectonic framework the most important parameters, where rift-related magmatism, basin environment and sedimentary provenance play a key role (Cawood et al. 2007, Li et al, 2007, Evans et al. 2016, Cawood and Pisarevsky 2017). The types of basins preserved in cratonic margins are often used to define the general tectonic environment for those margins during each consecutive basin development, thus the importance of identifying more precisely the tectonic framework of these basins through time and space. In most reconstructions of Rodinia, the São Francisco-Congo Craton is one of the hardest to position, specially due to the lack of paleomagnetic data and poor chronostratigraphic surveys (Bogdanova et al. 2009).

During early Statherian, the São Francisco Craton western margin witnessed an extensional event that opened a volcanic rift basin, represented by the Araí Group (Tanizaki et al. 2015). By late Calymmian the rift basin was covered unconformably by a broader, sag-type basin, represented by the Traíras Group (Martins-Ferreira et al. 2017). In the Stenian period, an even wider basin was developed over these sequences and its sedimentary record is represented by the Paranoá Group (Campos et al. 2013) a first-order sequence widely exposed in the cratonic margin and main subject of this study.

There is no agreement on an accurate definition regarding the type of basin or geotectonic environment for the Paranoá Group sequence. In regional studies, most of the authors cite the

group as related to passive margin environment in a broader sense (Matteini 2012, Campos et al. 2013, Pimentel 2016, Fuck et al. 2017). However, studies of greater detail do not come to characterize it as such, preferring the terms epicontinental basin (Faria 1995, Dardenne 2000), intracratonic basin (Chemale et al. 2012, Alvarenga et al. 2012) or rift-sag basin (Reis and Alkmim 2015). Another open question regards whether the Canastra Group is equivalent to the Paranoá or represents a younger sequence, deposited in a different environmental setting.

Fuck et al. (2008) state that, available geological data indicate that the western border of the São Francisco Craton was a passive margin, and that basin was most likely developed in the Mesoproterozoic. They add, however, that the opening of such basin is not yet well understood, but it must have had its closure episode around 900 Ma. Since that statement, no further studies were conducted to specifically investigate the Paranoá basin opening and further evolution.

In a recently published book about the São Francisco Craton, the lack of agreement on the Paranoá sequence basin type is evident. Heilbron et al. (2017), on chapter 1 defines the Paranoá Group as a rift-sag sequence, and the Canastra Group as rift to passive margin. Reis et al. (2017), on chapter 7 states that the Paranoá-Upper Espinhaço sequence can be subdivided into two 2nd-order sequences: a lower rift and an upper sag/passive margin. Fuck et al. (2017) on chapter 11 interprets the Canastra and Paranoá groups as equivalent and part of a passive margin sequence.

Recently, detailed geological mapping has uncovered tectono-stratigraphic evidences of pre- to syn-basin installation tectonic activity. These evidences, reported in this paper, have provided unprecedented information about the Paranoá basin history. Further investigation has revealed new aspects of basin evolution that may aid to better constrain the tectonic scenario of the Ectasian-Stenian extensional pulse that occurred throughout the São Francisco paleocontinent.

In the Neoproterozoic, all the mentioned Paleoproterozoic successor sedimentary sequences were affected by the Brasiliano orogenic event that imposed compressional deformation (Dardenne 2000, Fuck et al. 2004, Valeriano et al. 2004a, Alkmim and Martins-Neto 2012, Uhlein et al. 2012), which represents a considerable obstacle for Proterozoic tectono-stratigraphic studies.

Since borehole data is scarce and not representative, potential field studies have been used to aid structural and stratigraphic interpretation by identifying main structural features that affect the crust at different depths. International literature is abundant with examples of study cases that use potential data, especially gravimetric, to aid geological interpretation. (Gerard and

Griveau 1972, Boyce & Morris 2002, Engen et al. 2006, Salem et al. 2013, Dossing et al. 2014, Scisciani et al. 2014, Abtout et al. 2015, Chakravarthi & Ramamma 2015, Pallero et al. 2015).

Seeking out to contribute to the knowledge gap regarding the Paranoá sequence basin type and evolution, this study makes use of gravimetric data modeling and inversion, structural mapping and stratigraphic analysis in an attempt to characterize the basin in more detail by investigating the geometry and processes that controlled basin installation and further development.

2. Geological Setting

The Paranoá Group crops out in the Brasília Belt, an orogenic belt formed during the Neoproterozoic Brasiliano collage (Dardenne 2000; Valeriano et al. 2008; Brito-Neves et al. 2014; Pimentel 2016) in the western margin of the São Francisco Craton (Figure 1). The Brasília Belt external zone, where the study is focused, is a fold-and-thrust belt where rocks were locally submitted to metamorphism as high as low-greenschist facies, but are generally anchimetamorphic.

Despite not much affected by metamorphism, the preserved sedimentary sequences are folded and faulted in different styles and intensities across the belt. Isoclinal to open folds usually have undulated axis and are occasionally cut by transpressional shear zones generally parallel or diagonal to the fold axes.

In addition to the Paranoá basin, the São Francisco Craton western margin hosted a series of sedimentary basins during the Proterozoic. In the Northern Brasília Belt, Statherian to Ediacaran cratonic sequences have developed one over the other and are well exposed over hundreds of square kilometers. The Siderian/Rhyacian crystalline basement is also well exposed over a large area, allowing the mapping of its main structures and lithological heterogeneities. These main geological elements of the Northern Brasília Belt are briefly described below.

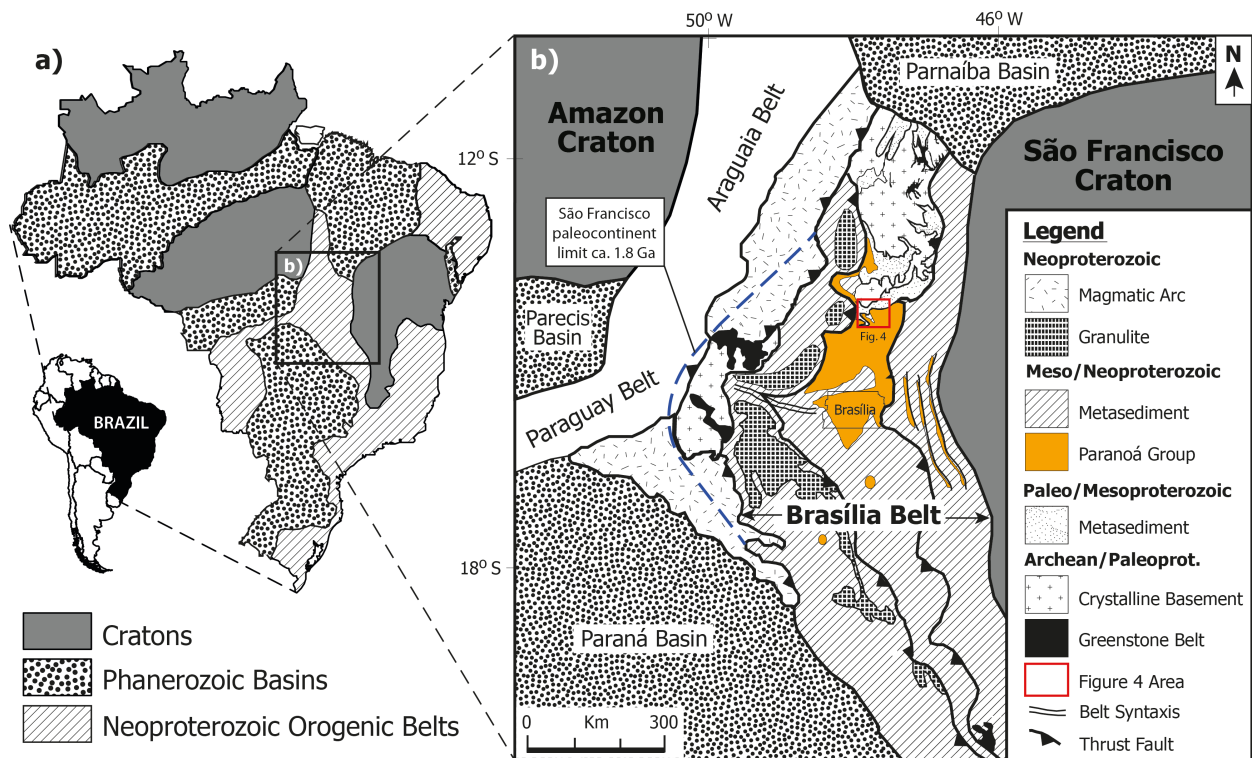


Figure 1 –Simplified geologic map showing a) the location of the Brasília Belt in central Brazil relative to main cratons and basins and b) Paranoá Group outcropping area in the São Francisco paleocontinent western margin.

2.1. Siderian/Rhyacian Crystalline Basement

The São Francisco Craton western margin crystalline basement consists of a collage of continental and arc-related terranes which were assembled and welded together during a sequence of orogenic episodes spanning Siderian and Rhyacian times that generated intense peraluminous and metaluminous magmatism (Cruz et al. 2003; Cuadros et al. 2017a).

2.2. Statherian Basin

The Araí Group represents a Statherian rift sequence and was divided in the Água Morna and Arraias formations, representing the pre-rift and syn-rift sequences, respectively (Martins-Ferreira et al. 2017). The post-rift sequence is absent either because it was eroded or not deposited. Bimodal volcanism occurs within both formations and was dated at ca. 1771 Ga (Pimentel et al. 1991). The Pirapora Aulacogen is believed to be of this age, possibly as part of the Araí-Espinhaço rift paleorift system Alkmim et al. (2012).

2.3. Calymmian Basin

The Traíras Group represents a sag-type basin sequence and its maximum depositional age was recently dated ca. 1542±31 Ma. by zircon provenance study (Martins-Ferreira et al. 2017).

2.4. Stenian/Tonian Basins

In the late Stenian period the Paranoá basin was installed over the Traíras Group sequences. The mechanisms for basin installation are not yet studied in detail. There is still debate whether the Canastra Group is a distal equivalent to the Paranoá Group or represents a younger sequence.

Youngest zircons found in the Canastra Group are dated ca. 1011 Ma. (Valeriano et al. 2004a,b) and ca. 1030 Ma. (Rodrigues 2008), suggesting Tonian age for this unit, thus younger than the Paranoá Group, where the youngest zircons ever found were dated ca. 1540 Ma. (Matteini et al. 2012), ca. 1428±12 Ma. (Martins Ferreira et al. 2017), and xenotime diagenetic age was dated ca. 1042±22 Ma (Matteini et al. 2012).

Tonian anorogenic magmatism ca. 875±9 Ma. (Silva et al. 2008) and rifting with maximum depositional age ca. 989±3 (Alcântara et al. 2017) Ma events are registered in the São Francisco craton NW and NE margins, but they possibly postdate the Paranoá sequence, according to the xenotime diagenetic age found by Matteini et al. (2012).

2.5. Ediacaran Basins

During the Ediacaran, the Brasiliano orogeny was taking place, the Araxá, Bambuí and Ibiá groups were interpreted as deposited, at least partially, in foreland and active margin environments that lasted until the Cambrian period (Dardenne 2000). The Bambuí Group covers unconformably the Siderian/Rhyacian crystalline basement, the Statherian Araí Group, the Callymmian Traíras Group and the Stenian Paranoá Group (Dardenne 2000). That widespread erosive unconformity was most likely generated after the Brasiliano orogeny uplift. The Bambuí group sequence begins with glacial deposits of debated age. The Macaúbas Group in the Araçuaí Belt and the Ibiá Group basal Cubatão Formation also contain glacial records, but are attributed to an earlier glacial event (Sturtian Glaciation), (Heilbron et al. 2017 and references therein).

2.6. Paranoá Group Geology

In its extensive area of occurrence, the Paranoá Group consists essentially of quartzite, phyllite, shale and marble. The metamorphic grade varies from anchimetamorphic in the Brasília Belt External Zone to greenschist facies near and at the Internal Zone, with few exceptions. In

the study area it occurs over the Araí Group and below the Bambuí Group, separated from both by an erosive unconformity marked by conglomerates.

The stratigraphy of the Paranoá Group presents some lateral and vertical variations that occur predominantly from east to west. However, Faria (1995) proposed an integrated stratigraphy that can be regionally correlated in the outermost portion of the Brasília Belt. The lithostratigraphic units were informally referred to by code letters from base to top: SM, R1, Q1, R2, Q2, S, A, R3, Q3, R4 and PC. The interpretation of the depositional paleo-environments was proposed by Faria (1995) and the formalization of these units into formations was proposed by Campos et al. (2013)(Figure 2).

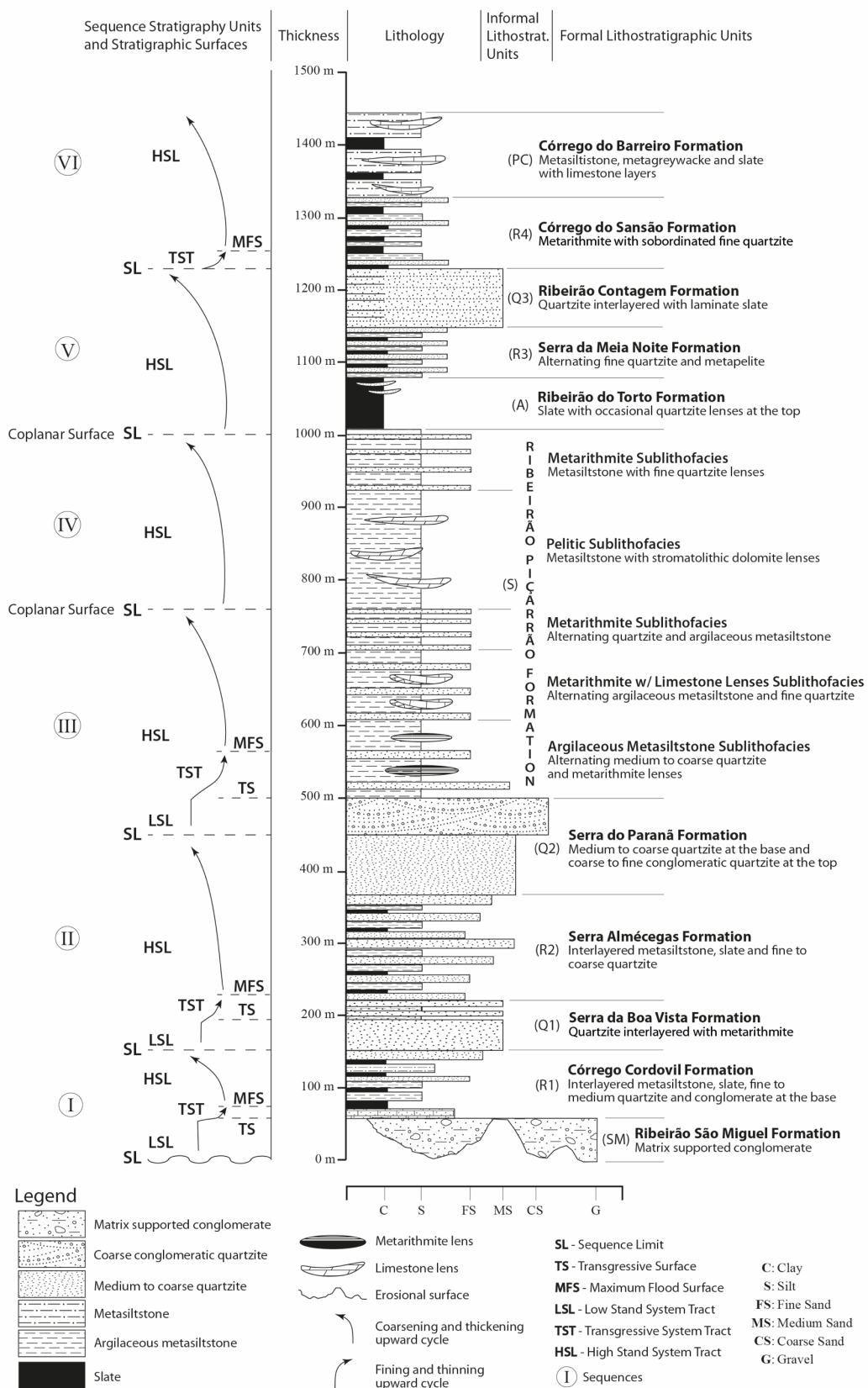


Figure 2 – Paranoá Group litho-stratigraphic column at its type locality (Alto Paraíso de Goiás / Distrito Federal) showing the related sequence stratigraphy cycles (modified from Campos et al. 2013).

The main features of each Paranoá Group formation are described as follows:

Ribeirão São Miguel Formation: (SM unit) matrix supported conglomerate, characterized by coarse sand with clay matrix and carbonatic cement, containing sub-angular to angular clasts, mainly pebbles, locally blocks and occasionally boulders of quartzite, metasilstone, quartz vein and marble. Fresh rock greenish-grey color changes to pinkish and reddish hues when weathered. This unit, although presenting a great regional spread, is not continuous and maximum thickness of 70 meters has been mapped.

Córrego Cordovil Formation: (R₁ unit) presents concordant and gradual contact with the basal conglomerate and is composed of a succession of impure carbonatic and clay-silt horizons evolving to rhythmic succession of thin layers of fine to medium feldspathic quartzite and silt-clay. Salt pseudomorphs and shrinking cracks are frequently observed close to the base. Mean thickness is 70 meters.

Serra da Boa Vista Formation: (Q₁ unit) fine to medium white quartzite are well stratified in decimetric banks with rare silt-clay layers at the top and frequent tabular cross-stratification. Total thickness reaches 80 meters.

Serra Almécegas Formation: (R₂ unit) represents an abrupt transition from the quartzite from the formation below. Centimeter thick layers of fine quartzite interlayered with siltstone and millimetric thick white clay. Cross-stratification truncated by waves and asymmetric wave-ripple marks are common in the thicker quartzite layers. Quick sand and fluidization structures are also common. The package is generally 150 meters thick.

Serra do Paranã Formation: (Q₂ unit) decimetric to metric layers of medium to coarse yellowish quartzite. Conglomeratic channels and beds with sub-angular to round pebbles are common at the top. Planar cross-stratification is common. Herringbone and overturned cross-stratification are common at the conglomeratic levels. Total thickness reaches 150 meters.

Ribeirão Piçarrão Formation: (S unit) is the most lithologic variable sequence, reaching mean thickness of up to 500 m. This formation can be subdivided in five lithofacies. Is subdivided in five lithofacies that comprise a sequence of greenish homogeneous metasilstone interbedded with sandy layers forming a metarhythmite, limestone and rare dolomite lenses locally occur. Its base is in general formed by pelitic/sandy metarhythmite. The lower sequence presents greater clay contribution and much less sandy contribution.

Ribeirão do Torto Formation: (A unit) presents transitional contact with the previous unit and is formed exclusively by homogeneous greenish-grey slate that becomes reddish when

weathered. Occasional sand layers with cross-bedding structures occur. Unit thickness is difficult to estimate precisely due to folding, but is interpreted around 70-80 meters.

Serra da Meia Noite Formation: (R₃ unit) sandy metarhythmite, characterized by decimetric to metric interbedding of quartzite and metapelite (silt and clay). Locally quartzite layers up to 10 meters are observed. Total thickness can reach 90 meters.

Ribeirão Contagem Formation: (Q₃ unit) composed of pure silicified fine white quartzite rich in fishbone, tabular and ribbed cross-bedding and asymmetric wave ripple structures. Maximum thickness reaches 100 meters.

Córrego do Sansão Formation: (R₄ unit) argillaceous metarhythmite composed of interbedded silty and clayish material besides thin layers of reddish quartzite. Sandy layers present laminations truncated by waves, cross lamination and hummocky structures. Thickness varies between 100 and 150 meters.

Córrego do Barreiro Formation: (PC unit) dominantly pelitic with grey shale and argillaceous metasiltstone associated to fine marble lenses that might contain preserved algal structures such as conic and columnar stromatolites. Decimetric to metric layers of medium to coarse dark quartzite occur. Thickness varies between 120 and 150 meters.

The Paranoá Group relative age was established by its lower and upper unconformities, regional correlations to other similar sequences and also by the stromatolites that occur in its carbonate rocks. Columnar and conophyton stromatolites indicate sedimentation age between 950 and 1150 Ma, positioning the unit in the Meso-Neoproterozoic boundary (Cloud & Dardenne 1973, Dardenne 1979). U-Pb results on detrital zircons indicate maximum deposition age at ca. 1540±10 Ma and probable deposition age around ca. 1042±22 Ma, defined by U-Pb age of xenotime diagenetic overgrowth in zircon (Matteini et al. 2012).

The three intracontinental extensional pulses that generated the Araí, Traíras and Paranoá sequences, were comprised by the Veadeiros Supergroup (Martins-Ferreira et al. 2017) representing respectively the lower, middle and upper portions of the supergroup and directly correlated to the lower, middle and upper Espinhaço Supergroup (Chemale et al. 2012) (Figure 3).

The Paranoá Group has been correlated to the Upper Espinhaço (Chemale et al. 2012; Guadagnin and Chemale 2015) or Espinhaço II sequence (Reis and Alkmim 2015), more specifically to the Caboclo and Morro do Chapéu Formation in the Chapada Diamantina region and Riacho do Bento, Mosquito and Fazendinha formations in the Northern Espinhaço and also to the Conselheiro Mata Group in the São Francisco Craton western margin (Heilbron et al.

2017). However, the correlation to the Caboclo Formation is still fragile, due to poor age constraints (Pb-Pb in carbonatic rocks; Babinski et al. 1993) and is considered premature by the authors since more geochronological data would be necessary to better establish that correlation. Nonetheless, the braided river and conglomeratic deposits of the Caboclo Formation could be related to the Stenian rifting recorded by the Sopa Brumadinho Formation.

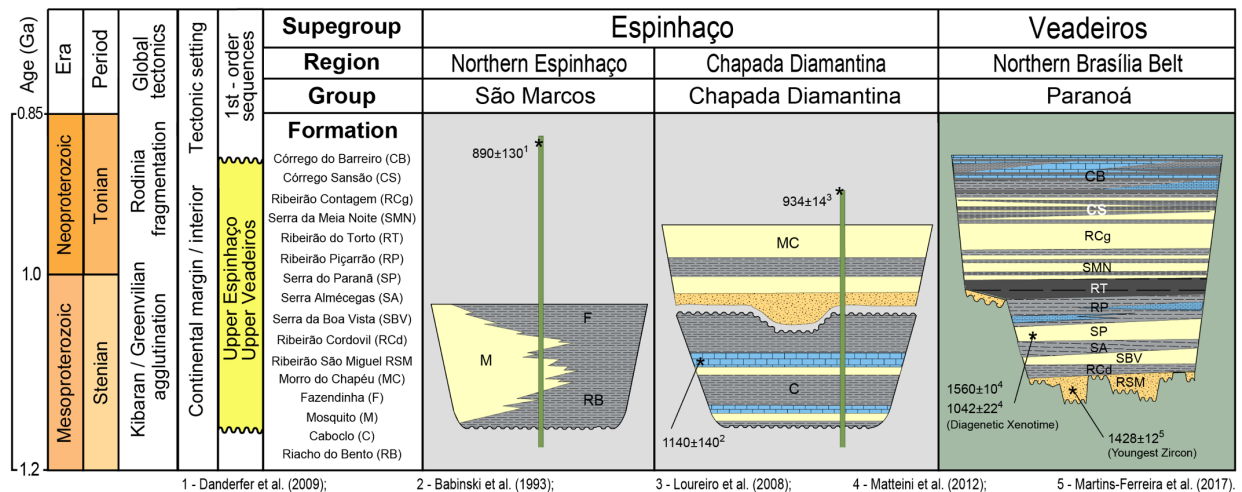


Figure 3 – Regional correlation of the Paranoá Group with other sequences in the São Francisco paleocontinent (modified from Martins-Ferreira et al. 2017).

Geological mapping in the Chapada dos Veadeiros region confirmed the existence of a regional scale fault, that has affected the Paranoá Group sequence at the time of sedimentation. The previously unnamed fault was already identified in previous maps, but no mention to it has been found in previous literature. The NNW-SSE trending structure was named Serra do Cristal Fault (SC Fault), once it hosts several quartz crystal mines. This fault's importance was previously ignored, and its detailed mapping revealed important aspects regarding the Paranoá basin evolution during the Ectasian/Stenian extensional pulse. The SC Fault separates two distinct deformational, stratigraphic and geomorphologic compartments (Figure 4).

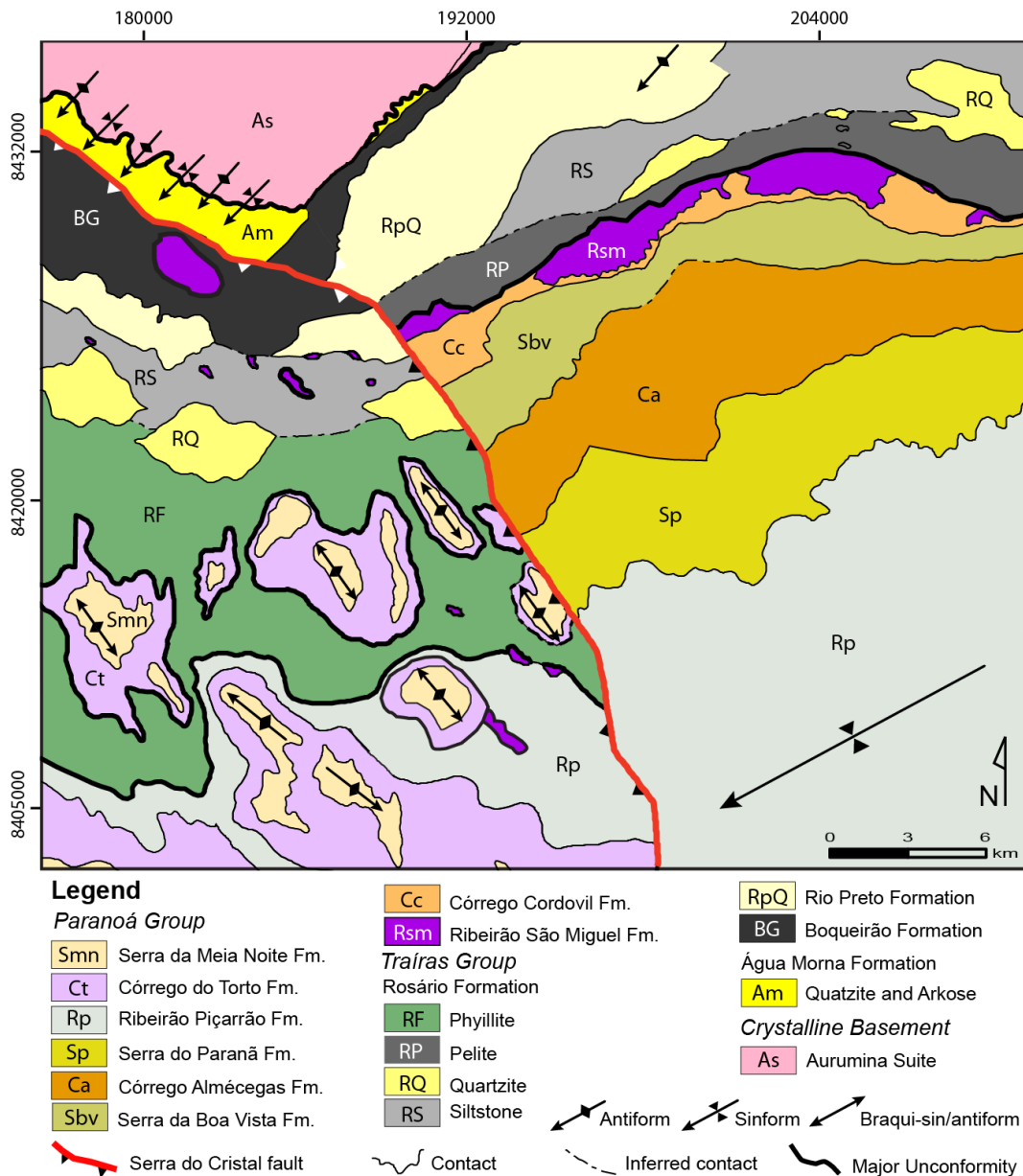


Figure 4 – Local geologic map displaying the Serra do Cristal Fault acting as a boundary between stratigraphic and deformational compartments.

To the east of the SC Fault, the Paranoá Group strata are smoothly folded, forming kilometer scale gentle folds with SW plunging axes while to the west of the fault the strata are intensely folded and faulted forming long kilometeric series of close to tight folds, sometimes recumbent with axes parallel to the fault direction (NNW-SSE). Both folding patterns end abruptly when they meet the main fault plane. Together with deformation, metamorphic grade also changes abruptly, from anchimetamorphic to low green-schist facies, and can be better observed in the pelitic facies of the Traíras Group, which to the east of the fault are described as siltstone and to the west are described as magnetite-chlorite-phyllite.

In addition, to the east of SC Fault, the Paranoá Group stratigraphic sequence is complete, as described by Campos *et al.* (2013), but to the west of the fault, the four lower formations, deposited over the Ribeirão São Miguel conglomerate, are absent. The following upper portion of the Paranoá Group rests over rocks of the Traíras Group directly, or are underlain by the São Miguel conglomerate, always along an erosive contact. The contact occurs with the Ribeirão do Torto Formation meta-argillite to the north and Córrego Piçarrão Formation meta-siltstone to the south. The greatest observable thickness of São Miguel basal conglomerates occurs near the SC Fault trace, especially in the western side.

The sudden absence of the Paranoá Group basal formations to the west of the SC Fault is investigated in this study. However, the possibility of tectonic control of the stratigraphy requires that regional architecture of tectonic structures is also investigated.

Campos *et al.* (2013) correlate the Paranoá Group stratigraphy in the Colinas do Sul region to the lower Paranoá Group, as did Fuck *et al.* (1988), despite the great resemblance of the pelite-carbonatic succession with the top Corrego do Barreiro Formation. After geological mapping along the interface between the Colinas do Sul and Alto Paraiso sequences, along the Serra do Cristal Fault, this work proposes the correlation of the Colinas do Sul column with the top of the Paranoá Group except by the basal São Miguel conglomerate.

As previously mentioned, the stratigraphic variation occurs abruptly at the SC Fault, so that to the west of the fault a stratigraphic gap of ca. 700 m occurs (Figure 5). However, the stratigraphic gap is not the only peculiarity of the western side, the Córrego do Barreiro Formation is thicker there, and only on the western side the Morro Vermelho conglomerate occurs, usually deposited over quartzite lens 1-10 meters thick. This conglomerate varies from clast to matrix supported, is reddish when weathered and contains poorly sorted angular to sub-rounded siltstone clasts (Figure 6a,b) and locally incorporates brachioid and venular quartzite boulders up to 6 m long (Figure 6c). The Morro Vermelho conglomerate thickness varies from zero to approximately 60 meters and can be noticed at distance as reddish and less vegetated elevations (Figure 6d). The mapping of its occurrences indicates that it was deposited in an erosive surface over the São Miguel Formation or directly over the Traíras Group, the basement for the sequence. The Morro Vermelho Conglomerate apparently coincides with the stratigraphic coplanar surface or flooding surface/sequence boundary that divides the Ribeirão Piçarrão Formation in sequences III and IV, probably representing a regional transgressive episode (Figure 2).

For simplification purposes, the formations that were not deposited over the higher basement block will be treated as Lower Paranoá sequence and the remaining formations as Upper Paranoá sequence. Therefore, the Lower Paranoá is comprised by the basal formations up to the intermediate Ribeirão Piçarrão Formation and the Upper Paranoá extends from the upper Ribeirão Piçarrão Formation to the Córrego do Barreiro Formation (Figure 5). A preliminary hypothesis is that basin subsidence may have caused paleofault reactivation and consequent abrupt relief variations, allowing erosion of the preceding formations.

As the basal formation, the São Miguel Formation is one of the most important units of the Paranoá Group, since it reveals details of basin installation mechanisms. It occurs on both sides of the SC Fault, but in the east side it occurs in much greater thickness, suggesting that initial depocenters were controlled by faulting.

The São Miguel Conglomerate contains angular to sub-angular venulated clasts (Figure 8a,b) and most rarely deformed clasts are found (Figure 8c). Metric scale blocks and boulders also occur, especially at the most basal deposits and near the SC Fault (Figure 8d). The contact with the underlying Traíras Group is marked by an erosive unconformity which is often planar (Figure 8e) but locally angular (Figure 8f).

The presence of angular and venulated clasts as well as boulders indicate faulting during basin installation, possibly reactivated during further basin evolution. Basement architecture was investigated via gravimetric inversion and modelling to clarify whether faulting was facilitated by previous structures.

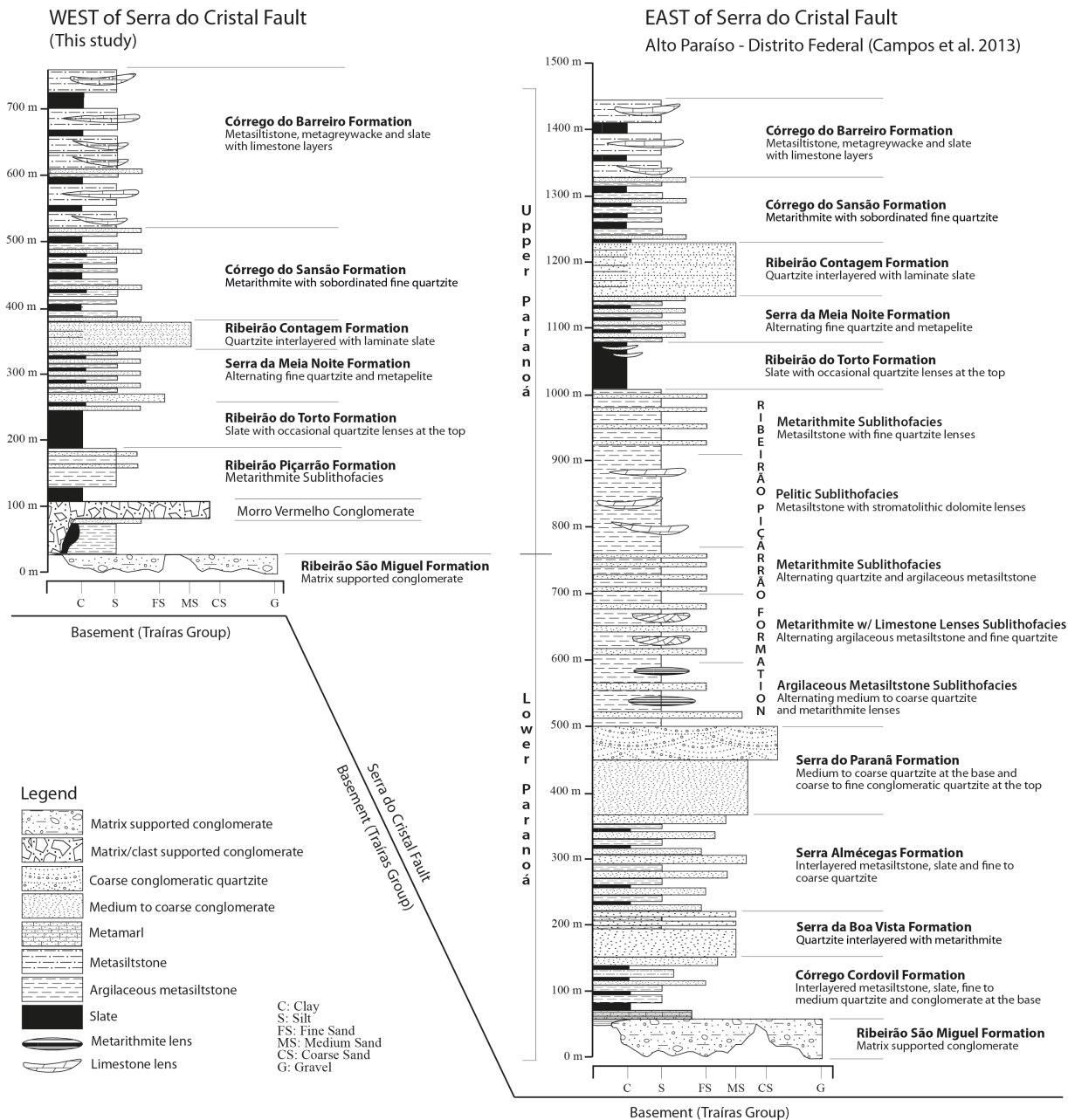


Figure 5 - Paranoá Group lithostratigraphic columns at both east and west sides of Serra do Cristal Fault. The west side column was intentionally built with a gap in order to emphasize the absence of most the basal formations. Basement is composed of Traíras Group rocks.

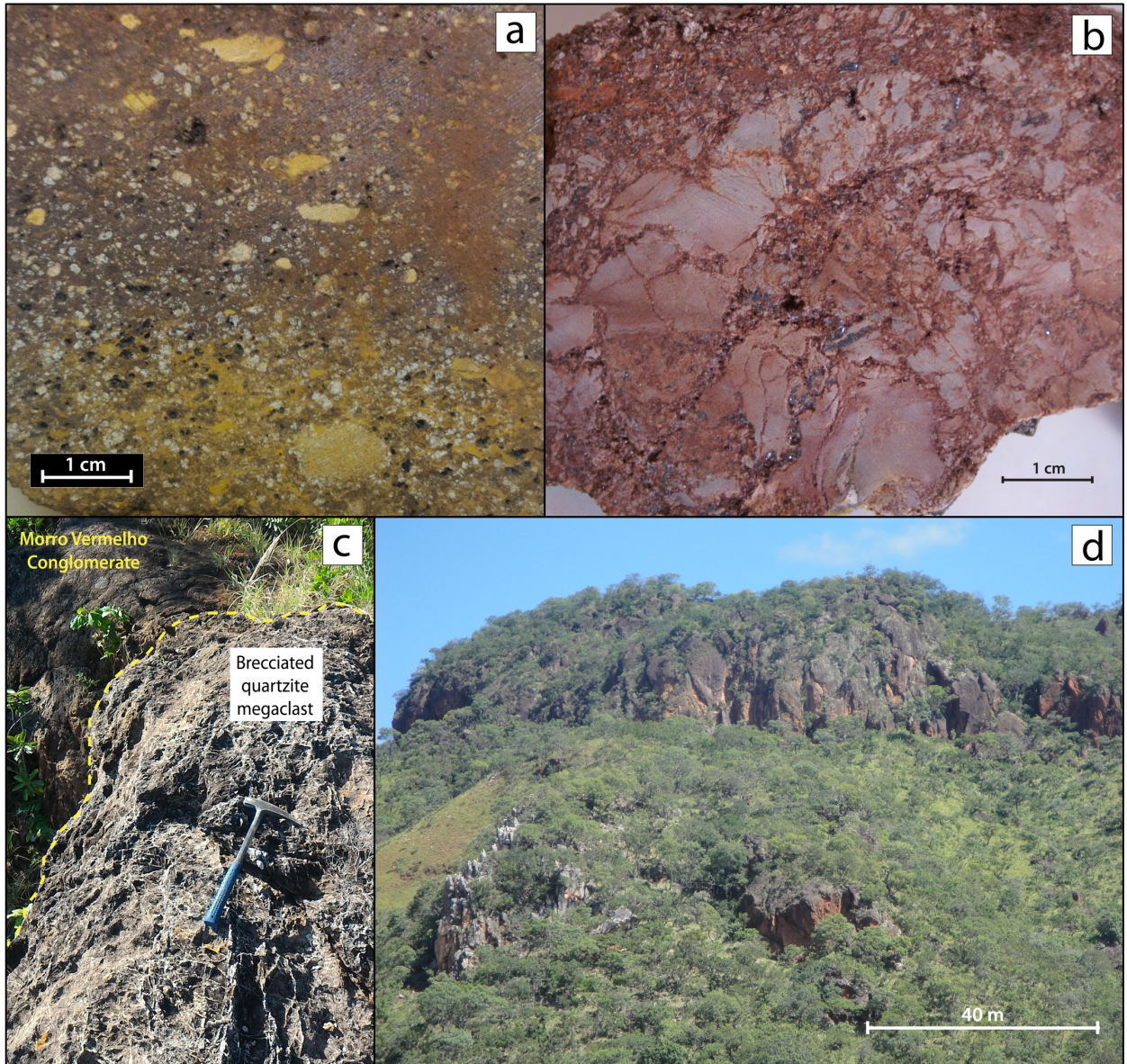


Figure 6 - Morro Vermelho Conglomerate **a,b)** hand samples **a)** occurring approximately 5 km from the Serra do Cristal Fault and presents reworked aspect with subrounded clasts; **b)** occurs ca. 200m from the fault and contain angular clasts of slate, siltstone and quartz vein, is matrix poor and contain occasional venulated and brecciated quartzite boulders **c)** Morro Vermelho conglomerate megaclast (boulder) composed of intensely venulated and brecciated quartzite boulder measuring ca. 6x4 m. **d)** ca. 120m elevation supported by the Morro Vermelho conglomerate, note underlain quartzite at the inferior left corner.

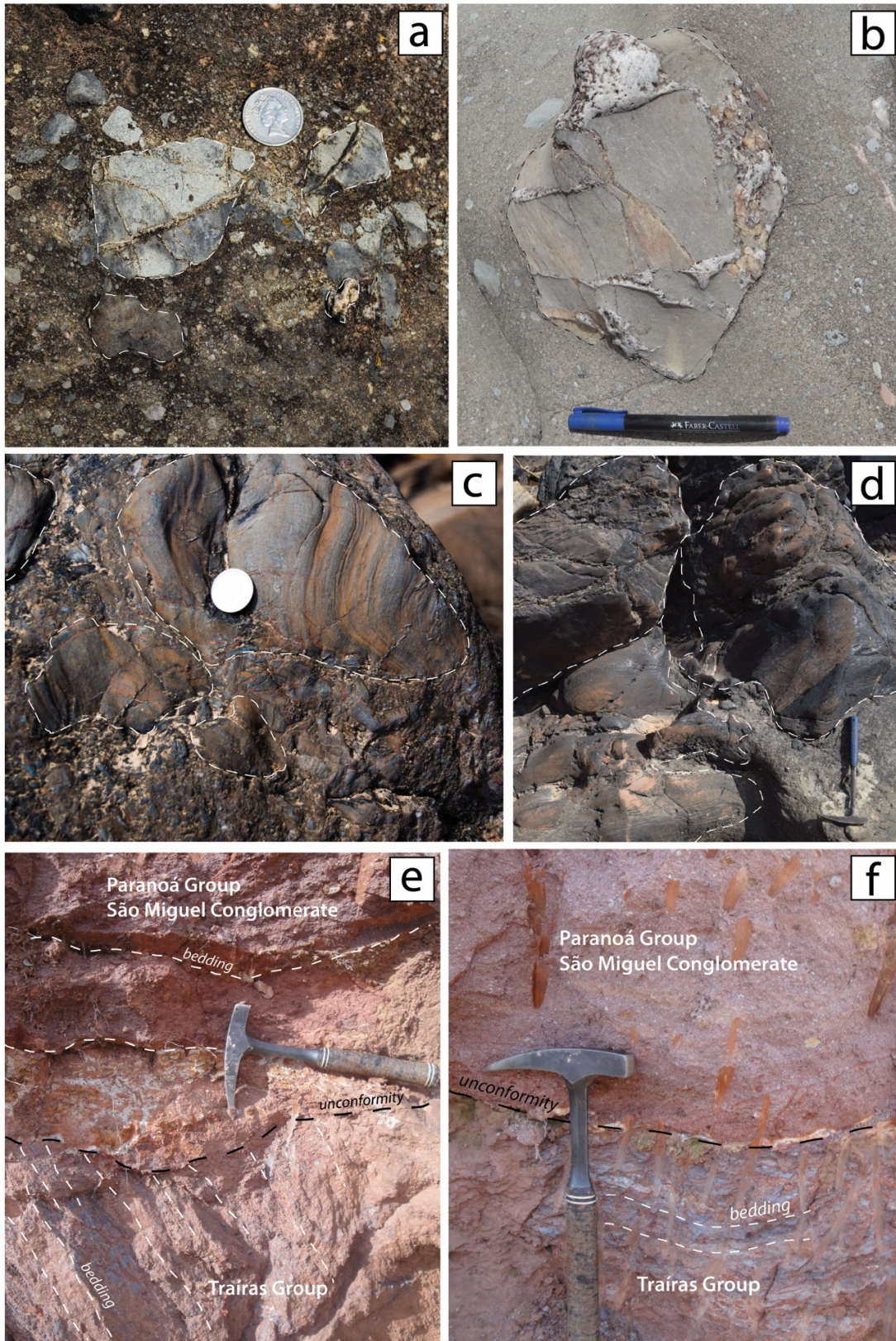


Figure 8 – São Miguel Conglomerate **a,b**) venulated phyllite angular to sub-angular clasts; **c**) slightly folded calcphyllite clasts (inside dashed contour); **d**) subangular metasiltstone megaclasts (inside dashed contour); **e**) angular unconformity and **f**) subparallel/erosive unconformity between the Trairas and Paranoá groups (dashed black line outlines unconformity surface and dashed white line outlines bedding).

3. Gravimetric data acquisition, preparation, inversion and modeling

Terrestrial gravimetric profiles were acquired along four NE-SW profiles transecting the Serra do Cristal Fault (Figure 9). The profiles are in average 20 km long with 500 m spacing between acquisition stations. The gravimetric data acquisition procedure involved the measurement of the gravity acceleration value using the Scintrex CG-5 gravimeter. The start and closure of each profile were carried out on IBGE (Brazilian Institute of Geography and Statistics) gravimetric bases for the recording of instrumental drift. The orthometric altitude and coordinates of each gravimetric station was carried out by means of the Trimble R4 GNSS receiver. The gravimetric data preparation involved corrections of instrumental drift, terrain and the calculation of Free-air and Bouguer anomalies, and the resulting curves are shown in figure 10.

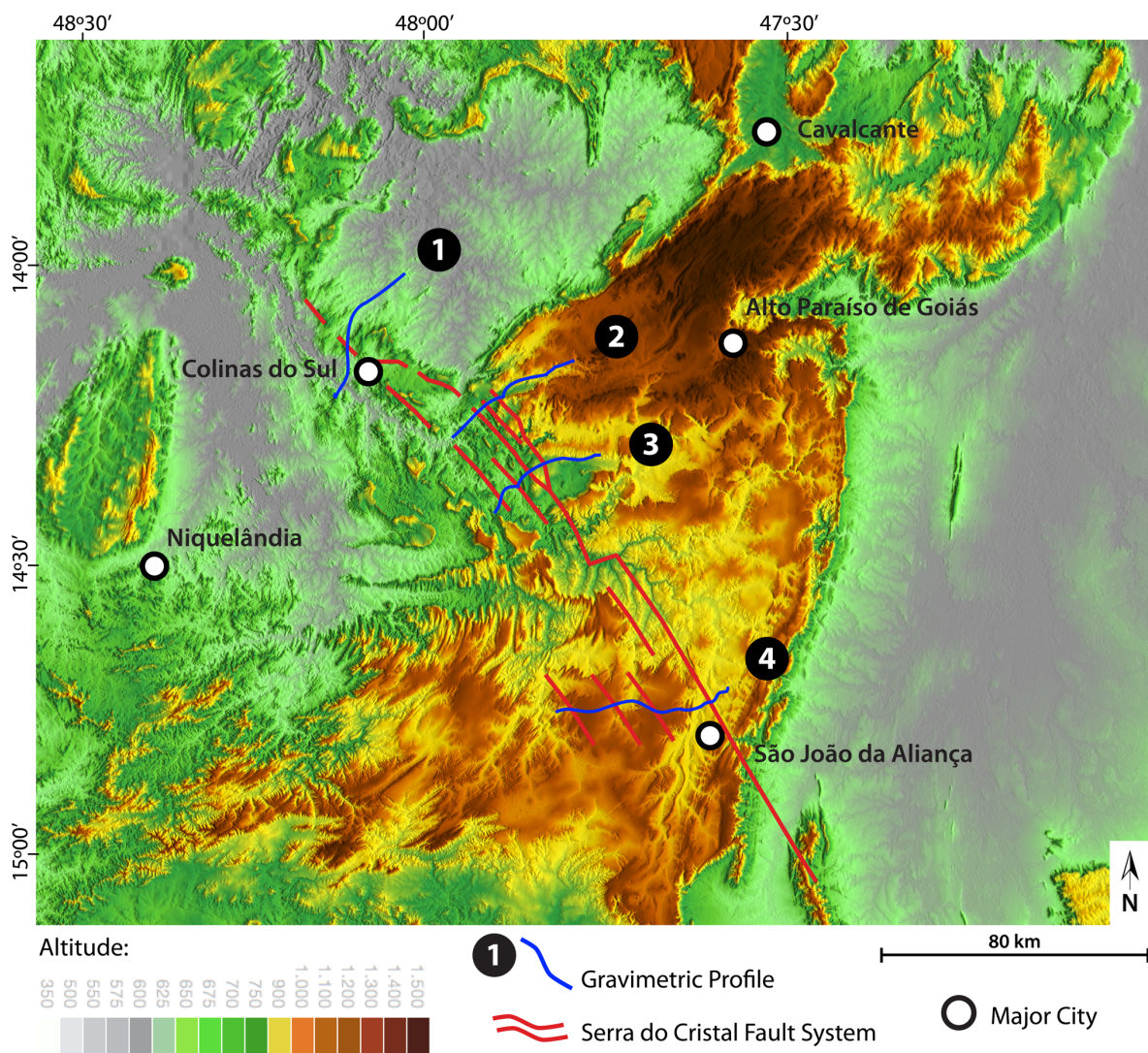


Figure 9 - Digital elevation model map displaying the location of the terrestrial gravimetric profiles and the Serra do Cristal Fault system traces.

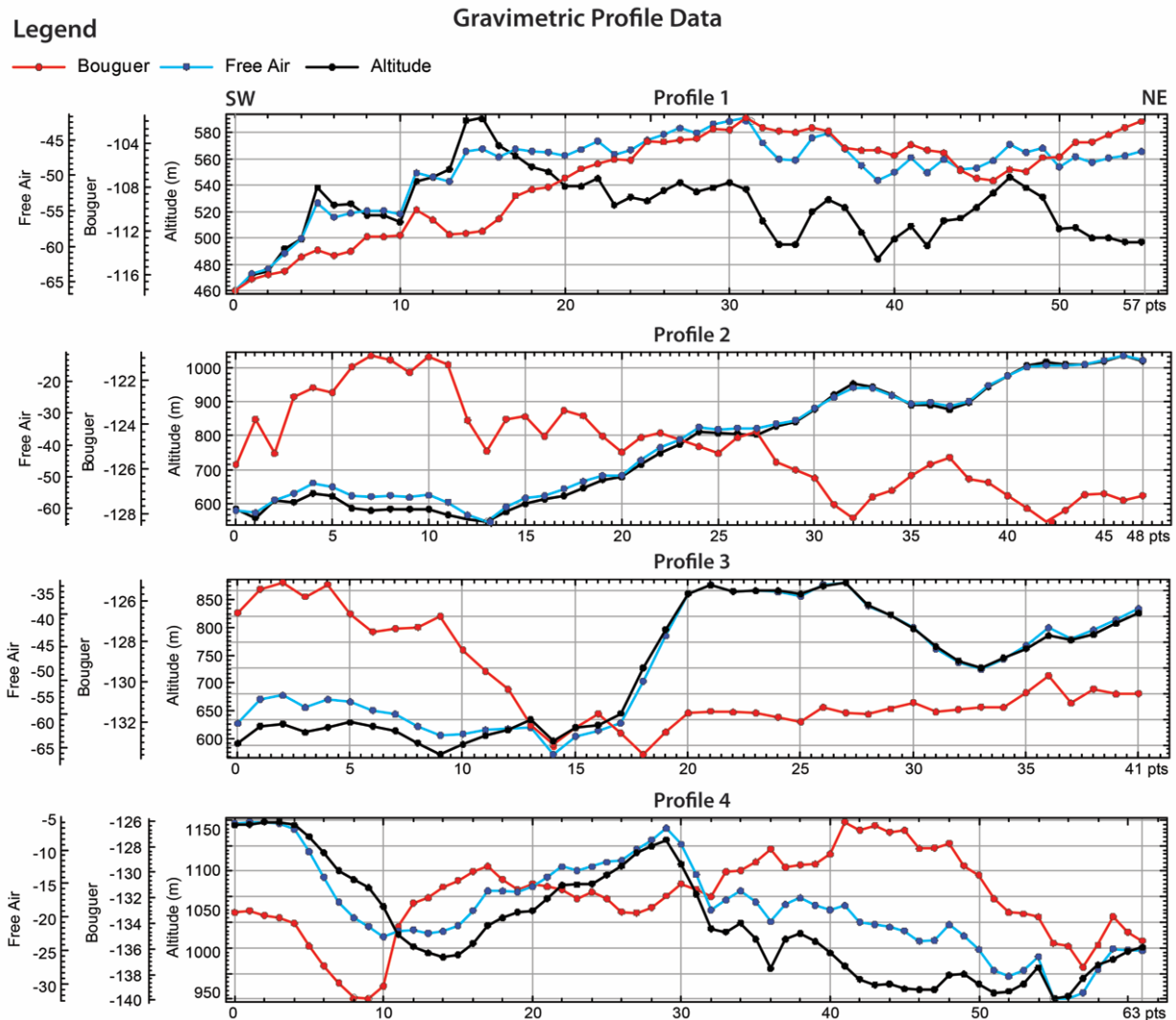


Figure 10 – Corrected gravimetric profile data showing altitude (m), Bouguer (mGal) and Free-air (mGal) data. See profiles location on figure 9.

Regional-residual data separation procedure involved subtracting the upward continuation results for 10000 m altitude from the Bouguer anomaly. The residual anomaly was used for gravimetric inversion and modeling.

Gravimetric profile data inversion was conducted using the softwares: forward modeling was carried out in LMG2D and inverse modeling in FORTRAN (Silva et al. 2001) and MATLAB by the algorithm for least-squares estimation proposed by Marquardt (1963). Gravimetric profile modeling was performed via Oasis Montaj software using GM-SYS extension. Field based geologic profiles with accurate surface and inferred subsurface geology were used for model control.

Rock density mean values for the main rock types were adopted from Sharma (1997), considering the stratigraphic proportions of quartzite, pelite and limestone in local sequence stratigraphy as well as depth, since rock density tends to increase with depth. The values were

tested during automated data inversion and the best-fit results were 2.45 g/cm³ for the Paranoá Group, 2.55 g/cm³ for the Traíras and Araí groups (undivided in this study) and 2.67 g/cm³ for crystalline basement rocks.

4. Results and Discussion

The gravimetric inversion results (Figure 11 – left column) served as general guides for the gravimetric profiles modeling. The main difference between anomalies induced by lithological variations and basement-topography variations is that the former have lower frequencies and higher amplitudes than the latter (Gibson 1998). Our gravity data show relatively low amplitude variations and high frequency variations, except for profile 1. The models fixed points were placed either where basement rocks outcrop or where the Bouguer anomaly is more constant. Considering that we are working close to a rifted crust, the modeling approach consisted in adjusting basin depth and basement blocks as to attain the lowest possible error values. Modeling errors achieved by our geological model were surprisingly low and the results revealed that basement architecture under the Paranoá and Traíras sequences is more complex than previously inferred and that the Araí rift continues to the south under the Paranoá sequence, as indicated by the basement horsts and grabens exhibited on the gravity modeling products (Figure 11 – right column).

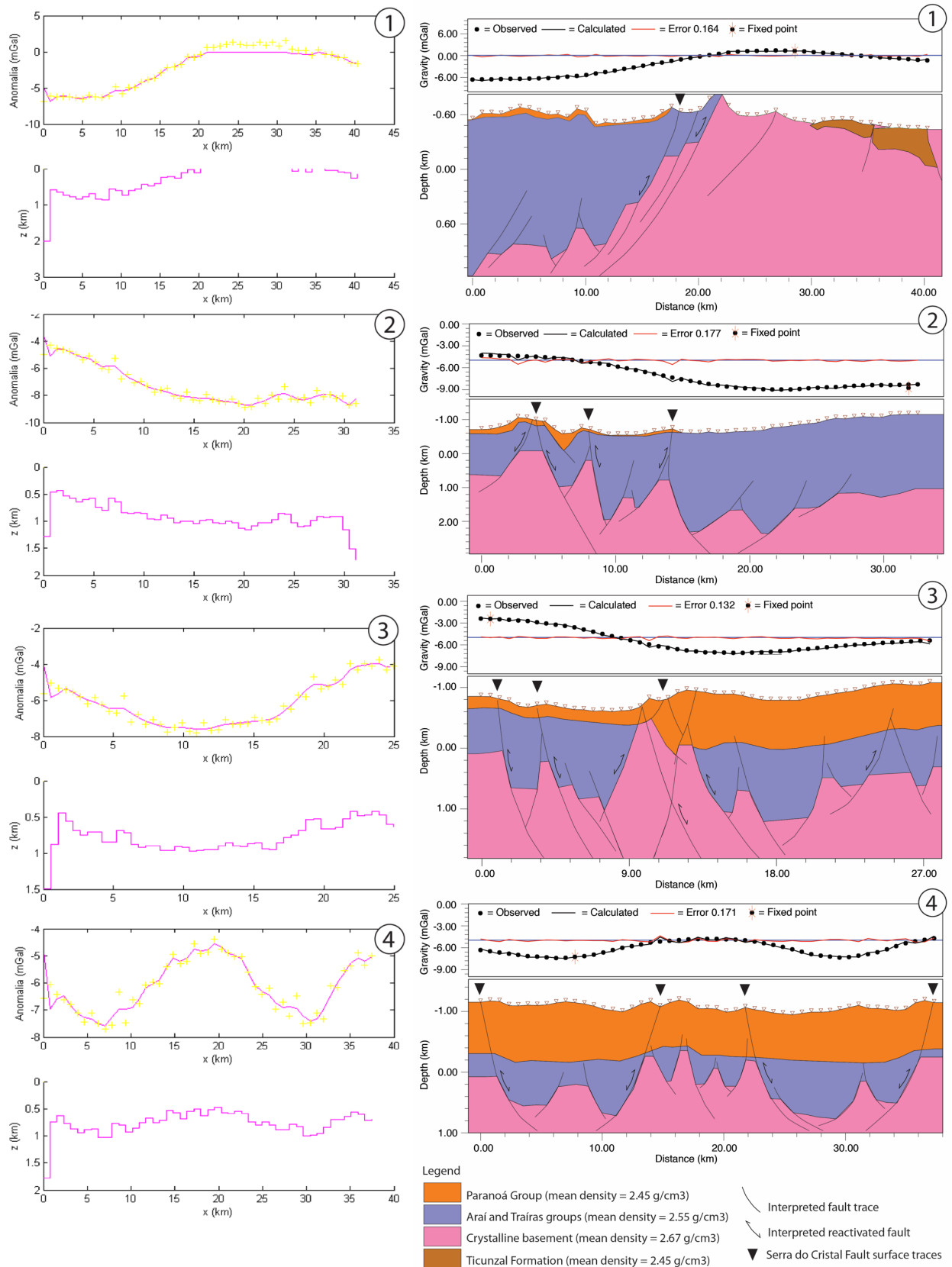


Figure 11 – Gravimetric inversion (left column) and direct gravimetric modelling (right column) results of terrestrial profiles 1 to 4 and interpreted reactivated basement faults. Serra do Cristal Fault System traces are indicated by black triangles.

The gravimetric models indicate that Paranoá sediments are underlain by Araí rift structures. A seismic profile further south also shows the presence of rift structures under Paranoá strata and suggests that the group is not allochthonous along the seismic line location (Figure 12). Similar basement architecture scenarios can be found elsewhere, as the Spitsbergen's Late Paleozoic successions, where intracratonic sequences are underlain by an earlier rift (Figure 13).

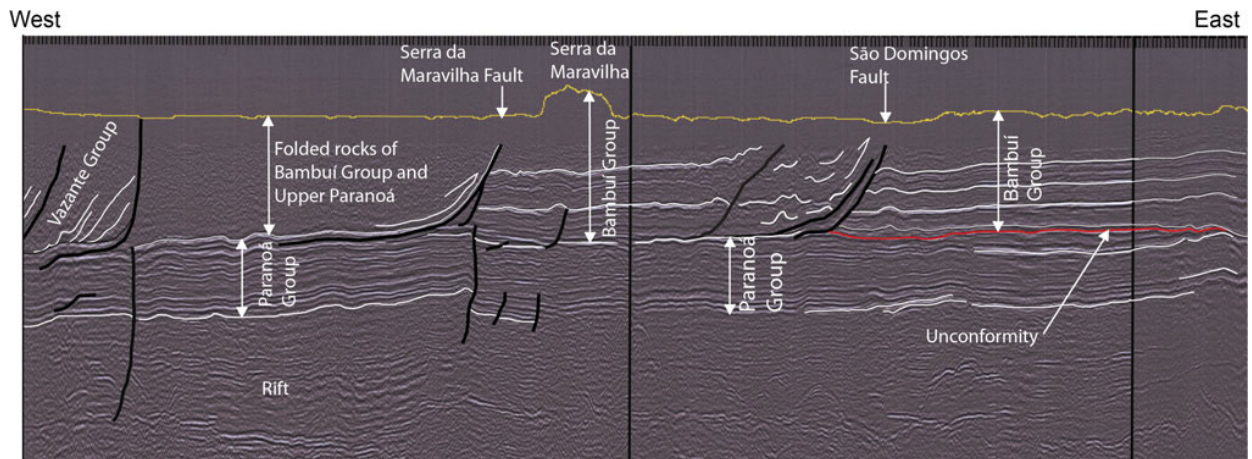


Figure 12 - Regional seismic profile across the external Brasília belt Bambuí, Vazante and Paranoá groups strata underlain by rift structures (from Alvarenga et al. 2012).

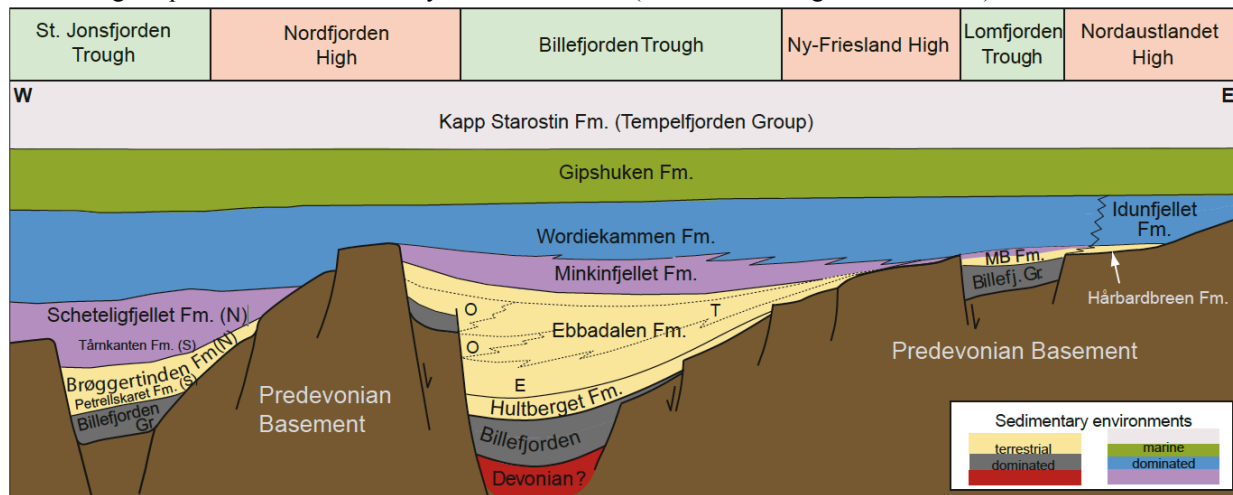


Figure 13 - Schematic E-W-trending cross-section, showing the framework of the lithostratigraphic units across Spitsbergen's Late Paleozoic trough and highs, Norway (Scheibner et al. 2015).

The Paranoá paleocurrent directions were measured by Faria (1995) in both basal and top formations on cross-bedding structures in three different regions, in the northern (n=210), central (n=194) and southern (n=432) areas of Paranoá occurrence. The paleocurrent data were plotted on map together with Araí Paleorift and Pirapora Aulacogen structures (Figure 14) and reveal an impressive match with the direction of rift and aulacogen main axes, suggesting that the Paranoá basin installation and further evolution was highly influenced by paleorift fault reactivation,

which initially allowed rapid subsidence and the formation of gravitational flux conglomerates and continued to exert influence also on the upper formations by dictating sediment flux direction (up to the Ribeirão Contagem Formation).

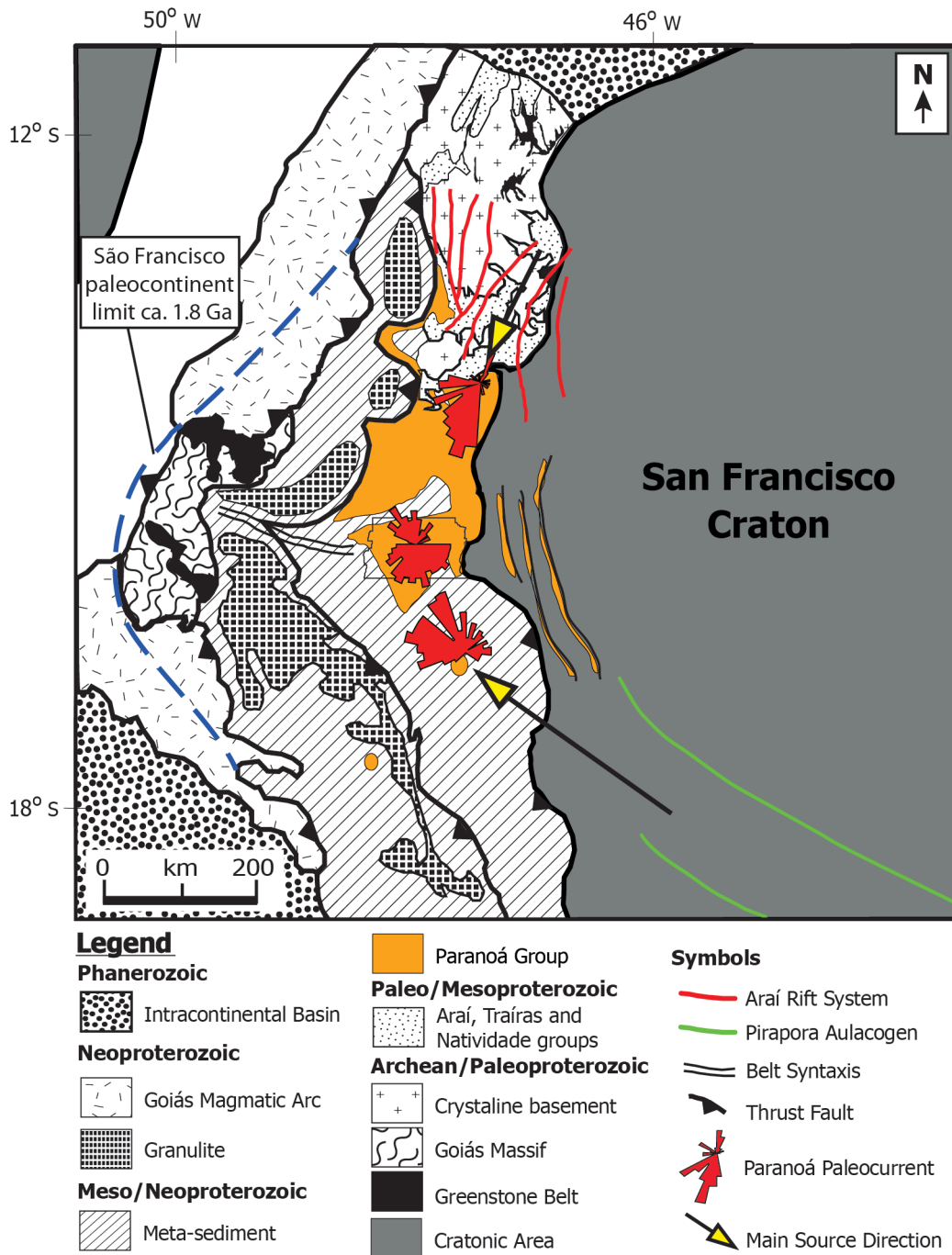


Figure 14 – Brasília Belt Map showing known rift and aulacogen paleo structures and Paranoá paleocurrent directions, measured by Faria (1995) on cross-bedding structures in three different regions, northern (n=210), central (n=194) and southern (n=432) areas of Paranoá occurrence.

4.1. Results Discussion According to Literature Review

The term *craton* is generally assumed to be the area of old, strong lithosphere beyond the reach of marginal tectonic processes (Sloss, 1988a). That definition is either time and space limited, time-limited because “old” is not defined, and space-limited because it is difficult to define a limit for orogenic deformation without considering crustal depth. Burgess (2008) considers the North American Craton to be the area underlain by Precambrian basement that has not been subject to plate-margin processes during the Phanerozoic, which does not exclude significant tectonic uplift, subsidence and deformation, simply that the tectonic processes affecting cratonic interiors are epirogenic and not orogenic.

Bradley (2008) argues that Mesoproterozoic passive margins seem to have lasted longer, reaching up to 600 Ma., however, affirms that data for that record basin lifespan is unreliable. Based on the overall data from passive margin basins he calculates a general average passive margin lifespan of 187 Ma. and cites the western São Francisco craton passive margin as having lasted 105 Ma, from 745 to 640 Ma., but does not cite the source from that data. According to the author, the abundance of passive margin basins is related to the supercontinent cycle, during assembly of a supercontinent a decrease in the world's population of passive margins is recorded, and during supercontinent break-up, passive margins increase in number.

Allen & Allen (2013) interpret the evolution of the cratonic basins as related to the rift-drift system, but formed by low extensional stretching factors. This mechanism may be associated with major faults observed below most intracratonic basins (e.g., Western Siberia Basin, Illinois Basin, Chad Basin). Armitage and Allen (2010) report that several mechanisms for the formation of cratonic basins have been proposed (see Hartley & Allen, 1994, Table 1), which are based on: thermal relaxation as a result of the extension or placement of warmer mantle; inferior continental lithosphere density variations; localized extension related to magmatic ascent or reactivation of pre-existing basins.

In a classic review paper exclusively on passive margins, Bradley (2008) defines the term “passive margin” as synonym of Atlantic-type margin in a broad sense, rifted margin or divergent margin. A passive margin is formed by rifting, followed by oceanic spreading, resulting in a plate consisting of oceanic and continental lithosphere sections welded by an igneous contact. The continental portion of a platform basin that leads to a passive margin basin is thus not considered a passive margin environment.

Intracratonic basins last longer than any other basin type, they commonly develop as initial rifts that fail to diverge further and undergo thermal subsidence, typically involving several

reactivations over several hundred million years. Rift basins developed over continental crust and may have three different fates, evolve to a proto-oceanic rift and continue to become passive margin or ocean basins or evolve to aulacogens, which can later generate intracratonic basins (Figure 15). Passive margin basins usually last 100-200 Ma. and intracratonic basins may last 200-500 Ma (Woodcock 2004).

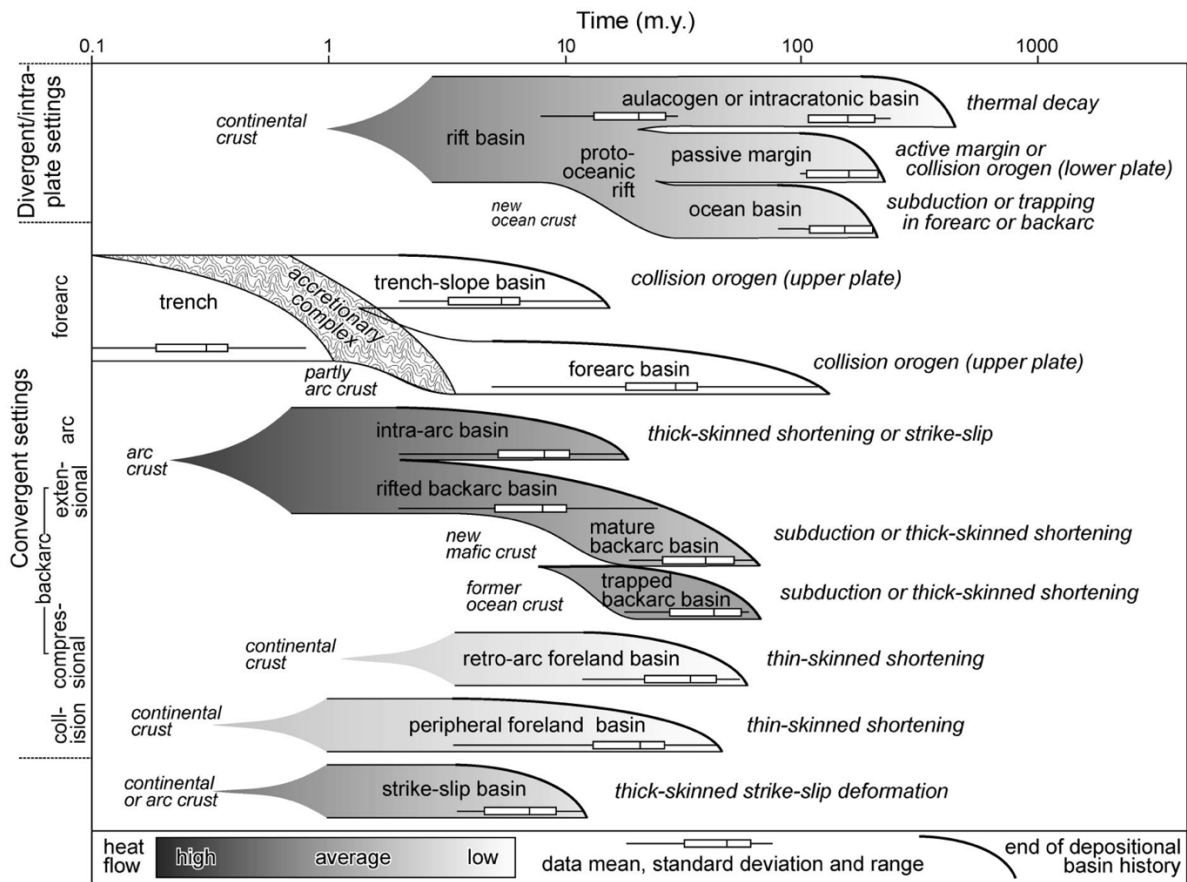


Figure 15 - Typical fates of different classes of sedimentary basin through time, plotted on logarithmic scale. Small italics show basement to each class. Large italics show deformational or thermal consequent fates of each class (from Woodcock 2004).

Passive margins are characterized by large thicknesses of sediments (up to ~ 12 km). Back stripping of restored stratigraphic cross-sections of ancient passive margins preserved in deformed orogenic belts reveal patterns of tectonic subsidence and uplift that resemble those in modern margins, so ancient passive margin sequences should be equally thick (Watts 2012).

For Coleman and Cahan (2012), passive margin basins, which extend onto oceanic crust, are those passive margin basins that typically develop between two cratonic masses and extend onto transitional and oceanic crust. There is obviously a continuum between this type of basin and pericratonic passive margin basins (including deltaic). An example of this type of basin is the Canada Basin of the United States and Canada.

According to Watts (2012), the Wilson Cycle implies that passive margins ultimately become sites of orogeny. Although the mechanisms by which this transition takes place are unclear (e.g. Erickson, 1993), there is evidence from back stripping of biostratigraphic data that foreland basins are underlain by stretched crust (e.g. the western deep Gulf of Mexico basin which overlies the western Gulf Coast margin, the Papuan basin which overlies the Northwest Australia margin and the west Taiwan basin which overlies the South China Sea margin).

Due to plate tectonics, all ancient passive margin basins were trapped within orogenic belts and are not preserved over ocean floor. However, many of these sequences can still be recognized by some tectono-stratigraphic evidences, as pointed out by Bradley (2008); shallow water deposits flanked by contemporary deep water facies, inferred from being deposited on stretched continental lithosphere or oceanic lithosphere, the existence of sequences of ophiolites and/or magmatic arcs pushed over the platform deposits, providing evidence that an ocean basin existed. Deep water facies and ophiolites also help in distinguishing between a passive margin involved in collision and a deformed (inverted) intracontinental basin that was flanked on all sides by continent and never developed a true ocean. However, the most characteristic feature of a passive margin is miogeoclinal prism - a sedimentary wedge that thickens to the sea and extends for 15 km or more from a trestle border.

Detail analyses of sedimentary structures and facies succession indicates that all the Paranoá sequence depositional environments correspond to shallow marine conditions, always influenced by storms and tides, with frequent structures of corrugated marks, cross strata, hummocky, etc., even on thinner granulometry sequences, in which shallow water stromatolites, salt pseudomorphs and intercalations of conglomeratic quartzite also occur frequently (Faria 1995, Dardenne 2000, Alvarenga et al. 2012, Campos et al. 2013). It is important to emphasize that strictly deep sedimentary environments were never mapped in the whole preserved succession. The variation of sand and clay material proportions is indeed related to small changes in water level, but also occurs due to subsidence rates variation and changes in sediment supply.

The Paranoá sequence presents satisfactory stratal preservation, however, detailed time control is otherwise difficult because as a Mesoproterozoic sequence, fossil dating is not a viable option and provenance studies are compromised by the lack of age diversity of igneous sources. Thus, available age data with detailed stratigraphic controls is still basic and exists only for the Lower Paranoá sequence. As a way to work around the problem, Alkmim et al. (2012) adopted the concept of 1st-order sequence (Martins-Neto et al. 2001, Catuneanu et al. 2005) for

chronocorrelated sequences at the São Francisco Craton eastern margin. Eriksson et al. (2012) used the same approach for several other cratons, including the São Francisco, in a global correlation study. They proposed the stratigraphic hierarchy as an approach to interpret the stratigraphic framework of sedimentary basins.

The technic involves separating the sequences in sub-sequences and identifying the bounding surfaces that divide them, based on their relative importance. The approach enables to identify tectonic changes reflected in the sedimentary record, providing thus a tectonic criterion for the subdivision of the rock record. According to Alkmim et al. (2012), a first-order sequence corresponds to an entire sedimentary basin-fill succession, regardless of the origin and life span of each particular basin. This first-order basin-fill succession can be in turn subdivided into second- and lower-order sequences, based on the steps involved in the tectonic evolution of the basin, irrespective of the time span between two same-order consecutive events. Besides basin subdivision, the approach also allows the classification of basin types, according to some key criteria, adopted in this paper.

According to Dickinson (1976a), first-order criteria for classifying sedimentary basin types are (1) type of nearby plate boundaries (2) proximity of plate boundary, and (3) basement crustal type. First-order classification based on criteria (1) and (2) can belong to divergent, intraplate, convergent, transform, hybrid, or miscellaneous settings. Each of these categories may vary depending on their basement crust type (continental, transitional, oceanic, or anomalous), as well as structural position, sediment supply, and inheritance of previous structures (Ingersol 2012). Basin classification and nomenclature have to be chosen based on characteristics of a basin at the time of sedimentation, thus any type of basin that later evolved to a passive margin should not be classified as such.

The first-order classification criteria were applied to the Paranoá basin at the study area, resulting as follows: (1) type of nearby plate boundaries: passive; (2) proximity of plate boundary: 100-200 km, and (3) basement crustal type: continental. According to the interpretation scheme proposed by Ingersol (2012, Table 1.1), the Paranoá sequence at the São Francisco Craton western margin could be classified as one of the Intraplate Continental Margin basin types, from which Continental Platform and Intracratonic basins apply.

Continental platforms are defined as basins over stable continents with thin and laterally extensive sedimentary strata, a modern example would be the Barents Sea and an ancient the Middle Paleozoic, North American basin.

Intracratonic basins are defined as broad cratonic basins over relatively thick crust underlain by fossil rifts, a modern example is the Chad basin and an ancient example is the Paleozoic Michigan Basin (Ingersol 2012). In addition, the intracontinental basin-generating subsidence generally occurs when lithospheric rigidity decreases over thousands of years, so that an uncompensated mass in the upper crust (remnants of fossil rifts) subsides over a wide area.

The terms intracontinental sag (Allen and Armitage 2012) and cratonic sag (Allen et al. 2015) are also used as synonyms of intracratonic basins. The use of *cratonic* or *continental* suffixes seems to be widely applied without strict criteria to distinguish between the two terms, but in fact it is the definition of craton that varies.

The most accepted limit for the São Francisco craton was proposed by Almeida (1981) as a modification from his original proposal (Almeida 1967) and includes the thin-skinned portion of the foreland fold-thrust belt. This limit is adopted in this paper and used for basin spatial distribution in figure 20. However, we recognize that the original proposal, which included only undeformed and unmetamorphosed Neoproterozoic rocks is more adequate for basin analysis in the western margin, because the newer limit includes the Vazante Group and a portion of the part of the Paranoá Group that is still more than 1 km thick and thus differs significantly from the truly cratonic, thin Paranoá sequence, only a few hundred meters thick.

Our data indicates that, in the studied area, the Paranoá sequence is underlain by a fossil rift, whose faults have been reactivated during basin installation, facilitating generalized subsidence and relative uplift of epirogenic arches that segmented the basin. As discussed before, basins over continental crust underlain by fossil rifts can be classified as intracratonic, however, intracratonic basins are usually much thicker than the 1.5 km presented by the Paranoá Group in the study area. In relation to that, Burgess (2008) evaluates that although areas surrounded by arches are sometimes referred to as intracratonic basins, such areas are still better considered as continental platforms or cratonic margin basins, rather than intracratonic basins due to the obvious differences in the thickness of total preserved sediment.

In contrast to intracratonic basins, cratonic margin basins accumulate sediment of uniform thickness over continental scales (Ingersol 2012) and their stratigraphic sequences can be transitional into passive margins, intracratonic basins, foreland basins, and other tectonic settings along continental margins (Figure 16) (Ingersoll and Busby, 1995; Burgess, 2008).

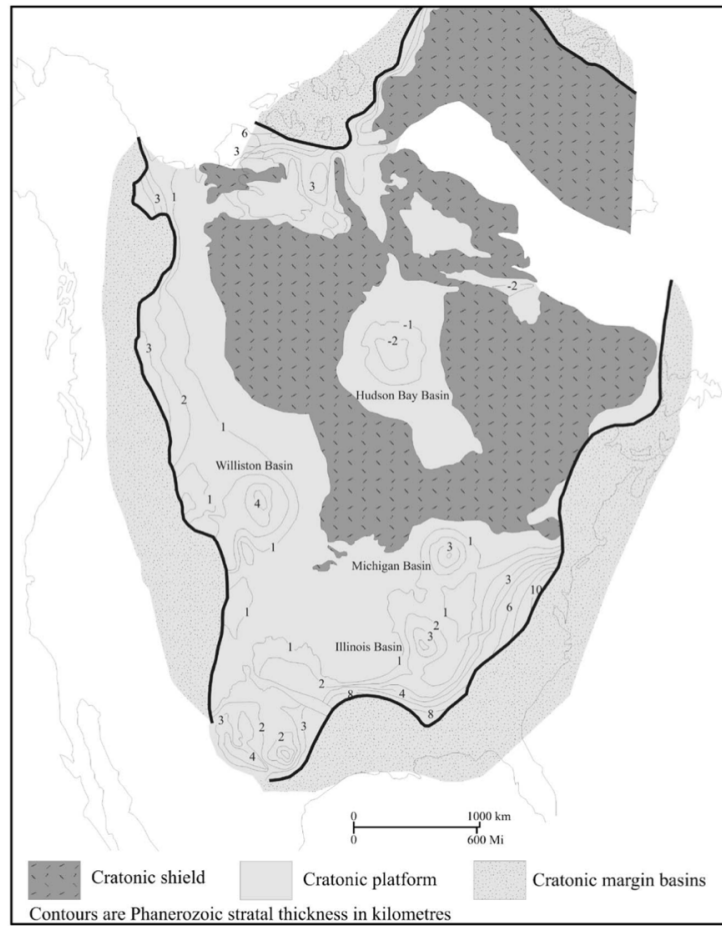


Figure 16 - Map of stratal thickness on the North American craton presented by Burgess (2008).

By the first-order characterization criteria the same basinal system with continuous strata can be characterized as different basin types according to its tectonic setting. Therefore, if analyzed in all its extension, the Paranoá sequence can be characterized as a continental platform basin when deposited over the São Francisco Craton, with thin and continuous strata. Over ancient rifts that affect the craton, it may be characterized as an intracratonic basin sequence, with greater sediment depths, since paleorift faults can facilitate greater subsidence rates. Over the younger, thinner, more recently accreted continental crust that formed the margins of São Francisco paleocontinent, it can be characterized as a cratonic margin type basin with limited subsidence (Figure 17). Nowhere, though, the Paranoá sequence displays passive margin characteristics, at least in the preserved sequences.

Between episodes of orogenic activity, the lithosphere is strengthened so that the achievement of local isostatic equilibrium is interrupted. Thus, an intracontinental basin can take hundreds of millions of years to achieve total isostatic compensation (Ingersoll and Busby, 1995; Howell and van der Pluijm, 1999).

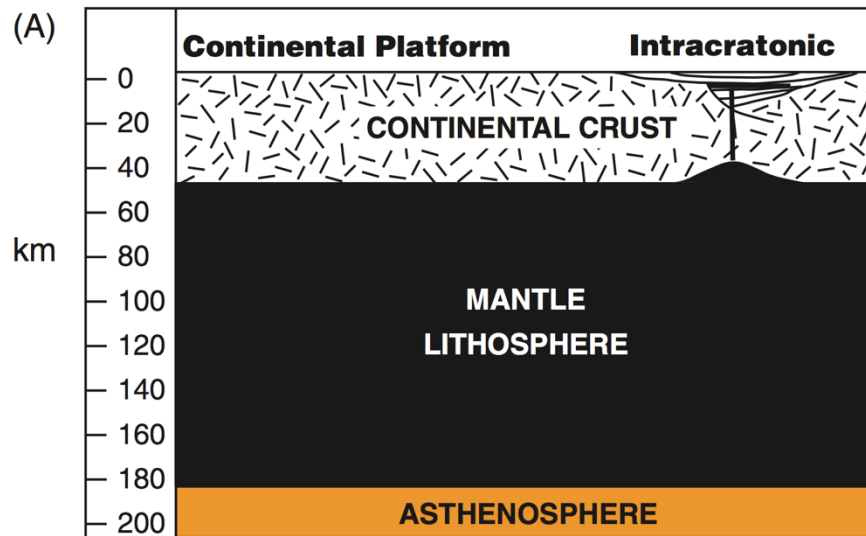


Figure 17 - True-scale actualistic analog models for sedimentary basins in intraplate setting (after Ingersoll 2012).

It is quite common for ancient intracratonic basins to be confused with other types of basins. Ingersoll and Busby (1995) warn that when a dormant ocean basin is full at sea level, it may appear superficially an intracratonic basin, but a dormant ocean basin, however, contains between 16-20 km of sedimentary strata piled over oceanic lithosphere or stretched, transitional continental crust, while the intracontinental basins contain only a few km of sedimentary strata piled on regular continental crust, with one or more fossil rifts below the center of the basin.

In dormant oceanic basins, the term "dormant" implies that there is no orogenic or tectonic activity within or adjacent to the basin; the term "oceanic" requires that the basin be superjacent to the oceanic lithosphere, in contrast to the intracontinental basins, which are typically overlying a partially fragmented continental lithosphere. Intracratonic basins may undergo repeated subsidence or tectonic inversion pulses (e.g., the modern North Sea) (Cooper and Williams, 1989, Cameron et al., 1992).

In the Glossary of Geology of the American Geological Institute, the term platform is defined as part of a continent covered by flat or gently sloping sedimentary rocks, not necessarily connected to a passive margin, covering a complex of rocks that have consolidated during previous deformations. Platform areas can be considered sedimentary basins in the sense that they are areas of sediment accumulation. It adds, however, that the platforms are not basins in the sense of an area experiencing differential subsidence in relation to the surrounding area of stable basement, this definition, which applies to intracontinental basins, but not to continental platforms.

The intracontinental basins within the platform show significantly greater thicknesses of preserved strata than the surrounding areas of the platform, greater than 4.5 km of strata in the

Michigan basin, compared to the average 1 km strata in the surrounding platform area. Subsidence mechanisms are not well understood in either case, but there is a clear distinction in terms of the thickness of preserved strata.

According to Armitage and Allen (2010), the most striking feature of the continental or cratonic basins is their long periods of subsidence, which begins during periods of continental dispersion and can continue through cycles of ocean closure and continental collision. The cratonic or continental basins are characterized by shallow marine and terrestrial sedimentation, and a relatively simple "cake layer" stratigraphy. They are located inside continental plates surrounded by passive margin basins, and are occasionally connected to the sea.

As pointed out by Burgess (2008), a significant feature in continental or cratonic platforms is the presence of epirogenic arches. An epirogenic arch is a high intraplataformal block that subsides less rapidly than the surrounding areas of the platform, leading to the formation of relatively lean strata, or is erected, leading to erosion and formation of local nonconformity (Figure 18). The arches are significant because they separate areas of the platform and appear to have acted as important elements in continental paleogeography. They are also important because they provide a clear indication of epirogenetic tectonic activity within the continent during sedimentation of the basin. The blocks that form epirogenic arches can be formed by previous extensional or compressional processes which generate weak zones separating crustal segments as well as density contrast between blocks of different source depth.

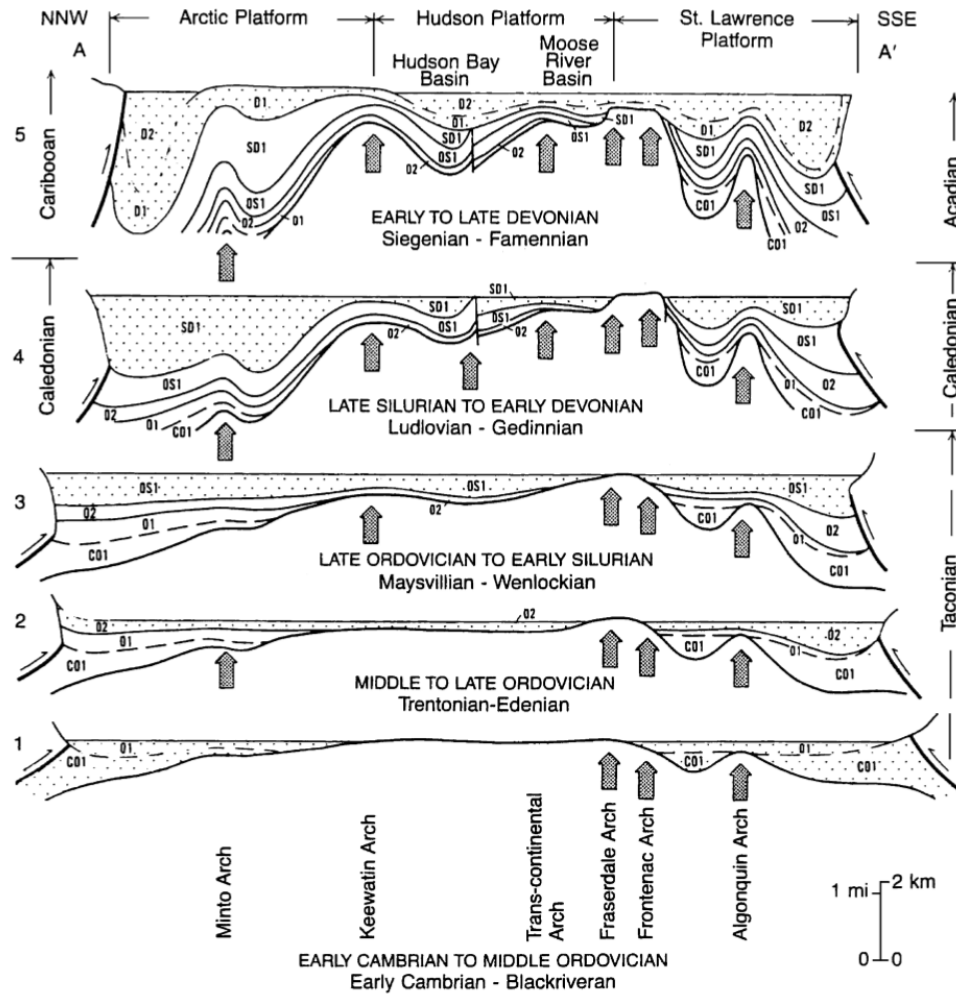


Figure 18 - Schematic cross-sections over the Canadian Craton showing the development of epirogenic arches in time (Sanford 1987).

Intracratonic basins represent the classic type of sedimentary basin. Global examples include the Williston, Michigan and Illinois basins in North America, the Maranhão basin in Brazil, and the Murzuk and Kufra in Africa, among others. Their interconnection to the sea is evidenced by intermittent phases of marine carbonate and evaporite sedimentation. During these periods they might, therefore, be more truly termed embayments rather than basins. Basins of this type are frequently separated by raised linear areas where sediment cover is thin or absent. These high areas are usually termed arches, paleo-highs, schwelle, axes of uplift, or positive areas (Selley 2000).

Intermittent uplift of the embayment open rim, closes the connection to the ocean, when evaporite or continental sedimentation may follow. The Murzuk basin of Libya is an excellent example of that process. The Proterozoic Paranoá basin also seems to fit perfectly in that scenario.

Epicratonic basins are those which lie on the edge of continental crust, similar to cratonic margin basins, and for being at the edge, instability is often seen in the sedimentary record. However, as the intracratonic basins, epicratonic basins also may lie within continental crust but be partially open to an ocean basin, once both often occur adjacent to one another, with little fundamental difference in genesis or fill (Selley and Sonnenberg 2014).

Examples of epicratonic basins include those of the Mississippi Gulf coast, the Niger basin and the Sirte basin of North Africa. Initial basin subsidence can be rapid due to marginal instability, resulting in an early phase of deep-water sedimentation or sub-aerial debris flow. Over previously faulted basement, rapid initial subsidence happens by local fault reactivation, which does not configure rifting, as in the classical sense, and subaerial or subaqueous gravitational debris flux deposits may occur. The Sirte basin basement, for instance, was extensively faulted some igneous activity occurred (Selley 1997).

Like intracratonic basins, epicratonic embayments can be infilled by both carbonates and terrigenous sediment. This differentiation is a generally function of the degree of uplift of the adjacent crust. Allen and Armitage (2012) account that some suites of cratonic basins originate as broad ramp-like realms of subsidence tilting down to the adjacent passive margin, and can be later “individualized” by secondary tectonic processes such as the emergence of intervening arches and domes.

The tectonic mechanisms and events that affect cratonic margins vary in duration and spatial extent, so that there is generally no clear distinction between marginal and epirogenetic processes. For example, the Colorado Plateau is a largely intact cratonic block, surrounded by an area of deformation related to margin processes, such as flexural load and plate effects (Burgess 2008). The processes that may explain the formation of intracratonic basins and cratonic platforms with related cratonic arcs are Eustasy, Extension and thermal re-equilibrium and Intraplate stress.

The distinction between tectonic and eustatic forces in temporal and spatial scales can be difficult since they represent different aspects of the same processes (Gurnis, 1990).

The extension of the lithosphere by tensive tensions generates extensional basins formed by the combination of initial subsidence governed by normal faults, followed by subsidence related to thermal rebalancing (McKenzie, 1978). This model has been widely used to explain several intracratonic basins around the world.

Burgess (2008) argues that the intraplate stress model, as a form of unconformity generation in the continental basins, has been criticized because synchronous activity in the

whole craton can be difficult to explain through this mechanism, since it would require that tension field stresses were developed simultaneously throughout the craton. However, he considers that, in the case of cratonic sequences, the synchronicity operates at scales of a few million years and therefore the synchronicity would not represent a model flaw.

In addition, examples of this type of synchronic effect have been documented, as the case of the African plate, where the obliteration of ophiolites in the northeast margin of Afro-Arabia seems to have generated an unconformity related to intraplate deformation during a period of 2 Ma. in the Senonian Period, covering more than 20 million km².

Repeated changes in the configuration of sedimentary basins and in the mechanisms governing their subsidence usually reflect changes in the stress systems affecting the craton and its thermo-mechanical properties (Nikishin et al. 1996). Changes in the intraplate stress regime are mainly controlled by processes affecting the plate boundaries (Zoback 1992; Zoback et al. 1993).

Intraplate stress variation over time may be related at some degree to orogenic processes and other plate tectonic events occurring in the cratonic margins. Marshak et al. (2003) describe the cratonic features of continental arches as evidences of intraplate stresses, especially due to the complex history of subsidence and uplift, which may be related to inherited basement faults reactivated by intraplate stress variations (e.g., Transcontinental Arch and Nemaha High).

Intraplate stresses play a large role in the generation of craton sequences by reactivating structures and conducting subsidence and uplift (Burgess 2008). The normal faults generated in the development of the Araí paleorift have been reactivated during the development of the Paranoá basin, involving the generation of the epirogenic arches that controlled basin installation and further evolution.

The source of intraplate stresses in the Congo-São Francisco paleocontinent responsible for the Upper Veadeiros and Espinhaço intracratonic basin sequences in the upper Mesoproterozoic might be related to compressive forces acting in plate margins. According to Bogdanova et al. (2009), the Irumide Belt on the craton's southern margin provides evidence of an active continental margin during the time span of 1055–1020 Ma. That age matches the Paranoá Group 1042 Ma. diagenesis age (Matteini et al. 2012).

4.2. Regional Correlation

In the São Francisco Craton eastern margin, three different rifting events are clearly recorded (ca. 1.73, 1.57 and 0.85 Ga). In the western margin, two rifting events are consensual at

ca. 1.77, and 1.57 Ga. (Pimentel et al. 1991), and other two are interpreted as being represented by the layered intrusions and related volcanism in the Goiás Massif, dated ca. 1.25 Ga. for the mafic-ultramafic upper layered intrusions and related volcanic sequences of Juscelândia and 0.79 Ga. for the lower layered intrusions (Ferreira Filho et al. 2010). However, the authors affirm that other chronocorrelated rifts are not described in the region and thus interpret the Goiás Massif, where the complexes occur, as an allochthonous block at the time of rifting.

In an opposing view, the Goiás Massif is demonstrated to have been accreted in the São Francisco Craton during the 2.1-2.0 Ga. Transamazonian Orogeny (Cordeiro and Oliveira 2017; Martins-Ferreira et al. 2017) by strong stratigraphic and geochronologic arguments. In that scenario, the rifting recorded by the complexes would have occurred in the São Francisco paleocontinent interior (See paleocontinent limit in figure 1) and could have a close relationship to the development of subsequent intracratonic basins. Besides the generation of crustal weaknesses by rift faulting, other mechanisms can induce subsidence over previously rifted crust.

The emplacement of great volumes of dense mafic-ultramafic rocks into the lower density granitic crust generates a density anomaly that may result, in time, in isostatic subsidence, through the mechanism of crustal densification, as predicted by Ingersol (2012).

Furthermore, there is in fact evidence, of a slightly diachronous extensional event recorded in mafic intrusive dikes at the Espinhaço Sequence in the 1.2-1.0 Ga. interval (K/Ar, Távora et al. 1967; Ar/Ar, Renne et al. 1990) which could be correlated to the 1.25 Ga. rifting in the Goiás Massif and strong evidence of a similar mafic magmatism at ca 1250 Ma. in the Congo Craton, recorded by the Post-Kibaran Intrusives (D'Agrella-Filho and Cordani 2017).

The existence of an ocean at the São Francisco paleocontinent western margin is consensual. It has been called Goiás Ocean and interpreted as an extension of the Pharusian Ocean, developed at the West African Craton eastern margin, comprising the Goiás-Pharusian Ocean.

The opening age of such oceanic basin, however, is still debatable and in fact all the four extensional pulses recorded in the São Francisco western margin could be viable candidates for having evolved to ocean spreading systems, leaving behind some aborted rift arms which we classify as intracontinental rifts. Theoretically, there is even the possibility that an ocean already existed at the western margin before these rifting events took place.

As reported by Condie (1992), there is strong evidence that the Pharusian ocean has opened at ca. 800 Ma. The event is well established by the emplacement of alkaline magmatism

associated to continental fragmentation, by an undoubted passive margin sequence and the existence of the Gourma trough, an aulacogen at high angle to the expected orientation of the Pharusian main rift.

If the assumption that the Goiás Ocean extended to the Pharusian ocean is correct, the ca. 790 Ma. mafic-ultramafic layered complexes that occur in the Goiás Massif are very likely the record of that ocean aperture. Extension-related mafic magmatism of this age is also recorded in the Congo Craton by the Gagwe Lavas ca. 790 Ma. and the Mbozi Mafic Complex ca. 750 Ma. (D'Agrella-Filho and Cordani 2017).

Turbidites are found present across the entire proto ocean floor of the Pharusian Ocean, evidence that it represented a "miniocean" (Kroner 1979). As a possible extension of this ocean, the 790 Ma. rifting might have also opened a small oceanic basin or even none at all.

The same can be true to the 1.25 Ga. event, despite of consensus about the generation of oceanic crust in this event, the models constructed so far propose a narrow oceanic basin that separated the Goiás Massif from the São Francisco paleocontinent at ca. 1.27-1.25 Ma. and closed between 750-630 Ma. (Moraes et al. 2003, 2006, Ferreira-Filho et al. 2010), a 500-640 Ma. interval, which is a too long timespan for an oceanic basin (187 Ma. mean life), specially a narrow one, unless it represented a dormant ocean, but no records of such basin type are found.

The model proposed by Moraes et al. (2006, Figure 5d) shows, at ca. 790 Ma., a narrow ocean opened ca. 1.27 Ga. separating the Goiás Massif from the São Francisco paleocontinent and a wide ocean behind the Goiás Massif hosting the intra-oceanic Goiás Magmatic arc (Figure 19). No mention is made, however, to the event that opened that wider ocean and the passive margin sequences related to it. In a contrasting view, De Waele et al. (2008) consider that the ocean separating the São Francisco-Congo and Amazonian cratons is not necessarily wide, or might have a complex history, since available paleomagnetic data do not exclude the possibility that these cratons have been near each other before 900–880 Ma and were separated again afterwards.

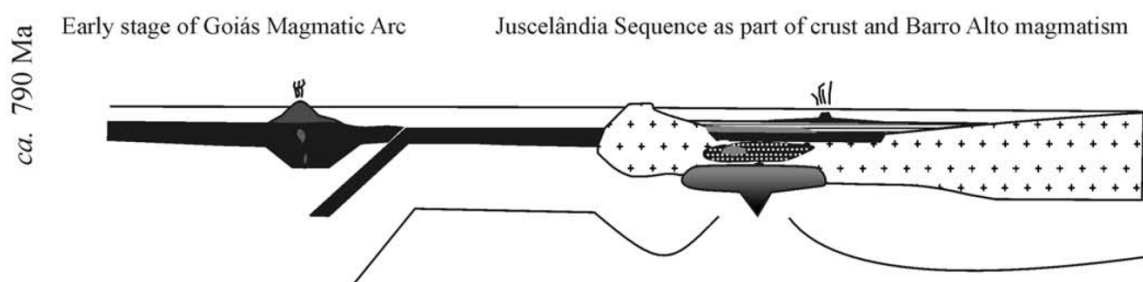


Figure 19 - Interpreted schematic section for the Juscelândia Rifting proposed by Moraes et al. (2006).

Rifting events records are slightly diachronous in the São Francisco Craton and its margins (Figure 20), especially in the western margin where distinct records of rift-related magmatism are present. Unfortunately, in addition to oceanic and transitional crusts consumption, the ocean closure and consequent continental collision during Neoproterozoic Brasiliano orogeny has imposed metamorphism from greenschist to granulite facies and a series of thrusts that have hindered the identification of ultimate evidences in favor of one of these events. It is possible, for instance, that a passive margin basin contemporary to the Paranoá basin is not preserved, or has not been identified yet, among the poorly mapped and understood metasedimentary sequences that occur over the Goiás Massif.

The ca. 1.25 Ga rifting is considered too old to represent the development of the passive margin in the western São Francisco Craton (Dardenne 2000; Pimentel et al. 2000). Pimentel et al. (2004) consider that the precursor rift of the western São Francisco Craton passive margin must have started around 1.0 Ga, as indicated by geochronological data of mafic dike swarms in the São Francisco and Congo cratons (Tack et al. 2001). The Paranoá diagenetic age ca. 1042 ± 22 Ma (Matteini et al. 2012) predates the most reasonable 1.0 Ga. assumption. Also, the Paranoá and its chronocorrelated sequences do not present any reliant characteristics that would allow their classification strictly as passive margin records, younger sequences are thus required to represent the São Francisco paleocontinent passive margin.

Furthermore, the recently proposed São Francisco paleocontinent limit at 1.8 Ga. (Cordeiro et al. 2017; Martins-Ferreira et al. 2017), coincides with a Neoproterozoic suture generated by the Goiás Magmatic Arc collage, occurred during Gondwana assembly. The Paranoá Group rocks outcrop at least 100 km far from that limit, which reinforces its classification as an intracontinental or continental margin basin.

Alternatively, younger sedimentary sequences in all São Francisco Craton margins show confident stratigraphic evidences of passive margin environment. These sequences are younger than the Paranoá Group and point to a ca. 1.0-0.8 Ga. ocean opening interpretation, episode known as Tonian Taphrogenesis (Alcântara et al. 2017). A basin-type classification based on tectonic setting is proposed for some of the sedimentary sequences in the São Francisco Craton and margins (Figure 20), constrained by the available data (Table 1) and guided by the presented literature review. Correlation is made with global events of glaciation and supercontinent cycles, as well as regional events of rifting and orogeny.

At the western margin, the Canastra Group is commonly interpreted as a lateral correlative of the Paranoá Group. However significant differences between the two support a

different interpretation. First, the Canastra group records shallowing upwards environment and the Paranoá records the opposite. Second, the youngest zircons ever found in the Paranoá are ca. 1.5 Ga. (Matteini et al. 2012, Martins-Ferreira et al. 2017) whereas the Canastra contain ca. 1.0 Ga. zircons (Valeriano et al. 2004a,b; Rodrigues 2008), allowing to suppose that the Canastra Group might be younger than the Paranoá. The Vazante Group is also younger, its detrital zircons point to a maximum depositional age of 935 ± 14 Ma (Rodrigues 2008; Rodrigues et al. 2012). Both Vazante and Canastra groups record turbidite sequences and syn-depositional mafic volcanic rocks (Dardenne 2000, Moura et al. 2016).

The Araxá and Canastra groups have been previously interpreted as preserved records of the Neoproterozoic passive margin that bordered the São Francisco paleocontinent western margin (Dardenne 2000, Valeriano et al. 2004a). The 1.0 to 0.9 zircons found in the Lower Araxá Group units, Lower Ibiá Group (Cubatão Formation), Canastra Group, and Vazante Group mark their maximum depositional age and these zircons might derive from rift-related magmatic events responsible for the opening of this oceanic basin. Mafic dike swarms spanning from 1.0 to 0.8 Ga. found in the São Francisco Craton corroborate for that hypothesis (Girardi et al. 2017). Many of these swarms record the onset of different rifting events related to a long-lasting regional extension phase (Figure 20), followed by development of passive margins in late Tonian-Cryogenian periods. Minimum sedimentation age for these passive margin sequences is given by regional metamorphism associated to subduction at ca. 650–610 Ma. (Valeriano et al. 2017, Caxito et al. 2017, Reis et al. 2017).

Younger units, such as the Bambuí Group, the Rio Verde Formation (Upper Ibiá group) and part of the Araxá Group, with detrital zircons as young as 0.6 Ga, show mixed sedimentary provenance, both from cratonic sources and presumably from magmatic arc terranes. The Bambuí sequence shows characteristics of Foreland basin environment, and is correlated to the Salitre Formation (Una Group) (Figure 20). The Rio Verde Formation records an active margin depositional environment, such as a fore-arc basin and, together with the Upper Araxá units, is correlated to the Salinas Formation (Upper Macaúbas Group) (Alkmim and Martins Neto 2012, Valeriano et al. 2017).

The initiation of the Goiás Magmatic Arc, indicated on figure 20 at ca. 900 Ma. is recorded by its oldest rocks formed in intra-oceanic island arc setting (Pimentel and Fuck 1992). The event marks the inversion from oceanic basin spreading to oceanic closure by crustal consumption and do not necessarily mean the installation of an active margin at the São Francisco Craton western margin at that time.

If the ca. 550 Ma. zircons in the Bambuí Group Sete Lagoas Formation (Pimentel et al. 2016) are confirmed not to have Pb loss, the Jequitaiá Glaciation does not necessarily need to be separated from it to match the older Gaskiers, Marinoan and Sturtian glaciations. There is evidence of a late Ediacaran (ca 547 Ma) glaciation in Namibia, Africa, called Baykonurian glaciation, that is also recorded in South America and Eurasia (Chumakov 2009, 2010, Germs and Gaucher, 2012) and could be related to the Jequitaiá Formation. It is very likely that Jequitaiá glacial centers were located in the craton (Martins Ferreira et al. 2013), where younger zircons were absent, explaining the non-occurrence of zircons younger than 880 Ma. in the Jequitaiá Formation.

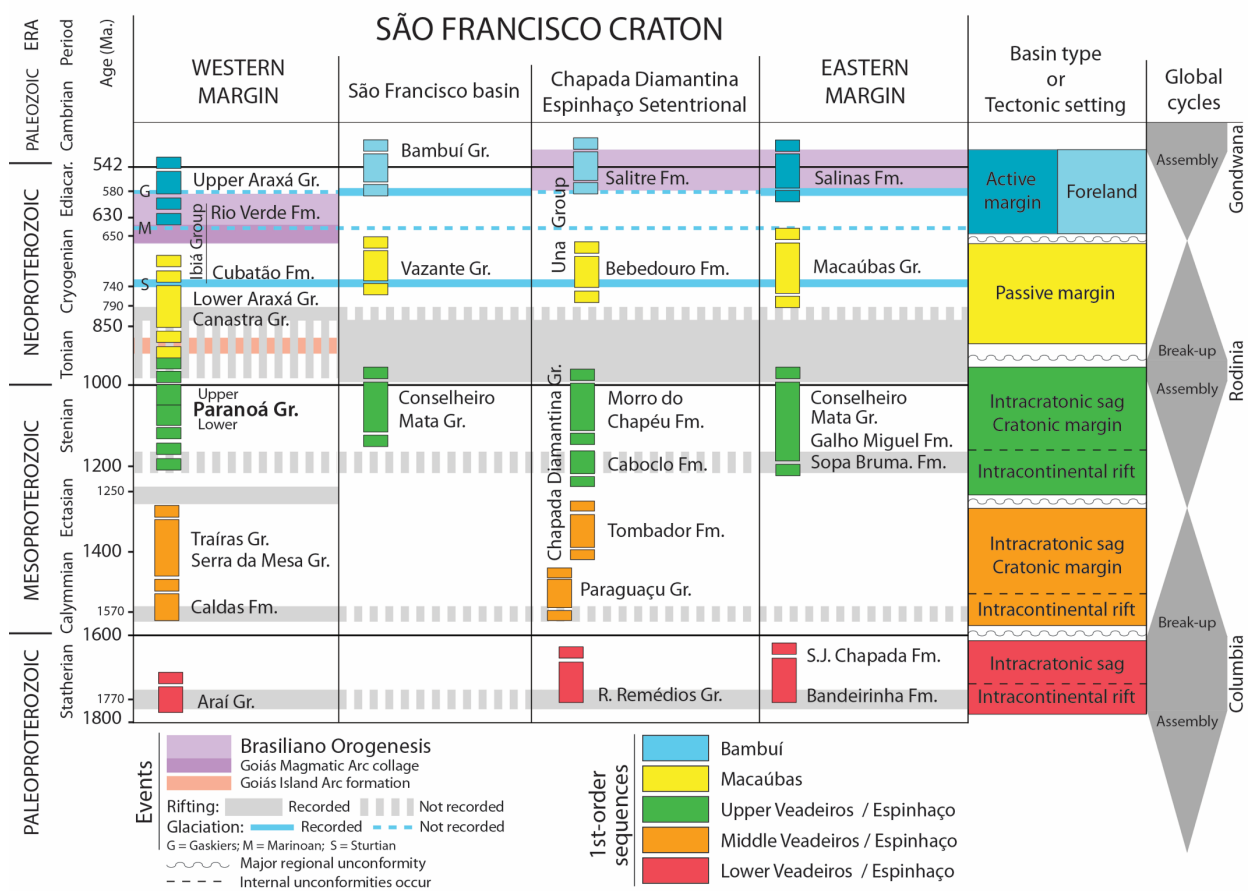


Figure 20 – Regional stratigraphic correlation chart of the Proterozoic São Francisco Craton and margins first-order sequences showing interpreted basin types, tectonic settings and events (modified from Alkmim and Martins-Neto 2012, Reis et al. 2017, with global age data acquired from Li et al. 2008, Bogdanova et al. 2009, Pisarevsky et al. 2003, Zhao et al. 2003, Johansson et al. 2014 and regional U-Pb age data compilation presented on table 1, correlation explanation in text).

Table 1 – Ages (U-Pb) and basin classification data compilation used for the regional correlation chart presented on figure 20. Major compilation source is derived from Heilbron et al. (2017), Pimentel (2016) and Ferreira Filho et al. (2010).

Unit	Max. Dep. Age (U-Pb)	Min. Dep. Age (U-Pb)	Environment or Basin Type	References
BambuÍ Group				
Três Marias Fm.	616; 575±9		Foreland	Rodrigues (2008); Kuchenbecker (2015)
Serra da Saudade Fm. Serra Sta. Helena Fm.	612; 622±7			Pimentel et al. (2001); Rodrigues (2008)
Sete Lagoas Fm.	557; 609			Paula-Santos et al. (2015) Rodrigues (2008)
Jequitai Formation	883		Glaciogenic	Rodrigues (2008)
Araxá Group				
Upper Araxá	607±1		Active margin	Valeriano et al. (2004a)
Lower Araxá	907±1		Passive Margin	Valeriano et al. (2004a)
Ibiá Group				
Rio Verde Fm.	636±21; 639±15		Foreland	Rodrigues (2008); Dias et al. (2011)
Cubatão Fm.	922 ± 16; 935±11		Passive Margin Glaciogenic	Dias et al. (2011); Rodrigues et al. (2010)
Canastra Group				
Chapada dos Pilões Fm.	971±98; 1070		Passive Margin	Dias (2011); Rodrigues et al. (2010)
Paracatu Fm.	1037±76; 1063±30;			Dias (2011); Rodrigues et al. (2010)
Serra do Landim Fm.	1079±45			Rodrigues et al. (2010)
Passos nappe	1011; 1030			Valeriano et al. (2004a, b); Rodrigues (2008)
Vazante Group				
Santo Antonio do Bonito Retiro Formation	997±29		Passive margin	Rodrigues et al. (2012);
Rocinha Formation	935±14	650–610		Rodrigues et al. (2012); Valeriano et al. (2017)
Serra do Garrote Fm.	1296±13			Rodrigues et al. (2012);
Morro do Calário Fm.	1137±8			Rodrigues et al. (2012);
Lapa Formation	1048±14			Rodrigues et al. (2012);
Paranoá Group				
			Cratonic Margin Intracratonic Sag	This study
Ribeirão do Torto Fm.	1558±13			Matteini et al. (2012)
Ribeirão Piçarrão Fm.	1610±11			Martins-Ferreira et al. (2017)
Ribeirão Contagem Fm.	1553±8			Matteini et al. (2012)
Serra do Paranã Fm.	1507±17			Martins-Ferreira et al. (2017)
Serra da Boa Vista Fm.	1540±10	1042±22 (Dep. Age)		Matteini et al. (2012)
São Miguel Formation	1428±12			Martins-Ferreira et al. (2017)
Serra da Mesa Group				
	1557±24		Intracont. Sag	Marques (2009)
TraÍras Group				
			Rift-Sag or Intracont. Sag	Martins-Ferreira et al. (2017)
Rosário Formation	1543±31		Intracont. Sag	Martins-Ferreira et al. (2017)
AraÍ Group				
Arraias Formation	-	1771 (Dep. Age)	Intracont. Rift	Pimentel et al. (1991)
Aurumina Suíte				
	-	2110 (Cryst.)	Syn-Orogenic	Cuadros et al. (2017a)
Ticunzal Formation	-	2160	Intracratonic Sag	Cuadros et al. (2017b)
Macaúbas Group				
Salinas Fm.	548	520	Active Margin	Kuchenbecker (2014)
Upper Macaúbas	-	-	Passive Margin	Heilbron et al. (2017)
Middle Macaúbas	-	-	Rift - Glacial Marine	Heilbron et al. (2017)
Lower Macaúbas	905; 875		Rift	Machado et al. (1989); Silva et al. (2008); Menezes et al. (2012)
Una Group				
Salitre Formation	-	-	Foreland	Pedreira et al. (1975)

Bebedouro Formation	-	-	Glaciogenic Margin	Passive	Guimarães (1996)
Conselheiro Mata Group	1417±22	906±2	Intracratonic Sag		Chemale et al. (2012); Abreu (1991)
Córrego Pereira Formation	1150±19	933±20	Intracratonic Sag		Kuchenbecker (2015); Dussin and Chemale (2012)
Galho do Miguel Fm.	1350	906±2	Intracratonic Sag		Lopes (2012); Abreu (1991)
Sopa-Brumadinho Fm.	1192	906±2	Intracratonic Sag		Guadagnin et al. (2013)
Tombador Formation	1415±41	934±14	Intracratonic Sag		Guadagnin and Chemale (2015); Loureiro et al. (2008)
São João da Chapada Fm.		1703±12	Intracratonic Sag		Chemale et al. (2012)
Bandeirinha Formation		1703±12	Intracont. Rift		Chemale et al. (2012)
Rio dos Remédios Group		1750	Intracont.Rift-Sag		Danderfer Filho et al. (2009)
Eastern Margin and SFC - Mafic Dikes		890±130; 934±14; 854±23 (Crystallization)	Extensional Event		Guadagnin et al. (2013); Loureiro et al. (2008); Danderfer Filho et al. (2009)
Western Margin - Upper Layered Intrusions		1250 (Crystalliz.)	Extensional Event		Ferreira Filho et al. (2010)
Western Margin - Lower Layered Intrusions		790 (Crystalliz.)	Extensional Event		Ferreira Filho et al. (2010)

4.3. Global Correlation to Supercontinent Cycles

The initiation of basins is not uniformly or randomly distributed in geological time, but they tend to initiate at times of break-up of supercontinents (Armitage and Allen 2012), except those basins that form by intraplate stress as some intracratonic basins, which tend to form during supercontinent assembly. However, correlation to supercontinent cycles is rarely straight forward, assembly and break-up events generally last very long periods, are not synchronous in all cratons, and most of them, especially Rodinia, are still controversial in both ages and cratonic participation (e.g., Pisarevsky et al. 2003; Hoffman 1991, Kröner & Cordani 2003, Cordani et al. 2003, Li et al. 2008, Bogdanova et al. 2009, Evans 2009, 2016, De Waele et al. 2008, Johansson et al. 2014).

As mentioned above, the conjugate São Francisco-Congo craton's position and participation in these supercontinents is still very debatable, except for Gondwana (assembly ca. 650-500 Ma.) and Pangea (Veevers 2004). Nonetheless, it was possible to establish an approximate timing of supercontinent break-up phases with the São Francisco paleocontinent regional widespread extension phases that generated intracontinental rifts, intracratonic sag basins and passive margin basins. On the other hand, supercontinent assembly can be related to the formation of major unconformities, intracratonic, active margin and foreland basins.

Tectonic events responsible for intraplate stress in the São Francisco paleocontinent may be related to the amalgamation of Rodinia. In a paleomagnetic study on Rodinia, Bogdanova et al. (2009) state that the major colliding events that resulted in the emergence of the stable continental structure called Rodinia occurred between 1050 and 900 Ma, which coincides with

the most probable age of installation and evolution of the Paranoá basin. The authors affirm, however, that the data about the Congo-São Francisco Craton in Rodinia are controversial due to the lack of paleomagnetic information and also to the occurrence of the Irumide Belt, that shows evidence of active continental margin in the south of the Congo Craton between 1055 and 1020 Ma, contradicting the involvement of the craton in the period of major Rodinia collisions. They conclude, however, that paleomagnetic data do not exclude the possibility that the Congo-São Francisco Craton was close to the Amazon Craton before ca. 900-880 Ma, and after that age they would have separated again.

Global peaks of juvenile crustal growth show a distinct peak around 1.2 Ga (Condie, 1998), which coincide with the ca. 1.25 Ga. upper layered intrusions and basaltic volcanism sequences in the São Francisco Craton western margin and also to the Rodinia main collisions. The assembly of Rodinia, which comprised most of Precambrian continental blocks, lasted ca. 400 m.y. (from 1300 to 900 Ma), encompassing the Paranoá basin lifespan. Its break-up took place between 830 and 650 Ma. most probably by mantle plume interaction (Bogdanova et al. 2009), corresponding to the age interval proposed for the passive margin sequences in the São Francisco Craton.

Pisarevsky et al. (2003) proposes a model the Rodinia break-up starts ca. 820-800 and end at ca. 780-770 and that the break-up might have been accompanied by mantle plume. They report that extensive mafic magmatism related to rifting is recognized in western Laurentia, SE Australia and South China and in western Laurentia at 730-780 Ma. and at 820-830 Ma. in SE Australia and South China. That magmatism can be related to the 790 Ma. rifting and mafic intrusive and volcanic sequences found in the São Francisco Palecontinent (Ferreira Filho et al. 2010).

The problem with correlating São Francisco craton tectonic events to Rodinia is that the position and timing constraints of São Francisco- Congo craton in the supercontinent has always been a challenge. Many authors have attempted to include it in new models without success (Pisarevsky et al. 2003; Evans 2009, 2016; Bogdanova et al. 2009, etc.), some have even proposed to exclude the pair from the models altogether (Kröner & Cordani 2003; Cordani et al. 2003). Some models are specifically built intending to unravel that mystery (De Waele et al. 2008).

Nonetheless, continents that did not participate in Rodinia might also have been affected by the intense mantle plume activity during the period of Rodinia break-up, explaining the several occurrences of mafic magmatism at ca. 900-750 Ma. in the São Francisco-Congo craton.

Johansson et al. 2014 propose the SAMBA model, in which the São Francisco-Congo craton is part of Rodinia but proposes a back-arc setting ca. 850-750 Ma. at the western São Francisco Craton margin, which is by all means consistent with our stratigraphic records.

According to Zhao et al. (2003), the assembly of Columbia supercontinent was completed during the global-scale 2.1-1.8 Ga collisional event, forming the 2.1-2.0 Ga Transamazonian Orogen in South America and its break-up commenced at ca.1.6 Ga and continued until ca. 1.2 Ga, as indicated by widespread 1.6-1.2 Ga continental rifting, anorogenic magmatism and emplacement of mafic dyke swarms in all cratonic blocks of Columbia. That event can be correlated to the development of Statherian Lower Veadeiros/Espinhaço and Callymian Middle Veadeiros/Espinhaço sequences.

5. Conclusions

Understanding the controls for a Proterozoic basin installation and evolution can be challenging in a thrust belt environment due to deformation and metamorphism. However, the use of gravimetric data and detailed stratigraphic surveys, provided new evidences that, allied with literature review, allowed to step forward in the direction of a more comprehensive view of the tectonic settings and events taking place during the Paranoá basin evolution in the São Francisco paleocontinent during the Mesoproterozoic.

The regional and global correlation of tectonic settings and events proposed is constrained by a comprehensive compilation of the most recent data available.

The results obtained in this study, together with well established concepts from reliable literature sources, have led to the following conclusions or suppositions:

- The Serra do Cristal Fault system has acted during Paranoá basin installation and further evolution, controlling basement highs and lows. It was later reactivated as reverse faults during the Brasiliano basin inversion.
- The development of epirogenic arcs acted as confining blocks during the Paranoá basin evolution and were likely favored by the reactivation of preexisting Araí Rift structures underneath, leading to the initial development of the Serra do Cristal Fault system.
- Despite being dominated by faulting and by basement blocks reworking, the initial phase of the Paranoá basin evolution should not be characterized as a rift system in the strict sense. The assumption is supported by the stratigraphic record, that indicates rapid transition to a fault-absent subsidence mechanism, evidenced by the lack of

repeated production of coarse conglomerates and the great stratal lateral continuity that follows the basal São Miguel Conglomerate.

- The Paranoá Group sequence outcropping at the western margin of the São Francisco paleocontinent unlikely represents a passive margin record in the strict sense. Alternatively, it presents clear stratigraphic and tectonic characteristics that allow its classification as a cratonic margin basin that might have acted as an epicratonic embayment basin in times of high stand sea level.
- If analyzed in all its extension, the Paranoá sequence can be characterized as different types of basins: continental platform basin over the São Francisco Craton, intracratonic sag over ancient rifts that affect the craton, and cratonic margin type basin over the younger, thinner, more recently accreted continental crust that formed the margins of São Francisco paleocontinent. Nowhere, thus, the Paranoá sequence discloses passive margin characteristics, at least in the preserved sequences.
- More suitable candidates to be characterized as passive margin sequences in the São Francisco paleocontinent western margin are the Canastra, Vazante, Lower Araxá and Lower Ibiá (Cubatão Formation) groups, which can be correlated to the lower Macaúbas and lower Una (Bebedouro Formation) groups in a regional context.
- As indicated by the literature review data, we interpret the 1.0-0.8 Ga. interval, known as Tonian Taphrogenesis, most likely represents the extensional event that led to passive margin development in the São Francisco paleocontinent western margin.

Acknowledgements

The authors are grateful to the University of Brasília (*UnB*) and the Brasília Seismological Observatory (*ObsIS*) for the laboratorial and financial support.

References

- Abtout, A.; Boukerbout, H.; Bouyahiaoui, B.; Derder, M. 2015. Gravimetric Evidences of Active Fault and Underground Structure of Chelif Seismogenic Basin (Algeria). *Journal of African Earth Sciences*. Volume 99, Part 2, November 2014, Pages 363–373.
- Alcântara, D.D.C.B.G., Uhlein, A., Caxito, F.D.A., Dussin, I., Pedrosa-Soares, A.C. 2017. Stratigraphy, tectonics and detrital zircon U-Pb (LA-ICP-MS) geochronology of the Rio Preto Belt and northern Paramirim corridor, NE, Brazil. *Brazilian Journal of Geology*, 47(2), 261-273.

- Alkmim, F.F.; Martins-Neto, M.A. 2012. Proterozoic first-order sedimentary sequences of the São Francisco craton, eastern Brazil. *Marine and Petroleum Geology*, 33(1), 127-139.
- Allen, P.A., Armitage, J.J. 2012. Cratonic basins. *Tectonics of Sedimentary Basins: Recent Advances*, 602-620.
- Allen, P. A., & Allen, J. R. 2013. *Basin analysis: Principles and application to petroleum play assessment*. John Wiley & Sons.
- Alvarenga, C.J.S., Dardenne, M.A., Vieira, L.C., Martinho, C.T., Guimarães, E.M., Santos, R.V., Santana, R.O. 2012. Estratigrafia da Borda Ocidental da Bacia do São Francisco. *Boletim de Geociências da Petrobrás* 20, 145–164.
- Armitage, J.J., Allen, P.A. 2010. Cratonic basins and the long-term subsidence history of continental interiors. *Journal of the Geological Society*, 167(1), 61-70.
- Babinski, M., Van Schmus, W.R., Chemale, F., Brito-Neves, B.B., Rocha, A.J.D. 1993. Idade isocrônica Pb/Pb em rochas carbonáticas da Formação Caboclo. Morro do Chapéu, BA. In: SBG, Simp. do Cráton do São Francisco, 2, 160-163.
- Bertoni, M.E., Rooney, A.D., Selby, D., Alkmim, F.F., Le Heron, D.P. 2014. Neoproterozoic Re–Os systematics of organic-rich rocks in the São Francisco Basin, Brazil and implications for hydrocarbon exploration. *Precambrian Research*, 255, 355-366.
- Bogdanova, S.V., Pisarevsky, S.A., Li, Z.X. 2009. Assembly and breakup of Rodinia (some results of IGCP Project 440). *Stratigraphy and Geological Correlation*, 17(3), 259.
- Boyce, J.I.; Morris, W.A. 2002. Basement-controlled faulting of Paleozoic strata in southern Ontario, Canada: new evidence from geophysical lineament mapping, *Tectonophysics*, Volume 353, Issues 1–4, 23 August 2002, Pages 151-171.
- Burgess, P.M. 2008. The nature of shallow-water carbonate lithofacies thickness distributions. *Geology*, 36(3), 235-238.
- Campos, J.E.G., Dardenne, M.A., Freitas-Silva, F.H., Martins-Ferreira, M.A.C. 2013. Geologia do Grupo Paranoá na porção externa da Faixa Brasília. *Brazilian Journal of Geology*, 43(3), 461-476.
- Catuneanu, O.; Martins-Neto, M.; Eriksson, P.G. 2005. Precambrian sequence stratigraphy. *Sedimentary Geology*, 176 (1-2), 67-95.
- Cawood, P.A.; Nemchin, A.A.; Strachan, R.; Prave, T.; Krabbendam, M. 2007. Sedimentary basin and detrital zircon record along East Laurentia and Baltica during assembly and break-up of Rodinia. *Journal of the Geological Society*, 164(2), 257-275.

- Cawood, P.A.; Pisarevsky, S.A. 2017. Laurentia-Baltica-Amaozonia relations during Rodinia assembly. *Precambrian Research*, 292, 386-397.
- Chakravarthi, V. and Ramamma, B. 2015. Determination of Sedimentary Basin Basement Depth: A Space Domain Based Gravity Inversion using Exponential Density Function, *Acta Geophysica*. 63, 1066-1079.
- Chemale, F., Dussin, I. A., Alkmim, F. F., Martins, M. S., Queiroga, G., Armstrong, R., Santos, M. N. 2012. Unravelling a Proterozoic basin history through detrital zircon geochronology: the case of the Espinhaço Supergroup, Minas Gerais, Brazil. *Gondwana Research*, 22(1), 200-206.
- Chumakov, N.M., 2009a. The Baykonurian glaciohorizon of the Late Vendian. *Stratigraphy and Geological correlations* 17, 373-383.
- Chumakov, N.M. 2010. Precambrian glaciations and associated biospheric events. *Stratigraphy and Geological correlations* 18, 467-479.
- Cloud, P. and Dardenne, M. 1973. Proterozoic age of the Bambuí Group in Brazil. *Geological Society of America Bulletin*, 84(5), 1673-1676.
- Coleman, J.L.Jr.; Cahan, S.M., 2012, Preliminary catalog of the sedimentary basins of the United States: U.S. Geological Survey Open-File Report 2012-1111, 27 p. (plus 4 figures and 1 table available as separate files). Available online at <http://pubs.usgs.gov/of/2012/1111/>.
- Condie, K.C. 1998. Episodic continental growth and supercontinents: a mantle avalanche connection? *Earth and Planetary Science Letters*, 163(1), 97-108.
- Cooper, M.A., Williams, G.D. (Eds.). (1989). *Inversion tectonics*. Cordani, U.G.; Teixeira, W.; D'Agrella-Filho, M.S.; Trindade, R.I. 2009. The position of the Amazonian Craton in supercontinents. *Gondwana Research*, 15(3), 396-407.
- Cordani, U.G., D'Agrella-Filho, M.S., Brito-Neves, B.D., Trindade, R. I. F. 2003. Tearing up Rodinia: the Neoproterozoic palaeogeography of South American cratonic fragments. *Terra Nova*, 15(5), 350-359.
- Cruz, E.L.C.C., Kuyumjian, R.M., Boaventura, G.R. 2003. Low-K calc-alkaline granitic series of southeastern Tocantins State: Chemical evidence for two sources for the granite-gneissic complexes in the Paleoproterozoic Almas-Dianópolis Terrane. *Brazilian Journal of Geology*, 33(2), 125-136.
- Cuadros, F.A., Botelho, N.F., Fuck, R.A., Dantas, E.L. 2017. The peraluminous Aurumina Granite Suite in central Brazil: an example of mantle-continental crust interaction in a Paleoproterozoic cordilleran hinterland setting? *Precambrian Research*, 299, pg.75-100.

- D'Agrella-Filho, M.S., Cordani, U.G. 2017. The Paleomagnetic Record of the São Francisco-Congo Craton. In São Francisco Craton, Eastern Brazil (pp. 305-320). Springer International Publishing.
- Dalziel, I.W.D. 1991. Pacific margins of Laurentia and East Antarctica–Australia as a conjugate rift pair: Evidence and implications for an Eocambrian supercontinent, *Geology* 19, 598 – 601.
- Danderfer Filho, A., De Waele, B., Pedreira, A.J., Nalini, H.A. 2009. New geochronological constraints on the geological evolution of Espinhaço basin within the São Francisco Craton—Brazil. *Precambrian Research*, 170(1), 116-128.
- Dardenne, M.A., Walde, D.H.G. 1979. A estratigrafia dos Grupos Bambuí e Macaúbas no Brasil central. 1º Simpósio de Geologia de Minas, Sociedade Brasileira de Geologia, Belo Horizonte, 43-54.
- Dardenne, M.A. 2000. The Brasília Fold Belt. In: U.G. Cordani, E.J. Milani, A. Thomaz Filho & D.A. Campos. 2000. Tectonic evolution of South America. 31 st International Geological Congress, Rio de Janeiro, SBG. p. 231-263.
- D'el-Rey Silva, L.J.H.; Vasconcelos, M.A.R.; Silva, D.V.G. 2008. Timing and role of the Maranhão River Thrust in the evolution of the neoproterozoic Brasília Belt and Tocantins Province, central Brazil. *Gondwana Research*, 13: 352-374.
- De Waele, B., Johnson, S.P., Pisarevsky, S.A. 2008. Palaeoproterozoic to Neoproterozoic growth and evolution of the eastern Congo Craton: its role in the Rodinia puzzle. *Precambrian Research*, 160(1),127-141.
- Dickinson, W.R. 1976. Sedimentary basins developed during evolution of Mesozoic–Cenozoic arc–trench system in western North America. *Canadian Journal of Earth Sciences*, 13(9), 1268-1287.
- Døssing, A.; Hansen, T.M.; Olesen, A.V.; Hopper, J.R.; Funck, T. 2014. Gravity inversion predicts the nature of the Amundsen Basin and its continental borderlands near Greenland, *Earth and Planetary Science Letters*, Volume 408, Pages 132-145
- Engen, Ø., Frazer, L., Wessel, P., Faleide, J., 2006. Prediction of sediment thickness in the Norwegian–Greenland Sea from gravity inversion. *J. Geophys. Res.* 111 (B11), B11403.
- Erickson, S.G. 1993. Sedimentary loading, lithospheric flexure, and subduction initiation at passive margins. *Geology*, 21(2), 125-128.
- Eriksson, P.G., Catuneanu, O., Bumby, A.J. 2012. First-and second-order global sequence stratigraphic correlations and accommodation charts for the Kaapvaal, Karelian, São Francisco (-Congo) and Slave cratons: An introduction. *Marine and Petroleum Geology*, 33(1), 1-7.

- Evans D.A.D., 2009, The palaeomagnetically viable, long-lived and all-inclusive Rodinia supercontinent reconstruction, in Murphy J.B., et al., eds., *Ancient orogens and modern analogues: Geological Society of London Special Publication 327*, p. 371–404, doi:10.1144/SP327.16.
- Evans, D.A.D.; Trindade, R.I.F.; Catelani, E.L.; D'Agrella-Filho, M.S.; Heaman, L. M.; Oliveira, E. P.; Salminen, J. M. 2016. Return to Rodinia? Moderate to high palaeolatitude of the São Francisco/Congo craton at 920 Ma. *Geological Society, London, Special Publications*, 424(1), 167-190.
- Faria, A.D. 1995. Estratigrafia e sistemas deposicionais do Grupo Paranoá nas áreas de Cristalina, Distrito Federal e São João D'Aliança-Alto Paraíso de Goiás. Unpublished PhD thesis, Universidade de Brasília, 199.
- Ferreira Filho, C. F., Pimentel, M. M., de Araujo, S. M., Laux, J. H. (2010). Layered intrusions and volcanic sequences in Central Brazil: geological and geochronological constraints for Mesoproterozoic (1.25 Ga) and Neoproterozoic (0.79 Ga) igneous associations. *Precambrian Research*, 183(3), 617-634.
- Fuck R.A., Marine J.O., Dardenne M.A. Figueiredo A.N. 1988. Coberturas metassedimentares do Proterozóico Médio: os grupos Araí e Paranoá na região de Niquelândia - Colinas, Goiás. *Revista Brasileira de Geociências*, 18(1):54-62
- Fuck, R.A., Brito-Neves, B.B., Schobbenhaus, C. 2008. Rodinia descendants in South America. *Precambrian Research*, 160(1), 108-126.
- Fuck, R.A., Marini, O.J., Dardenne, M.A., De Figueiredo, A. N. 2017. Coberturas metassedimentares do Proterozóico Médio: os grupos Araí e Paranoá na região de Niquelândia-Colinas, Goiás. *Revista Brasileira de Geociências*, 18(1), 54-62.
- Gerard, A., Griveau, P., 1972. Interprétations quantitatives en gravimétrie et magnétisme à partir de cartes transformées de gradient vertical. *Geophys. Prospect.* 20, 459–481.
- Germis, G.J.B.; Gaucher, C. 2012. Nature and extent of a late Ediacaran (ca. 547 Ma) glacigenic erosion surface in southern Africa. *South African Journal of Geology*, 115(1), 91-102.
- Gibson, R.I., & Millegan, P.S. (Eds.). 1998. *Geologic applications of gravity and magnetics: Case histories*. Society of Exploration Geophysicists and American Association of Petroleum Geologists.
- Girardi, V.A., Teixeira, W., Mazzucchelli, M., de Oliveira, E.P., da Costa, P.C.C. 2017. Mafic Dykes: Petrogenesis and Tectonic Inferences. In *São Francisco Craton, Eastern Brazil* (pp. 145-169). Springer International Publishing.
- Grotzinger, J.P., Bowring, S.A., Saylor, B.Z. and Kaufman, A.J., 1995. Biostratigraphic and geochronologic constraints on early animal evolution. *Science*, 270, 598–604.

- Guadagnin, F. and Chemale, F. 2015. Detrital zircon record of the Paleoproterozoic to Mesoproterozoic cratonic basins in the São Francisco Craton. *Journal of South American Earth Sciences*, 60, 104-116.
- Guimarães J.T. 1996. A Formação Bebedouro no Estado da Bahia: Faciologia, estratigrafia e ambientes de sedimentação. Dissertação de Mestrado, Instituto de Geociências, Universidade Federal da Bahia, 146p
- Gurnis, M. 1990. Bounds on global dynamic topography from Phanerozoic flooding of continental platforms. *Nature*, 344(6268), 754.
- Hartley, R.W., Allen, P.A. 1994. Interior cratonic basins of Africa: relation to continental break-up and role of mantle convection. *Basin Research*, 6(2-3), 95-113.
- Heilbron, M., Cordani, U.G., Alkmim, F. F. (Eds.). 2017. São Francisco Craton, Eastern Brazil: Tectonic Genealogy of a Miniature Continent. Springer.
- Hoffman, P.F. 1991. Did the breakout of Laurentia turn Gondwanaland inside-out. *Science*, 252(5011), 1409-1412.
- Howell, P.D., van der Pluijm, B.A. 1999. Structural sequences and styles of subsidence in the Michigan basin. *Geological Society of America Bulletin*, 111(7), 974-991.
- Busby, C. J., Ingersol, R. V. 1995. *Tectonics of Sedimentary Basins*, 579 pp.
- Ingersol, R. V. 2012. Composition of modern sand and Cretaceous sandstone derived from the Sierra Nevada, California, USA, with implications for Cenozoic and Mesozoic uplift and dissection. *Sedimentary Geology*, 280, 195-207.
- Johansson, Å. 2014. From Rodinia to Gondwana with the ‘SAMBA’ model—a distant view from Baltica towards Amazonia and beyond. *Precambrian Research*, 244, 226-235.
- Kröner, A. 1979. Pan African plate tectonics and its repercussions on the crust of northeast Africa. *Geologische Rundschau*, 68(2), 565-583.
- Kröner, A., Cordani, U. 2003. African, southern Indian and South American cratons were not part of the Rodinia supercontinent: evidence from field relationships and geochronology. *Tectonophysics*, 375(1), 325-352.
- Kuchenbecker, M. 2014. Relações entre coberturas do Cráton do São Francisco e bacias situadas em orógenos marginais: o registro de datações U-Pb de grãos detríticos de zircão e suas implicações geotectônicas. Universidade Federal de Minas Gerais, Brazil www.bibliotecadigital.ufmg.br.
- Li, Z.X., Bogdanova, S.V., Collins, A.S., Davidson, A., De Waele, B., Ernst, R.E., Fitzsimons, I.C.W., Fuck, R.A., Gladkochub, D.P., Jacobs, J., Karlstrom, K.E., Lu, S., Natapov, L.M., Pease, V.,

- Pisarevsky, S.A., Thrane, K., Vernikovsky, V., 2008. Assembly, configuration, and break-up history of Rodinia: A synthesis. *Precambrian Res.* 160, 179–210.
- Lopes, T.C. 2012. O Supergrupo Espinhaço na Serra do Cabral, Minas Gerais: contribuição ao estudo de proveniência sedimentar. Master's Dissertation. Universidade Federal de Minas Gerais. 136p.
- Machado N., Schrank A., Abreu F.R., Knauer L.G., Almeida-Abreu P.A. 1989. Resultados preliminares da geocronologia U/Pb na Serra do Espinhaço Meridional. In: SBG, Simp. Geol. MG., 5, Anais, p. 1-4
- Marquardt, D.W., 1963, An algorithm for least-squares estimation of nonlinear parameters: *J. Soc. of Ind. and Applied Math.*, 2, 601-612.
- Marques, G.C. (2009). Geologia dos grupos Araí e Serra da Mesa e seu embasamento no sul do Tocantins. Master's Thesis. Universidade de Brasília. Instituto de Geociências.
- Marshall, J. S., Idleman, B. D., Gardner, T. W., Fisher, D. M. 2003. Landscape evolution within a retreating volcanic arc, Costa Rica, Central America. *Geology*, 31(5), 419-422.
- Martins-Ferreira, M.A.; Chemale, F.; Dias, A.N.C.; Campos, J.E.G. 2017. Proterozoic intracontinental basin succession in the São Francisco Craton western margin constrained by detrital zircon geochronology. *Journal of South American Earth Sciences*. Submitted for Publication on the *Journal of South American Earth Sciences*.
- Martins-Neto, M.A.; Pedrosa Soares, A.C.; Lima, S.A.A. 2001. Tectono-sedimentary evolution of sedimentary basins from Late Paleoproterozoic to Late Neoproterozoic in the São Francisco craton and Araçuaí fold belt, eastern Brazil. *Sedimentary Geology* 141/142, 343e370.
- Matteini, M., Dantas, E. L., Pimentel, M. M., de Alvarenga, C. J. S., Dardenne, M. A. 2012. U–Pb and Hf isotope study on detrital zircons from the Paranoá Group, Brasília Belt Brazil: Constraints on depositional age at Mesoproterozoic–Neoproterozoic transition and tectono-magmatic events in the São Francisco craton. *Precambrian Research*, 206, 168-181.
- McKenzie, D. 1978. Some remarks on the development of sedimentary basins. *Earth and Planetary science letters*, 40(1), 25-32.
- Moore, E.M. 1991. Southwest U.S.–East Antarctica (SWEAT) connection: a hypothesis, *Geology* 19, 425 – 428.
- Moraes, R., Fuck, R.A., Pimentel, M.M., Gioia, S.M.C.L., Figueiredo, A.M.G. 2003. Geochemistry and Sm–Nd isotopic characteristics of bimodal volcanic rocks of Juscelândia, Goiás, Brazil: Mesoproterozoic transition from continental rift to ocean basin. *Precambrian Research*, 125(3), 317-336.

- Moraes, R., Fuck, R.A., Pimentel, M.M., Gioia, S.M., de Hollanda, M.H., Armstrong, R. 2006. The bimodal rift-related Juscelândia volcanosedimentary sequence in central Brazil: Mesoproterozoic extension and Neoproterozoic metamorphism. *Journal of South American Earth Sciences*, 20(4), 287-301.
- Moura, F.G.; Martins-Ferreira M.AC.; Campos J.E.G. 2016. Petrografia e caracterização geoquímica preliminar de rochas vulcânicas do topo do Grupo Vazante. Conference Paper. 48° Brazilian Geological Congress, Porto Alegre - RS, Brazil, Volume: 1.
- Nikishin, A.M., Ziegler, P.A., Stephenson, R.A., Cloetingh, S.A.P.L., Furne, A.V., Fokin, P.A., ... & Gorbachev, V.I. 1996. Late Precambrian to Triassic history of the East European Craton: dynamics of sedimentary basin evolution. *Tectonophysics*, 268(1-4), 23-63.
- Pallero, J.L.G.; J.L. Fernández-Martínez, J.L.; S. Bonvalot, S.; Fudym, O. 2015. Gravity inversion and uncertainty assessment of basement relief via Particle Swarm Optimization. *Journal of Applied Geophysics*. Volume 116, May 2015, Pages 180–191.
- Paula-Santos, G.M., Babinski, M., Kuchenbecker, M., Caetano-Filho, S., Trindade, R.I., Pedrosa-Soares, A. C. 2015. New evidence of an Ediacaran age for the Bambuí Group in southern São Francisco craton (eastern Brazil) from zircon U–Pb data and isotope chemostratigraphy. *Gondwana Research*, 28(2), 702-720.
- Pedreira, A.J.; Arcanjo, J.B.A.; Pedrosa, C.J.; Oliveira, J.E.; Silva, B.C.E. 1975. Projeto Bahia – Geologia da Bacia de Rio de Contas. Salvador, DNPM/CPRM. 112 p. (Relatório 3).
- Pimentel, M. M., Heaman, L., Fuck, R. A., Marini, O. J. 1991. U-Pb zircon geochronology of Precambrian tin-bearing continental-type acid magmatism in central Brazil. *Precambrian Research*, 52(3), 321-335.
- Pimentel, M.M., Fuck, R.A. 1992. Neoproterozoic crustal accretion in central Brazil. *Geology*, 20(4), 375-379.
- Pimentel, M.M., Dardenne, M., Fuck, R., Viana, M., Junges, S., Fischel, D., Dantas, E.L. 2001. Nd isotopes and the provenance of detrital sediments of the Neoproterozoic Brasília Belt, central Brazil. *Journal of South American Earth Sciences*, 14(6), 571-585.
- Pimentel, M.M., Jost, H., Fuck, R. A. 2004. O embasamento da Faixa Brasília e o arco magmático de Goiás. *Geologia do continente sul-americano: evolução da obra de Fernando Flávio Marques de Almeida*, 355-358.
- Pimentel, M.M. 2016. The tectonic evolution of the Neoproterozoic Brasília Belt, central Brazil: a geochronological and isotopic approach. *Brazilian Journal of Geology*, 46, 67-82.

- Pisarevsky, S.A., Wingate, M.T., Powell, C.M., Johnson, S., Evans, D.A. 2003. Models of Rodinia assembly and fragmentation. *Geological Society, London, Special Publications*, 206(1), 35-55.
- Piuzana, D., Pimentel, M.M., Fuck, R.A., Armstrong, R. 2003a. SHRIMP U–Pb and Sm–Nd data for the Araxá Group and associated magmatic rocks: constraints for the age of sedimentation and geodynamic context of the southern Brasília Belt, central Brazil. *Precambrian Research*, 125(1), 139-160.
- Piuzana, D., Pimentel, M.M., Fuck, R.A., Armstrong, R. 2003b. Neoproterozoic granulite facies metamorphism and coeval granitic magmatism in the Brasília Belt, Central Brazil: regional implications of new SHRIMP U–Pb and Sm–Nd data. *Precambrian Research*, 125(3), 245-273.
- Reis, H. L. and Alkmim, F. F. 2015. Anatomy of a basin-controlled foreland fold-thrust belt curve: The Três Marias salient, São Francisco basin, Brazil. *Marine and Petroleum Geology*, 66, 711-731.
- Reis, H.L., Alkmim, F.F., Fonseca, R.C., Nascimento, T.C., Suss, J.F., Prevatti, L.D. 2017. The São Francisco Basin. In *São Francisco Craton, Eastern Brazil*, 117-143. Springer International Publishing.
- Renne, P.R., Onstott, T.C., D’Agrella-Filho, M.S., Pacca, I.G., Teixeira, W., 1990. $^{40}\text{Ar}/^{39}\text{Ar}$ dating of 1.0–1.1 Ga magnetizations from the São Francisco and Kalahari cratons: tectonic implications for Pan-African and Brasiliano mobile belts. *Earth and Planetary Science Letters* 101, 349–366.
- Rodrigues, J.B. 2008. Proveniência de sedimentos dos grupos Canastra, Ibiá, Vazante e Bambuí – um estudo de zircões detríticos e Idades Modelo Sm–Nd. (PhD Thesis) Instituto de Geociências, Universidade de Brasília, Brasília, (128 pp.).
- Rodrigues, J.B., Pimentel, M.M., Buhn, B., Matteini, M., Dardenne, M.A., Alvarenga, C.J.S., Armstrong, R.A., 2012. Provenance of the Vazante Group: new U–Pb, Sm–Nd, Lu–Hf isotopic data and implications for the tectonic evolution of the Neoproterozoic Brasília Belt. *Gondwana Research* 21 (2–3), 439–450.
- Salem, A., Green, C., Campbell, S., Fairhead, J. D., Cascone, L., & Moorhead, L. 2013. Moho depth and sediment thickness estimation beneath the Red Sea derived from satellite and terrestrial gravity data. *Geophysics*, 78(5), G89-G101.
- Sanford, B. V. 1987. Paleozoic geology of the Hudson Platform. Pgs 483-505.
- Scheibner, C.; Blomeier, D.; Forke, H.; & Gesierich, K. 2015. From terrestrial to shallow-marine depositional environments: reconstruction of the depositional environments during the Late Carboniferous transgression of the Lomfjorden Trough in NE Spitsbergen (Malte Brunfjellet Formation). *Norwegian Journal of Geology*, 95(2), 127-152.

- Scisciani, V.; Agostini, S.; Calamita, F.; Pace, P.; Cilli, A.; Giori, I.; Paltrinieri, W. 2014. Positive inversion tectonics in foreland fold-and-thrust belts: A reappraisal of the Umbria–Marche Northern Apennines (Central Italy) by integrating geological and geophysical data, *Tectonophysics*, Volume 637, 10 December 2014, Pages 218-237
- Selley, R.C. 1997. The basins of Northwest Africa: structural evolution. *Sedimentary Basins of the World*, 3, 17-26.
- Selley, R.C.; Sonnenberg, S.A. 2014. *Elements of petroleum geology*. Academic Press.
- Selley, R.C. 2000. *Applied sedimentology*. Elsevier.
- Silva L.C., Pedrosa-Soares A.C., Teixeira L., Armstrong R. 2008. Tonian rift-related, A-type continental plutonism in the Araçuaí Orogen, eastern Brazil: new evidence for the breakup stage of the São Francisco Congo Palecontinent. *Gondwana Research*, 13:527-537.
- Silva, J.B.C., Medeiros, W.E. e Barbosa, V.C.F. 2001. Potential-field inversion: Choosing the appropriate technique to solve a geologic problem: *Geophysics*, 66 , 511-520.
- Sloss, L.L. 1988. Forty years of sequence stratigraphy. *Geological Society of America Bulletin*, 100(11), 1661-1665.
- Tack, L., Wingate, M.T.D., Liégeois, J.P., Fernandez-Alonso, M., Deblond, A. 2001. Early Neoproterozoic magmatism (1000–910 Ma) of the Zadinian and Mayumbian Groups (Bas-Congo): onset of Rodinia rifting at the western edge of the Congo craton. *Precambrian research*, 110(1), 277-306.
- Távora, F.I., Cordani, U.G., Kawashita, K., 1967. Determinações de idade K/Ar em rochas da região central da Bahia. In: *Anais 21º Congresso Brasileiro de Geologia*, Sociedade Brasileira de Geologia, Curitiba, pp. 214–224.
- Uhlein, A., Trompette, R., Egydio-Silva, M. 1995. Rifteamentos superpostos e tectônica de inversão na borda sudeste do Cráton do São Francisco. *Revista Geonomos*, 3(1).
- Uhlein, A., Fonseca, M.A., Seer, H.J., Dardenne, M.A. 2012. Tectônica da faixa de dobramentos Brasília–setores setentrional e meridional. *Revista Geonomos*, 2(20).
- Uhlein, A., Trompette, R., Alvarenga, C., 1999. Neoproterozoic glacial and gravitational sedimentation on a continental rifted margin: the Jequitai–Macaúbas sequence (Minas Gerais, Brazil). *Journal of South American Earth Sciences* 12, 435–451.
- Valeriano, C.M., Machado, N., Simonetti, A., Valladares, C.S., Seer, H.J., Simões, L.S.A., 2004. U–Pb geochronology of Southern Brasília belt (SE-Brazil): sedimentary provenance, Neoproterozoic orogeny and assembly of West Gondwana. *Precambrian Research* 130, 27–55.

- Valeriano, C.M., Pimentel, M.M., Heilbron, M., Almeida, J.C.H., Trouw, R.A.J., 2008. Tectonic evolution of the Brasília Belt, Central Brazil, and early assembly of Gondwana. In: Pankhurst, R.J., Trouw, R.A.J., Brito-Neves, B.B., De Wit, M.J. (Eds.), *West Gondwana: Pre-Cenozoic Correlations Across the South Atlantic Region*. Geological Society, London, Special Publications. 294, pp. 197–210.
- Valeriano, C.D.M., Teixeira, W., Heilbron, M., Simões, L.S.A. 2017. Southern Brasilia belt (SE Brazil): tectonic discontinuities, K-Ar data and evolution during the Neoproterozoic Brasiliano orogeny. *Revista Brasileira de Geociências*, 30(1), 195-199.
- Veevers, J.J. 2004. Gondwanaland from 650–500 Ma assembly through 320 Ma merger in Pangea to 185–100 Ma break-up: supercontinental tectonics via stratigraphy and radiometric dating. *Earth-Science Reviews*, 68(1), 1-132.
- Watts, A.B. 2012. Models for the evolution of passive margins. *Regional Geology and Tectonics: Phanerozoic Rift Systems and Sedimentary Basins*, 1, 33.
- Woodcock, N.H. 2004. Life span and fate of basins. *Geology*, 32(8), 685-688.
- Zoback, M.L. 1992. First-and second-order patterns of stress in the lithosphere: The World Stress Map Project. *Journal of Geophysical Research: Solid Earth*, 97(B8), 11703-11728.
- Zoback, M.D., Apel, R., Baumgärtner, J., Brudy, M., Emmermann, R., Engeser, B., Vernik, L. 1993. Upper-crustal strength inferred from stress measurements to 6 km depth in the KTB borehole. *Nature*, 365(6447), 633-635.

CAPÍTULO 6

Artigo 5:

Caracterização estrutural preliminar da Saliência do Moqué, sistema de dobramentos e cavalgamentos da zona externa da Faixa Brasília Norte

Caracterização estrutural preliminar da Saliência do Moqué, sistema de dobramentos e cavalgamentos da zona externa da Faixa Brasília Norte

Marco Antônio Caçador Martins-Ferreira¹ marcoc@unb.br (Corresponding Author); José Eloi Guimarães Campos¹

1. Instituto de Geociências, Universidade de Brasília, Campus Universitário Darcy Ribeiro, Asa Norte, CEP 70.910-900, Brasília - DF - Brasil.

Abstract

Salients are local or regional orogenic curvatures that represent important tectonic features once they can reveal key aspects of orogenic evolution. The Moqué Salient is located in the western border of the São Francisco Craton where the Neoproterozoic Brasília Belt has developed during the Brasiliano orogeny. The curvature was developed in the external Brasília Belt, a foreland fold-thrust orogenic belt. The salient is 60 km long on its N-S axis and 45 km on E-W axis, and is formed by an arched to the foreland fold-and-thrust sequence that affects Proterozoic sedimentary covers. Detail structural mapping and gravimetric data were applied to investigate the Moqué Salient evolution and compare it to orogenic curvature models in order to characterize the local and regional tectono-structural context in which the salient was developed. The study revealed some specific aspects about Brasília Belt tectonic evolution: 1 - during the Moqué Salient development, the crystalline basement was already elevated to the north and the Serra do Cristal Fault was already a significant discontinuity; 2 - basement relief seems to have influenced nappe propagation; 3 - the Moqué Salient represents one of the last features generated in the Brasília Belt during Brasiliano compressive tectonics; 4 - the Moqué Salient fits the Divergent Transport model and later controlled Lateral or Oblique Ramp Boundaries; and 5 - localized differential stress responsible for divergent transport can be explained by the Niquelândia mafic-ultramafic intrusion exhumation.

Keywords: Moqué Salient, Brasiliano orogeny, orogenic curvature, divergent transport, oblique ramp boundary.

1. Introdução

A Saliência do Moqué está localizada na borda oeste do Cráton São Francisco, mais especificamente na zona externa da Faixa Brasília Norte (Figura 1), entre a Falha do Rio dos Bois e a Falha do Paranã (Figura 3), onde se desenvolveu um cinturão de dobras e cavalgamentos de antepaís durante a orogênese Brasileira, no Neoproterozoico.

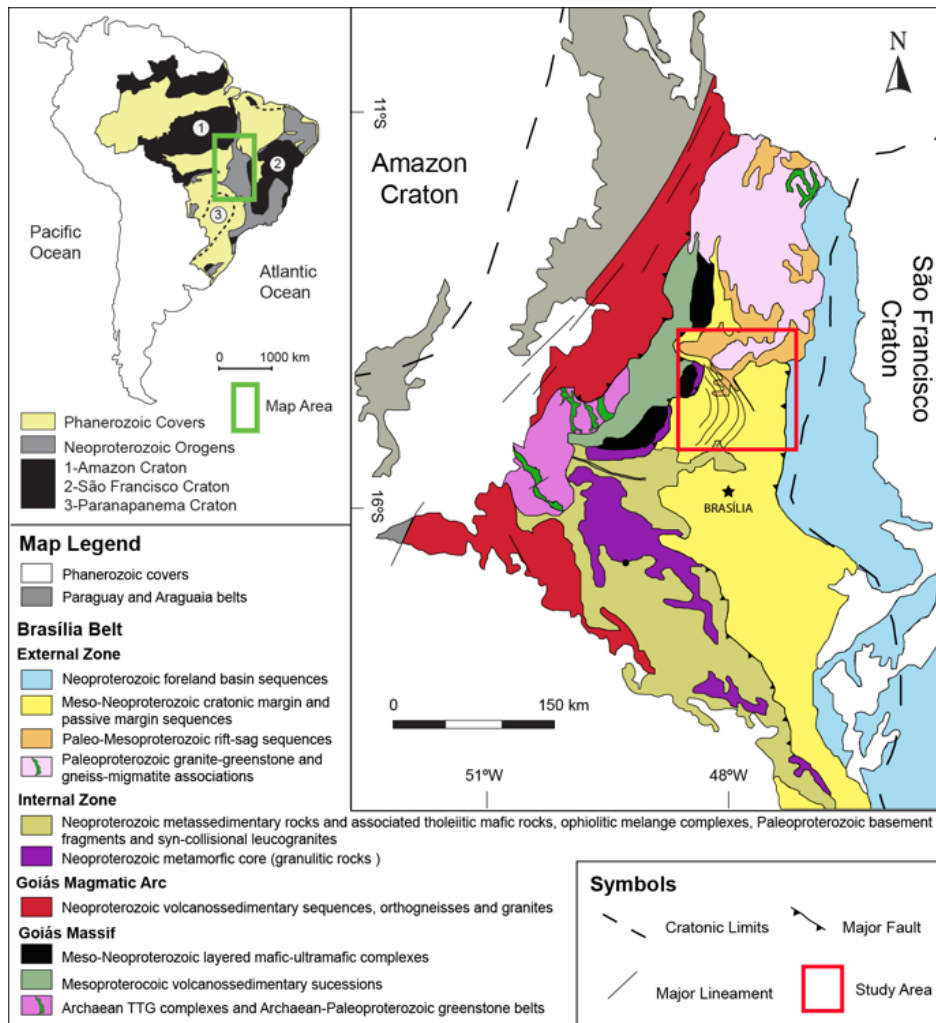


Figura 1 - Localização da Saliência do Moquéim na Faixa Brasília (Modificada de Fuck, 2004).

Trata-se de uma feição de escala regional, com extensão aproximada de 60 km no seu eixo N-S e 45 km no seu eixo E-W, formada por dobras e cavalgamentos arqueados, gerados em contexto de falha de empurrão que afeta as rochas do Grupo Paranoá (Campos et al. 2013) em toda a sua extensão e em menor proporção as rochas do Grupo Traíras (Martins-Ferreira et al. 2017) em sua porção norte.

Segundo Bates & Jackson (1987), o termo saliência é usado, em geologia, para se referir a uma área na qual os planos axiais das dobras são convexos em direção à zona externa de um cinturão de dobras. Os planos axiais do trem de dobras do Moquéim são curvos ao longo da saliência formando convexidade voltada para a zona externa da Faixa Brasília e os eixos dessas dobras encontram-se ondulados em escala quilométrica, facilitando erosão diferencial e gerando vales perpendiculares à direção de ondulação.

O presente artigo trata de documentar os avanços no trabalho de caracterização da Saliência do Moquéim e do contexto tectono-estrutural local e regional em que está inserida.

2. Métodos

A Saliência do Mequém abrange uma área bastante extensa, com aproximadamente 2.500 km² de relevo bastante acidentado e de acesso difícil, onde vastas áreas não possuem estradas. A impossibilidade de cobrir toda a área com mapeamento geológico-estrutural gerou a necessidade de desenvolvimento de métodos de mapeamento diferenciados, adaptados para o mapeamento de grandes feições.

Foi desenvolvido um método de seleção de afloramentos em áreas-chave a partir de imagens de satélite e modelos digitais de elevação aliado ao conhecimento prévio da área. Locais de interseção de estruturas regionais foram priorizados. Nos locais escolhidos como áreas-chave foram realizados trabalhos de mapeamento geológico-estrutural de detalhe.

O método de mapeamento do trem de dobras do perfil central caracterizou-se pelo registro fotográfico de toda a sequência de dobras, a montagem das fotos em sequência e a vetorização das camadas dobradas e falhas de cavalgamento (Figura 2). Foram montadas 4 seções e os vetores de cada seção foram integrados em sequência afim de confeccionar o perfil completo da zona central da Saliência do Moquém. Simultaneamente ao registro fotográfico da serra, foram mapeados em detalhe os afloramentos ao longo da estrada paralela a todo o perfil.

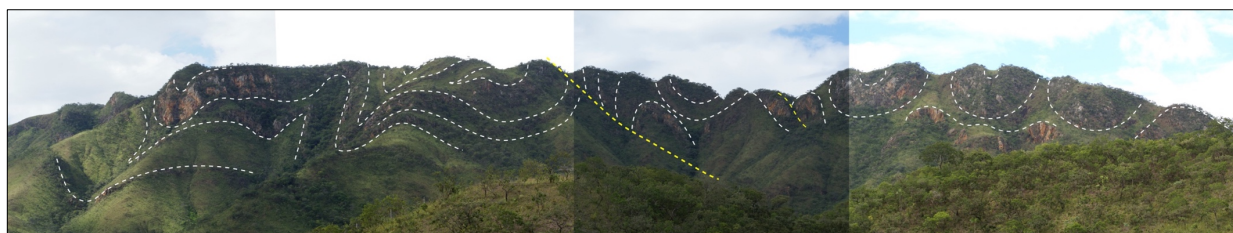


Figura 2 - Exemplo de resultado do método de montagem fotográfica sequencial e vetorização de camadas dobradas e falhas de cavalgamento (linhas pontilhadas) da seção 1, usada na montagem do Perfil Central (localização do perfil na Figura 5).

Para fins de análise dos controles subsuperficiais no desenvolvimento da Saliência do Moquém, foi confeccionado o mapa de continuação ascendente com elevação de 3.000 utilizando-se o software Oasis Montaj. O produto de continuação ascendente foi gerado a partir da anomalia Bouguer de dados gravimétricos terrestres levantados pelo IBGE. Sobre o mapa foram plotados os principais elementos estruturais da área estudada afim de se tentar estabelecer correlações morfológicas e genéticas entre as estruturas mapeadas e o relevo do embasamento cristalino.

3. Resultados

As dobras que compõem a Saliência do Moquém variam desde suaves até apertadas e localmente são isoclinais, verticais a reclinadas com eixo curvado em forma de arco e, portanto,

com vergência variável entre as direções NE e SE (Figura 3). As dobras ocorrem em escala hectométrica a quilométrica em sucessões ricas em quartzitos e decimétricas a decamétricas em sequências com predominância de pelitos com eixos variando desde horizontais a subverticais. Comumente apresentam flancos invertidos falhados e cavalgados. De maneira geral, os cavalgamentos são mais verticalizados a oeste e se tornam horizontalizados para leste (Figura 4).

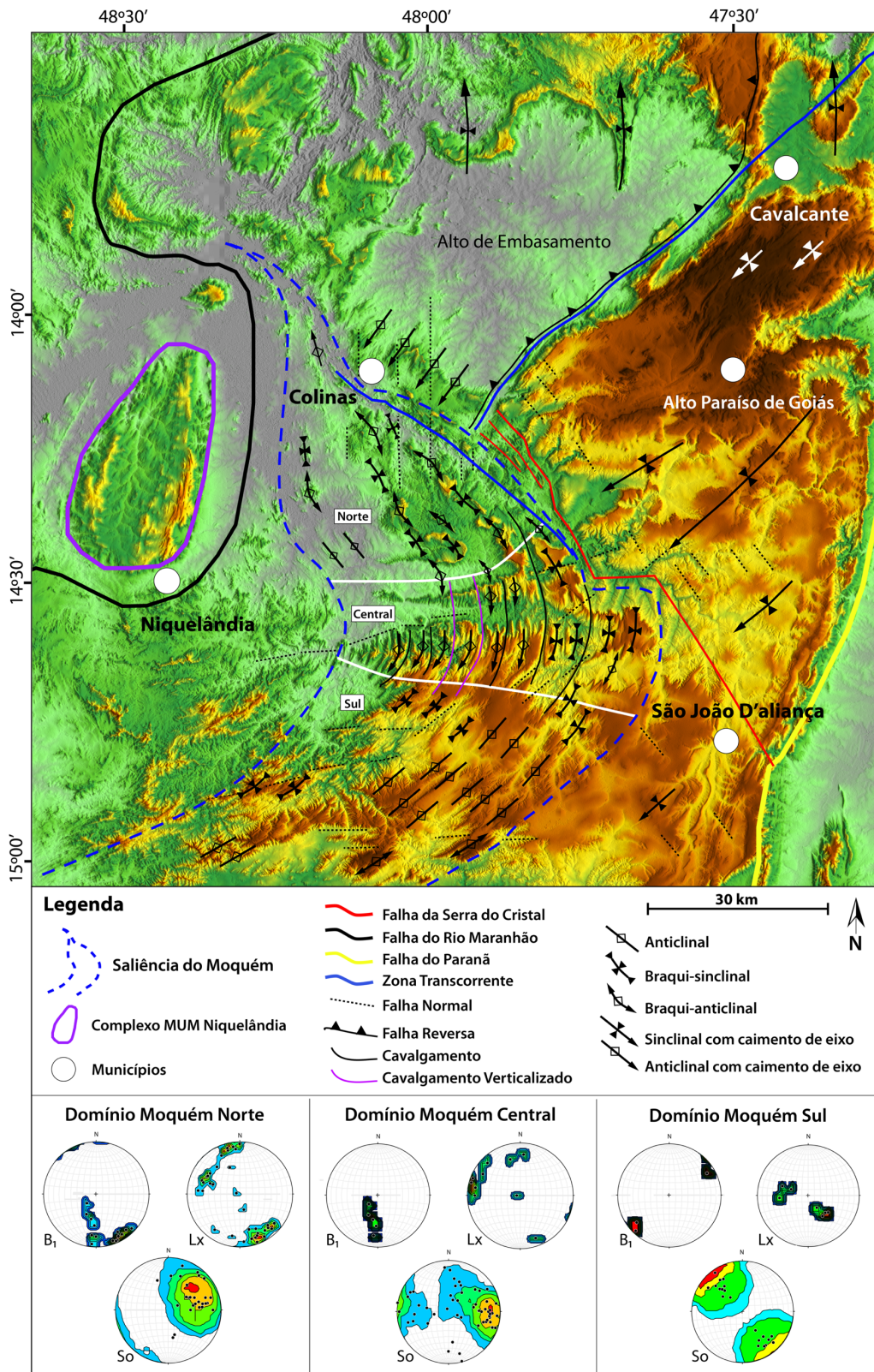


Figura 3 - Mapa estrutural simplificado sobre modelo digital de elevação da área de estudo. Em azul tracejado encontra-se destacada a Saliência do Moquéem. Em traço branco, os limites entre os setores norte, central e sul da saliência.

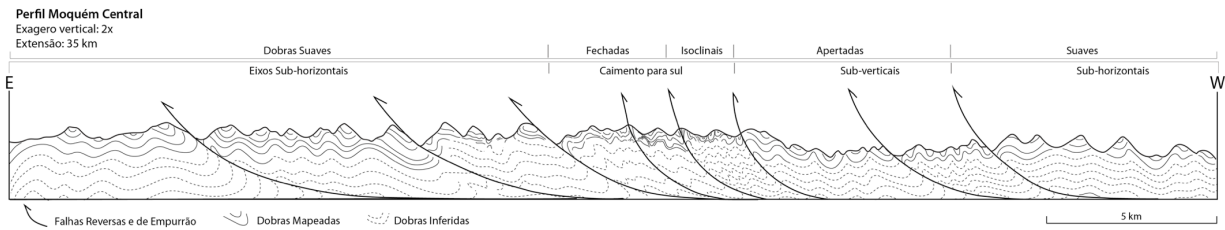


Figura 4 - Perfil foto-mapeado do trem de dobras do domínio moquém central. O perfil possui direção E-W, extensão de 35 km e exagero vertical de 2 vezes (localização do perfil na Figura 5).

As falhas de cavalgamento se unem ao plano de descolamento principal, que ocorre aproveitando a discordância entre os Grupos Paranoá e Araí, ora ao nível do conglomerado São Miguel (base do Gr. Paranoá), que ocorre recristalizado, e mais frequentemente no calcifilito de topo do Grupo Araí, que ocorre intensamente dobrado e venulado.

Em sua extremidade NE, a saliência e seu trem de dobras e cavalgamentos terminam abruptamente na Falha da Serra do Cristal e no alto do embasamento a norte, que impediram sua propagação até a altura do seu eixo central E-W. Todavia, a porção sul da saliência avança para SE em direção ao município de São João da Aliança por mais algumas dezenas de quilômetros. Na figura 5 os pontos 1, 2 e 3 indicam a propagação da porção sul da saliência no sentido SE.

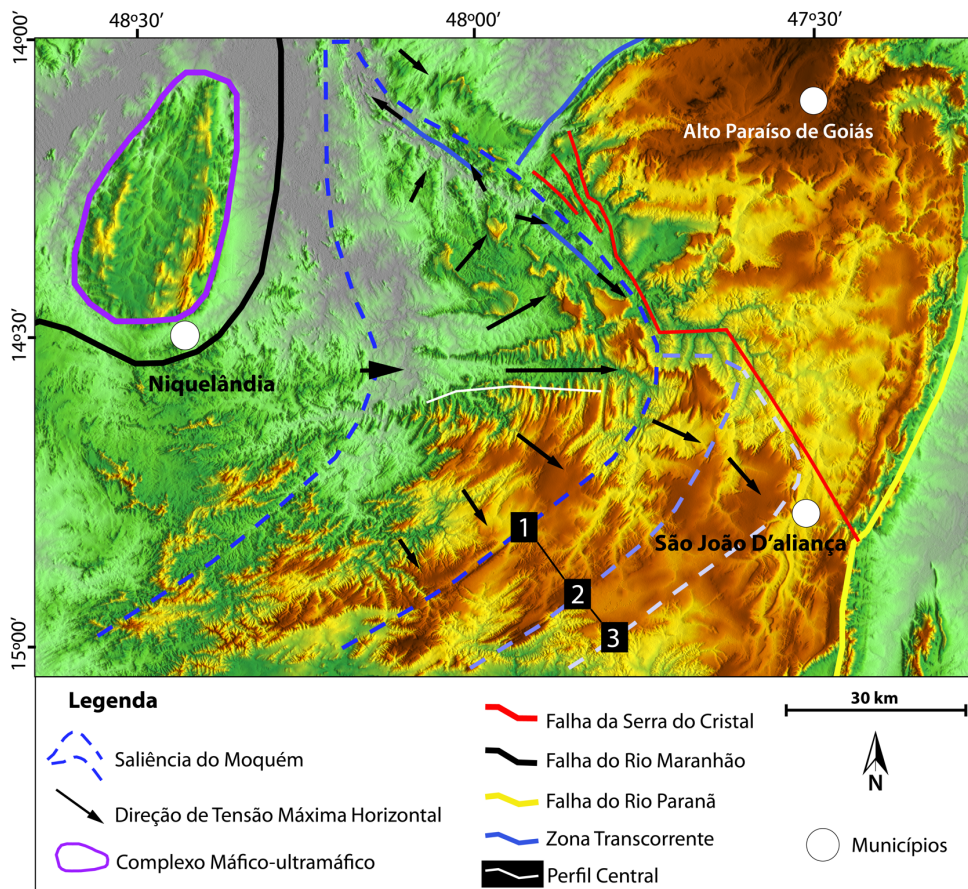


Figura 5 - Saliência do Moquém indicada sobre modelo digital de elevação. Setas pretas indicam a direção de transporte tectônico ou stress horizontal máximo (SHMax). A localização do perfil central está indicada pela linha branca contínua.

O impedimento da propagação da porção norte da saliência pelo alto do embasamento e pela Falha da Serra dos Cristais gerou zonas transcorrentes destrais e sinistrais paralelas aos eixos das dobras gerando escape lateral, principalmente próximo à falha, onde o transporte assume direção NW e SE, respectivamente, como ilustrado na Figura 5, onde ocorrem as foliações de transcorrência ($S_{3\text{-transcorrente}}$) e de crenulação extensional tardia ($S_{3\text{-normal}}$). O avanço apenas da porção sul gerou feições extensionais na porção norte da saliência e falhas de rasgamento no domínio central.

Os planos de transcorrência são restritos aos corredores de cisalhamento direcional de forma que $S_{3\text{-transcorrente}}$ ocorre apenas localmente. Nos calcifilitos o cisalhamento gera estiramento mineral (L_X) desde centimétrico a métrico como fitas de calcita (Figura 6A) e *slickensides* com estrias bem marcadas (Figura 6B) e de mesma direção geral que o L_X mineral.

A fase final de D_3 é marcada pela geração de foliação regional de direção subperpendicular a $S_{3\text{-transcorrente}}$, a S_2 e conseqüentemente aos eixos das dobras F_2 e é caracterizada por clivagem de crenulação distensiva $S_{3\text{-normal}}$ (Figura 6C) que aparentemente é co-genética à boudinagem nos ritmitos (Figura 6D), sendo a movimentação normal bem marcada por planos submilimétricos de crenulação. As superfícies de $S_{3\text{-normal}}$ são marcadas por cristais de biotita rotacionados e alinhados. A foliação $S_{3\text{-normal}}$ crenula a foliação $S_{0/1}$ em regime distensivo e imprime geometria sigmoidal.

Na porção norte e centro-norte da saliência, as zonas de transcorrência são sinistrais no limite com o alto de embasamento e destrais no limite com a Falha da Serra do Cristal e não ocorrem no domínio sul. Em ambos os casos são anastomosadas e relativamente estreitas, apesar de apresentarem limites graduais, a largura máxima das zonas transcorrentes não ultrapassa os 150 metros.

No domínio central da Saliência do Moquém, ocorrem veios de quartzo escalonados rotacionados, de escala métrica a decimétrica com pelo menos duas fases de rotação (Figura 6E), e nas barras de calcita em escala centimétrica a milimétrica, (Figura 6F), formando excelentes indicadores cinemáticos de movimentação sinistral.

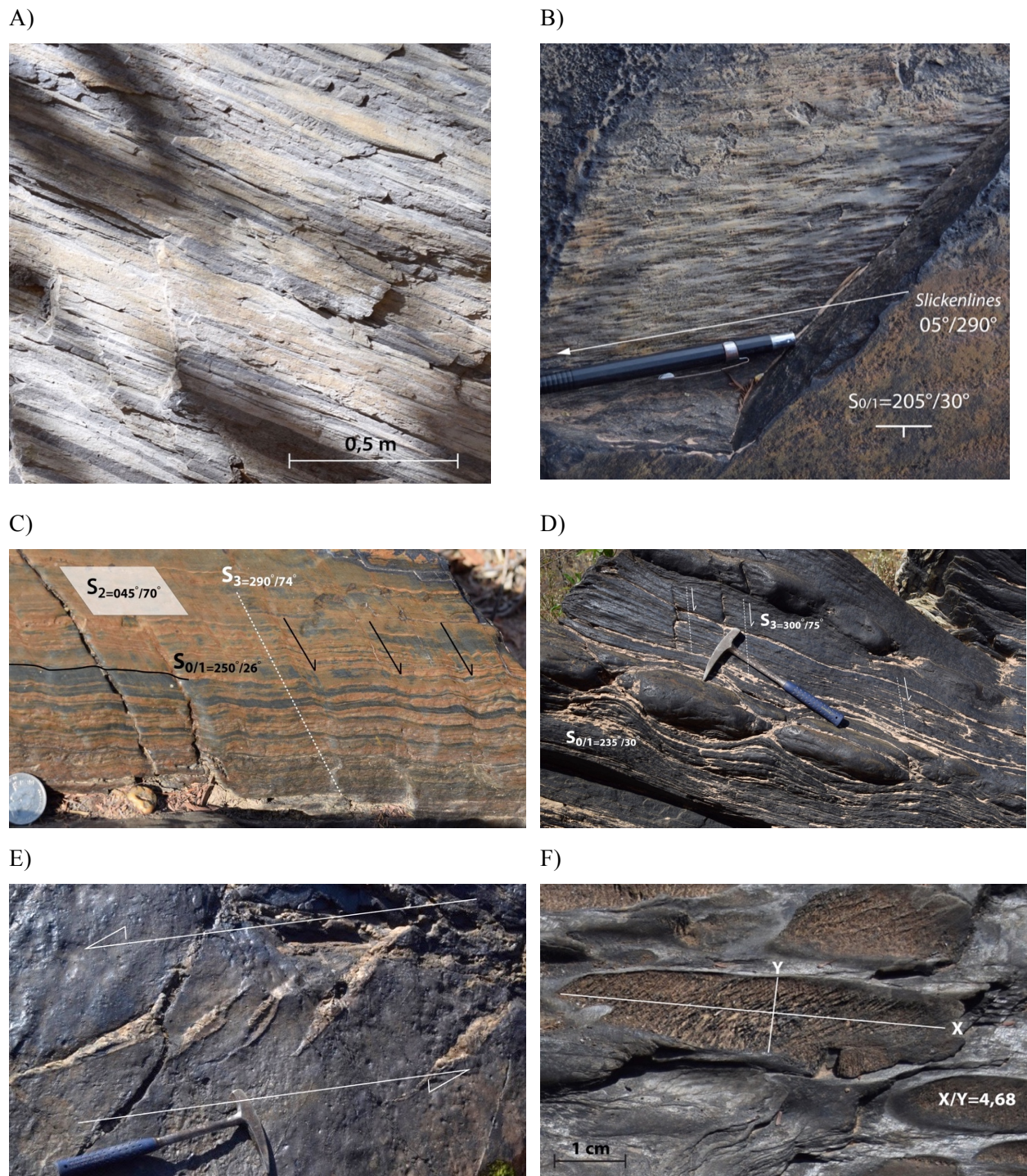


Figura 6 – A) *Slickenlines* desenvolvidas sobre superfície silicosa (*slickenside*) paralela a $S_{0/1}$. **B)** Lineação de estiramento mineral (L_x) de escala métrica formando fitas de calcita. $L_x = 15/160$. As fitas de calcita são paralelas a S_3 -transcorrente e encontram-se crenuladas por clivagem S_3 -normal. **C)** Detalhe da foliação S_3 -normal crenulando S_0 . Vista sobre o plano de S_2 . **D)** Boudinagem de camada de quartzito intercalada em metassiltitos do Grupo Araí. A direção de boudinagem é perpendicular a S_3 -normal = $300^\circ/75^\circ$. **E)** Lente carbonática deformada nos calcifilitos da Fm. Traíras do Grupo Araí a oeste da Falha da Serra do Cristal. Razão axial no plano XY do elipsoide de deformação (S_3 -transcorrente): $X/Y = 4,68$. **F)** Veios escalonados rotacionados indicando cinemática sinistral e regime semi-rúptil.

As relações de campo entre as foliações descritas acima são bem representadas no

afloramento da Figura 7 Paralelamente às zonas de transcorrência geradas por $S_{2/3}$ -transcorrente ocorrem brechas de caráter dúctil-rúptil caracterizadas por fragmentos submétricos a centimétricos de rocha metassedimentar estirados, dobrados e rompidos envoltos por nódulos globulares de escala métrica de veio de quartzo leitoso.

Os estereogramas das foliações de $S_{0/1}$, $S_{2/3}$ -transcorrente e S_{3} -normal no domínio central mostram dobras com vergência para ENE cortadas por transcorrências paralelas aos seus planos axiais e todo o conjunto crenulado por foliação NE (Figura 8).



Figura 7 - Afloramento da unidade calcifilito, topo do Grupo Araí, afetado pelo corredor de transcorrência do Subsistema Tocantinzinho que gera a foliação local S_{3} -transcorrente, que por sua vez é crenulada pela foliação regional tardia S_{3} -normal. A intersecção entre essas duas foliações gera lineação de intersecção (L_i) bastante expressiva em campo.

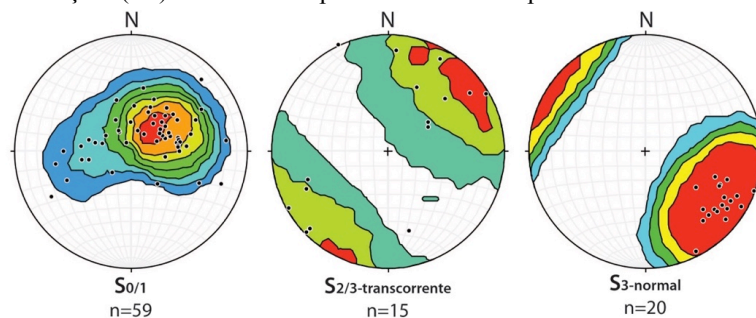


Figura 8 - Projeção estereográfica (hemisfério inferior) das foliações mapeadas na porção centro-norte do domínio Moquém.

Com o intuito de investigar possíveis controles em subsuperfície, foi gerado o mapa residual por subtração da continuação ascendente para 3.000 m da anomalia Bouguer. Na Figura 9, as estruturas regionais e locais mapeadas na região foram plotadas sobre o mapa residual Bouguer. As cores quentes representam maiores densidades e as cores frias representam menores densidades.

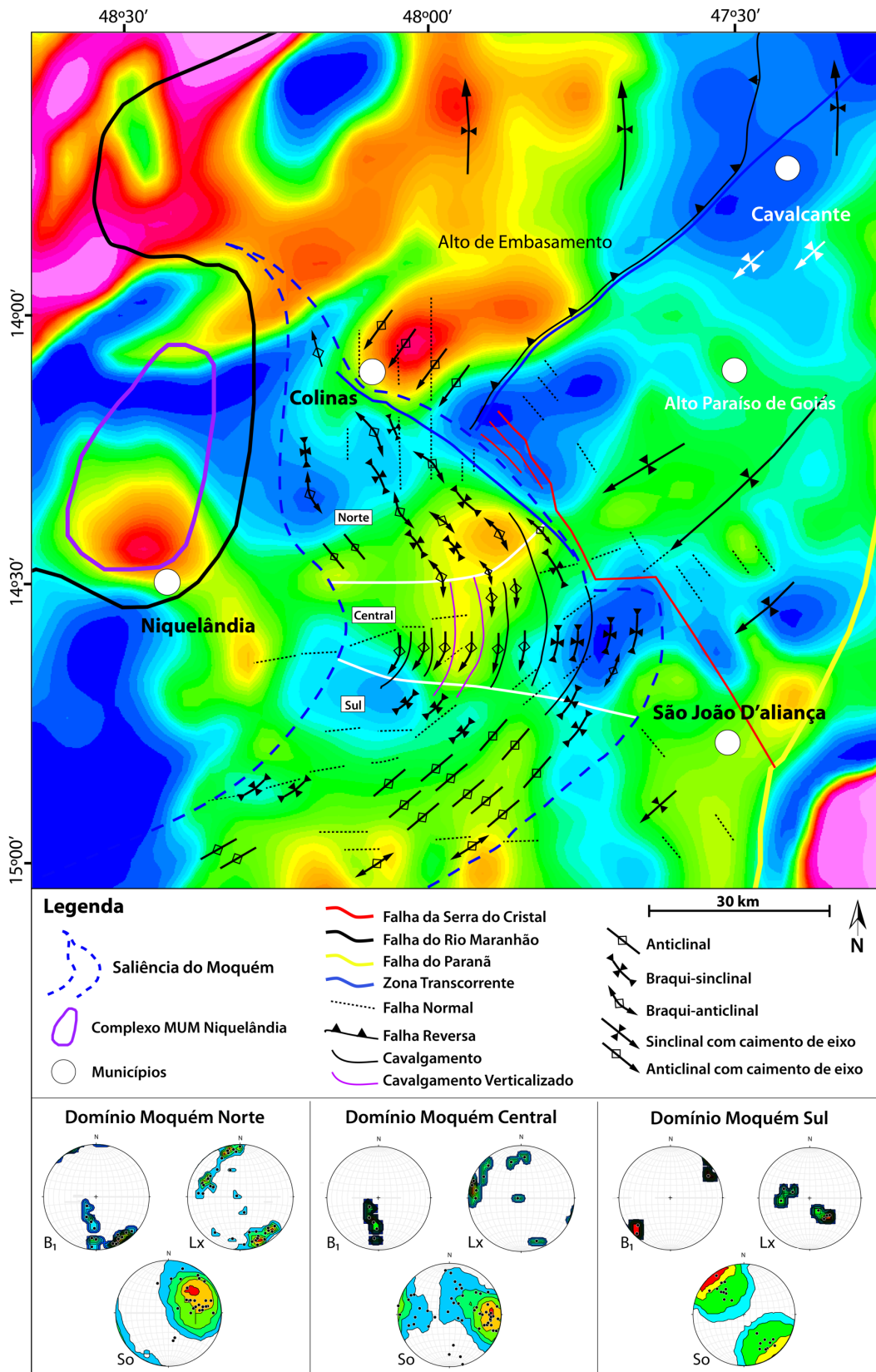


Figura 9 - Mapa estrutural simplificado sobre mapa residual da anomalia Bouguer. Estruturas replicadas da figura 3.

Nota-se claramente o alto de embasamento como anomalia de mais alta densidade e a Falha da Serra do Cristal indicada por uma variação brusca de densidade, ambos limitando o avanço do domínio norte da saliência. Fica evidente também o corpo denso do complexo máfico-ultramáfico de Niquelândia aproximadamente ao centro e a oeste da saliência.

Internamente à saliência, é possível correlacionar altos e baixos de embasamento à espessura de sedimentos, que resultam em cores mais frias onde há maior espessura e mais quentes onde a espessura sedimentar é menor. De modo aproximado, pode-se observar que as regiões de embasamento mais acidentado propiciam a geração de braqui-sinclinais e braqui-anticlinais, e áreas planas (em verde) tendem a desenvolver eixos de dobras horizontais. Os cavalgamentos verticalizados (em roxo) coincidem com anomalia semicircular de embasamento mais alto (em amarelo).

4. Discussão

Analisando a Saliência do Moquém a partir da integração dos dados coletados e à luz dos modelos de evolução de curvaturas orogênicas (Figura 10) compilados e adaptados por Whisner et al. (2014), pode-se inferir que inicialmente a saliência evoluiu de acordo com o modelo de Transporte Divergente (c), por transporte diferencial primário com trajetórias divergentes, ausência de rotação de eixo vertical de falha, avanço máximo no centro da saliência, extensão da saliência paralela à direção de transporte no centro e perpendicular nas laterais.

a. Bow and Arrow Rule	b. Orocline	c. Divergent Transport	d. Transport Parallel Simple Shear (Uniform Displacement Differentially Consumed)	e. Tear Fault Boundaries	f. Lateral or Oblique Ramp Boundaries
Primary differential parallel transport; No vertical-axis rotation of transport; No vertical-axis rotation of fault; Maximum thrust-front advance in middle of salient	Primary uniform parallel transport; Passive vertical-axis rotation of transport during bending; Vertical-axis rotation of fault (i.e. folding of fault); Extensional strain on the outer side of salient; Contractional strain on the inner side of salient	Primary differential transport with divergent trajectories; No vertical-axis rotation of fault; Maximum thrust-front advance in middle of the salient; Stretching parallel to the regional transport in the leading portion of the salient; Stretching perpendicular to the regional transport in the trailing portion of the salient	Primary differential parallel transport; No vertical-axis rotation of transport; Vertical-axis rotation of fault; Maximum thrust-front advance in middle of salient; Non uniform shortening across salient	Primary uniform parallel transport; Uniform thrust-front advance; Strike-slip faults at ends; No vertical-axis rotation of transport; No vertical-axis rotation of fault; No tangential extension; Original foreland-convex basin geometry	Primary parallel transport; Uniform or nonuniform thrust-front advance; Vertical-axis rotation of transport at edges due to lateral ramp or oblique ramp related structures; No vertical-axis rotation of fault; Contractional deformation at the edges of the salient; No tangential extension; Original foreland-convex basin geometry

Figura 10 - Modelos idealizados para explicar curvaturas orogênicas ilustrando a evolução de falhas individuais em cada cenário. Linhas tracejadas representam material anterior à deformação; linhas cheias representam o mesmo material após avanço; setas cheias indicam a direção de transporte; setas vazias indicam a direção de propagação de falha; setas com ponta dupla indicam direções de extensão tangencial da saliência (Whisner et al. 2014).

Todavia, em um segundo momento, a propagação da saliência foi impedida de avançar pelo alto do embasamento a norte e pela Falha da Serra do Cristal a centro-norte, momento em que se adequa ao modelo de Barreiras de Rampas Laterais ou Oblíquas (f), desenvolvendo avanço de empurrão não uniforme e contração intensa das extremidades barradas da saliência (braqui-anticlinais e braqui-sinclinais no norte e centro), além de falhas de rasgamento no domínio central, porém sem transporte primário paralelo, como no modelo.

A falha da Serra do Cristal teve papel importante no desenvolvimento da saliência, uma vez que o bloco elevado situado a leste funcionou como uma barreira à propagação do sistema de dobramentos, o qual é interrompido a leste desta estrutura. O contraste estratigráfico observado na sucessão basal da sequência Paranoá indica que esta estrutura já existia à época do início da deposição do Grupo Paranoá e muito provavelmente teve controle na deposição dos sedimentos da sequência Araí, sendo considerada uma descontinuidade associada ao embasamento.

A Saliência do Moquéim pode ser comparada à Saliência da Pensilvânia (Pennsylvania Salient) na Cordilheira Apalachiana Central, que possui características semelhantes e por décadas foi alvo de estudos quanto à sua gênese e significado tectônico, sempre com o desenvolvimento de hipóteses contraditórias (Lefort & Van der Voo, 1981; Gates & Valentino, 1991; Wise, 2004; Ong et al., 2007).

Recentemente, Benoit et al. (2014) avaliaram os dados sísmicos disponíveis por meio de análise de função do receptor e concluíram que a Saliência da Pensilvânia ocorre controlada por um corpo máfico de proporções quilométricas alojado em subsuperfície, resultado de um rifte Neoproterozoico abortado.

De forma similar, a Saliência do Moquéim encontra-se adjacente ao corpo máfico-ultramáfico de Niquelândia (a oeste), também atribuído a um rifteamento Neoproterozoico (Ferreira Filho et al. 2010) e apresenta vergência geral paralela à Falha do Rio Maranhão, estrutura facilitadora da exumação do corpo à superfície (D'el Rey Silva et al. 2008).

O mapeamento estrutural de detalhe evidencia 3 fases de deformação progressiva em regime compressivo, definidas pelo padrão de deformação que resulta, além de uma quarta fase extensional de relaxamento. A primeira fase D1 de direção média E-W gera foliação S1 subparalela a S0 e dobras centimétricas a decimétricas, resultado de fluxo inter-estratal nas rochas dos grupos Paranoá e topo do Araí.

A fase D2 gera foliação S2 plano-axial de dobras abertas a isoclinais, localmente verticais, de escala decamétrica a quilométrica com vergência geral variável entre NE e SE,

devido a mecanismo de transporte divergente gerado por tensão diferencial pontual. Ainda na fase D2 ocorrem cavalgamentos ao longo de falhas de propagação de dobras em todo o domínio.

Na fase final de compressão, D3, a progressão do processo de encurtamento gera escape lateral encurtando os eixos das dobras D2, facilitado pelo relevo do embasamento. O escape lateral evolui para cisalhamento transcorrente, que gera foliação S3- transcorrente aproveitando os planos de S2 nas zonas de charneira ou os flancos invertidos das dobras D2, ou os próprios planos de cavalgamentos.

Interrompida a tensão compressiva, na fase D4, a gravidade é responsável por mudar a tensão principal para a vertical, gerando falhas normais de direção N-S no domínio norte, SW-NE no domínio sul, W-E no domínio central e extremo sul e NW-SE em toda zona de frente da saliência.

4. Conclusões

- Durante o desenvolvimento da Saliência do Moquém o embasamento já se encontrava alto a norte e a Falha da Serra do Cristal já representava uma descontinuidade significativa. Essas duas feições exerceram importante controle na evolução da saliência como rampas laterais. A Falha da Serra do Cristal pode representar uma estrutura com gênese ligada ao início da evolução da sequência Araí, a partir de anisotropias do embasamento.
- O relevo do embasamento pode ter exercido controle na propagação da saliência já que possui relação direta com a ocorrência de braqui-anticlinais e braqui-sinclinais.
- A Saliência do Moquém representa uma das últimas feições regionais geradas na Faixa Brasília durante as fases compressivas da orogênese Brasileira.
- O desenvolvimento da Saliência do Moquém se deu a partir do mecanismo de Transporte Divergente controlado pelo modelo de Barreiras de Rampas Laterais ou Oblíquas.
- A tensão diferencial pontual responsável pelo transporte divergente pode ter sido produzida pela exumação do corpo denso do complexo máfico-ultramáfico de Niquelândia.

Agradecimentos

Ao Instituto de Geociências da Universidade de Brasília pelos meios destinados a viabilizar os trabalhos de campo e de escritório.

Referências Bibliográficas

- Bates, R.L.; Jackson, J.A.J. (Edts). 1987. Glossary of Geology. 3rd. Edition. American Geological Institute, Alexandria-USA.
- Benoit, M.H.; Ebinger, C.; Crampton, M. 2014. Orogenic bending around a rigid Proterozoic magmatic rift beneath the Central Appalachian Mountains, Earth and Planetary Science Letters, Volume 402, 15 Pages 197-20. ISSN 0012-821X.
- Campos, J.E.G.; Dardenne, M.A.; Martins-Ferreira, M.A.C.; Freitas-Silva, F.H. 2013 Geologia do Grupo Paranoá na porção externa da Faixa Brasília. Revista Brasileira de Geociências. V.43(3): Páginas 461 a 476.
- D'el-Rey Silva, L.J.H.; Vasconcelos, M.A.R.; Silva, D.V.G. 2008. Timing and role of the Maranhão River Thrust in the evolution of the neoproterozoic Brasília Belt and Tocantins Province, central Brazil. Gondwana Research, 13: 352-374.
- Ferreira Filho, C.F.; Pimentel, M.M.; Maria de Araujo, S.; Laux, J.H. 2010. Layered intrusions and volcanic sequences in Central Brazil: geological and geochronological constraints for Mesoproterozoic (1.25 Ga) and Neoproterozoic (0.79 Ga) igneous associations. Precambrian Res 183:617-634.
- Gates, A.E., Valentino, D.W., 1991. Late Proterozoic rift control on the shape of the Appalachians: the Pennsylvania reentrant. J. Geol. 99, 863-873.
- Lefort, J.P., van der Voo, R., 1981. A kinematic model for the collision and complete suturing between Gondwanaland and Laurussia in the Carboniferous. J. Geol. 89, 537-550.
- Ong, P.F., van der Pluijm, B.A., van der Voo, R., 2007. Early rotation and late folding in the Pennsylvania salient (U.S. Appalachians): evidence from calcite- twinning analysis of Paleozoic carbonates. Geol. Soc. Am. Bull. 119, 796-804. <http://dx.doi.org/10.1130/B26013.1>.
- Tanizaki, M.L.N.; Campos, J.E.G; Dardenne, M.A. 2015. Estratigrafia do Grupo Araí: registro de rifteamento paleoproterozoico no Brasil Central. Brazilian Journal of Geology, 45(1): 95-108.
- Whisner, S.C.; Schmidt, C.J.; and Whisner J.B. 2014. Structural analysis of the Lombard thrust sheet and adjacent areas in the Helena salient southwest Montana, USA, Journal of Structural Geology, v. 69 pp. 351-376.
- Wise, D.U. 2004. Pennsylvania salient of the Appalachians: a two-azimuth transport model based on new compilations of Piedmont data. Geology 32, 777-780. <http://dx.doi.org/10.1130/G20547.1>.

CAPÍTULO 7

DISCUSSÕES FINAIS E CONCLUSÕES

Neste capítulo, as conclusões dos artigos científicos que compõem a presente tese são sintetizadas e analisadas de forma conjunta, de maneira a integrar os estudos conduzidos em cada artigo. De forma a se evitar repetições, pontos de detalhamento dos temas estudados não são aqui abordados. É aconselhável, portanto, que o leitor se dirija aos artigos para ter acesso às conclusões específicas de cada tema estudado.

Parte dos resultados obtidos excedeu às expectativas, de modo que as abordagens consideradas bem-sucedidas merecem ser enfatizadas.

- A coleta de amostras para estudo de proveniência de sedimentos vinculada ao trabalho de mapeamento de detalhe e conseqüente detalhamento estratigráfico, permitiu a ampliação das possibilidades interpretativas e ao mesmo tempo garantiu maior convicção nos resultados alcançados.
- O uso de dados gravimétricos para modelagem da arquitetura do embasamento encoberto e da estruturação básica do rifte Araí se mostrou eficaz na região, todavia, os resultados positivos estão atrelados diretamente ao conhecimento geológico da região, acumulado durante as últimas décadas por diversos autores.
- O método de mapeamento geomorfológico usando critérios e informações geológicas detalhadas provou ser uma ferramenta importante para a análise preliminar em estudos de evolução tectônica de uma região. A aplicação do método permitiu a identificação de áreas-chave nas quais devem ser realizados trabalhos de mapeamento estrutural de detalhe, bem como perfis-chave onde devem ser mapeadas seções geológico-estruturais, com o intuito de levantar dados de campo para análise mais aprofundada da evolução tectônica local e regional.

Os estudos conduzidos no âmbito do projeto de pesquisa que configura a presente tese de doutoramento indicam que o Paleocôntinente São Francisco não representa uma plataforma isostaticamente estável e imutável que foi influenciada apenas por variações de nível eustático. Suas bacias interiores mostram que, na escala de tempo de centenas de milhões de anos, a área foi influenciada por diversos processos tectônicos distensivos e compressivos que exerceram

importante controle isostático durante toda a sua história evolutiva. As sequências intracontinentais Araí, Traíras e Paranoá, registram a evolução deste ambiente dinâmico nos períodos Paleo e Mesoproterozoico.

7.1. Evolução Tectônica das Bacias Araí e Traíras

Um sistema de riftes intracratônicos paleoproterozoicos circunda o Cráton São Francisco e separa o Bloco de Sete Lagoas (Artigo 3 - Figura 16). Reativações desse sistema no Mesoproterozoico geram abertura de novas bacias com características que sugerem dinâmica de subsidência e preenchimento condicionada à localização e direção das zonas de fraqueza preexistentes. O rifte Araí está associado, no tempo, ao sistema de riftes do Espinhaço e pode também ser contemporâneo à abertura do Aulacógeno Pirapora.

Estruturas básicas do rifte Araí foram identificadas, como falhas principais que separaram grabens e horsts, seus centros vulcânicos e plutônicos. Trata-se, todavia, de estudo preliminar e abordagens mais detalhadas podem revelar um conjunto de estruturas muito mais complexas. Entretanto, esta primeira abordagem básica apresenta avanços importantes que podem servir de base para pesquisas futuras que envolvam o rifte Araí. Além disso, o método de análise gravimétrica aliado ao mapeamento geológico, desenvolvido para este estudo, pode ser adaptado para a investigação de outros sistemas rifte antigos.

As principais estruturas do rifte Araí são o *Gráben* Nova Roma; *Gráben* Arraias-Cavalcante; *Horst* de Bezerra; *Gráben* da Prata; *Horst* Paranã Horst; *Gráben* Paranã; Falha da Água Morna e Falha da Goiana-Aboboreira. A Falha da Goiana-Aboboreira foi identificada como sendo o principal centro vulcânico do rifte (Artigo 3 - Figura 14).

O Rifte Araí pode ser classificado como do tipo intracontinental abortado, passivo, estreito a divergente, apresentando três braços principais, representados pelos grábens listados acima. Apesar de sua produção magmática, representada por granitos anorogênicos com textura rapakivi e vulcanismo bimodal associado, o rifte foi classificado como passivo, já que os volumes de vulcanismo são ínfimos se comparados com riftes intracontinentais do tipo ativo, em termos de espessura e área de abrangência dos derrames.

Contrastes de densidade significativos foram detectados a uma profundidade máxima de 38 km, que coincide com a profundidade estimada para a geração de granitos do tipo rapakivi e coincide também com o corpo de baixa velocidade detectado por estudos sísmicos.

Os contrastes de densidade abruptos, interpretados como gerados pelos falhamentos do rifte, foram mapeados e indicam que a zona de maior rejeito dessas falhas hoje está entre 4 e 8

km de profundidade, e zona de atenuação dessas falhas entre 8 e 12 km, possivelmente representando zona de transição dúctil-rúptil da crosta à época do rifteamento, se considerados o soerguimento e consequente a denudação pós-Brasílica.

Os dados geocronológicos gerados a partir de amostras coletadas em trabalho detalhado de levantamento estratigráfico, indicam que a sequência sedimentar conhecida como Formação Traíras, previamente incorporada ao Grupo Araí, possui idade máxima de deposição de 1543 ± 31 Ma. e é, portanto, separada da Formação Arraias no tempo por um lapso de no mínimo 228 Ma. Além do grande lapso temporal, as duas formações são separadas por expressiva discordância erosiva marcada por espessos níveis de conglomerados clasto-suportados.

Propõe-se, portanto, que a Formação Traíras seja separada do Grupo Araí e elevada à categoria de grupo. A proposta de formalização do Grupo Traíras inclui a elevação dos seus membros para o status de formações, respeitados os critérios do Código Brasileiro de Nomenclatura Estratigráfica. Com a individualização do Grupo Traíras, evidencia-se a distinção entre este e a bacia estritamente do tipo rifte representada pelo Grupo Araí, facilitando os trabalhos de correlação regional e o entendimento do ciclo de pulsos extensionais que afetaram o paleocontinente São Francisco no Proterozoico, com exceção do pulso extensional Toniano.

O Grupo Araí, portanto, fica restrito aos sedimentos relacionados ao rifte Estateriano e é dividido nas Formações Água Morna e Arraias. A Formação Arraias por sua vez é dividida em cinco membros, Prata, Cubículo, Buracão, Mutum e Ventura.

A instituição do Supergrupo Veadeiros é proposta como agrupamento de todas as sequências de primeira ordem formadas nos pulsos extensionais do Proterozoico na margem oeste do Cráton São Francisco. O intuito é diferenciar as sequências de fases extensionais intracratônicas daquelas de outros contextos tectônicos e estabelecer correlação com o Supergrupo Espinhaço, localizado na margem leste do Cráton. O Supergrupo Veadeiros é dividido nas fases inferior, médio e superior, respectivamente representadas pelas sequências Araí, Traíras e Paranoá e diretamente correlacionáveis com as fases inferior, média e superior do Supergrupo Espinhaço.

7.2. A Borda Oeste do Paleocontinente São Francisco

Os dados geocronológicos disponíveis na literatura, somados aos dados publicados nesta tese permitem inferir que o Grupo Serra da Mesa é crono-correlato ao Grupo Traíras. Estudos de aprofundamento no tema envolvendo detalhamento geocronológico com rígido controle estratigráfico da amostragem são necessários afim de estabelecer correlações crono-

estratigráficas mais sólidas entre essas sequências. Esses grupos são separados pela Falha do Rio Maranhão, anteriormente descrita como possível zona de sutura e conseqüentemente a borda do paleocontinente São Francisco no período Estateriano.

Todavia, tanto as sequências Traíras/Serra da Mesa, quanto os granitos anorogênicos da Suíte Serra da Mesa ca. 1,55 Ga. ocorrem em ambos os lados da Falha do Rio Maranhão, sugerindo que a borda oeste do paleocontinente São Francisco antes da formação dessas bacias e intrusão dos granitos deve ser mais a oeste, na Falha do Rio dos Bois, que separa o paleocontinente do Arco Magmático de Mara Rosa, acrescido no Neoproterozoico.

Outra evidência dessa hipótese é que o Maciço de Goiás, situado entre a Falha do Rio Maranhão e a Falha do Rio dos Bois, é a provável fonte proximal dos zircões de idade Riaciana com valores de $\epsilon_{\text{Hf}}(t)$ positivos encontrados na sequência Traíras, e como tal, já faria parte do paleocontinente durante a evolução dessa bacia no Mesoproterozoico.

Essas e outras evidências apresentadas por outros autores, abordadas em detalhe no Capítulo 3, indicam que o limite oeste do paleocontinente São Francisco, após o período Estateriano, pode coincidir com o limite oeste do Maciço de Goiás, marcado pela Falha do Rio dos Bois em superfície.

7.3. Evolução Tectônica da Bacia Paranoá

A integração dos estudos estratigráficos, estruturais e gravimétricos indicam que a Falha da Serra do Cristal representa uma descontinuidade crustal rasa, se comparada às demais descontinuidades crustais da região, relacionadas ao desenvolvimento do rifte Araí.

O controle local da sucessão estratigráfica do Grupo Paranoá indica que a falha deve ter sido gerada ou reativada na fase de instalação dessa bacia. Essa descontinuidade foi reativada também durante a orogênese Brasiliana, quando foram impressos cisalhamentos compressivos e transpressivos paralelos à falha, que também marca o limite oeste do alto de embasamento de Cavalcante, gera rampas frontais de cavalgamento e separa domínios com metamorfismo e deformação distintos.

A Falha da Serra do Cristal apresenta direção aproximadamente perpendicular às principais falhas profundas interpretadas como geradoras do rifte Araí, o que leva à interpretação de que se trata de falha de transferência do sistema Araí, reativado na instalação da sequência Paranoá. A reativação propiciou a configuração de altos estruturais com fases de soerguimento e subsidência distintos do restante da bacia, ao modelo dos Arcos Epirogênicos, comuns em bacias de margem continental relacionadas a bacias intracratônicas.

Os dados de proveniência U-Pb e Lu-Hf obtidos para o Conglomerado São Miguel em diferentes pontos de amostragem são semelhantes aos dados obtidos para o Grupo Araí e indicam que o conglomerado representa, em grande parte, produto do retrabalhamento local das rochas do Grupo Araí. A presença de clastos venulados e *boulders* de quartzitos de até 3 m de diâmetro de rochas do Grupo Araí como clastos nos conglomerados sugere eventos rúpteis extensionais anteriores ou contemporâneos à deposição do conglomerado.

A ocorrência restrita do Conglomerado Morro Vermelho apenas a uma faixa de 7 km a oeste da Falha da Serra do Cristal pode indicar reativações de estruturas prévias da fase de instalação da bacia durante processo de subsidência e/ou continuidade da tectônica extensional. Não há, todavia, indícios de eventos de elevada taxa de subsidência como era de se esperar em eventos de instalação de margem passiva, apenas falhamentos localizados de escala de algumas centenas de metros, indícios que contribuem para a hipótese de se tratar de bacia intracratônica ou bacia de margem cratônica.

Em conformidade com os resultados apresentados nos artigos da presente tese, o ambiente geotectônico de sedimentação do Grupo Paranoá na Zona Externa da Faixa Brasília é melhor caracterizado como bacia do tipo margem cratônica, adjacente e crono-correlata às bacias intracratônicas (de maior espessura) do Mesoproterozoico (ex. Espinhaço Médio).

É possível que a sequência Paranoá possa ter evoluído posteriormente ou ter sido conectada a uma bacia de margem passiva. Todavia o registro sedimentar preservado do Grupo Paranoá, na área de estudo, não permite caracterizá-lo como produto de bacia do tipo margem passiva em sentido estrito.

A presença de zircões mais jovens, de até 1.0 Ga. na sequência Canastra e a ausência desses zircões na sequência Paranoá, pode indicar que os sedimentos do Grupo Canastra foram depositados em bacia mais jovem, superposta aos sedimentos do Grupo Paranoá. Além disso, a presença de fácies típicas de águas profundas no Grupo Canastra, descritas por Dardenne (2000) sugerem que essa bacia possa ser representativa de ambiente de margem passiva na Faixa Brasília, conforme sugerido na figura 20 do artigo 4.

A partir de dados disponíveis na literatura, o intervalo entre 1,0 a 0,8 Ga., conhecido como tafrogênese toniana, é sugerido como possível fase de evolução de bacia de margem passiva na margem oeste do Cráton São Francisco.

7.4. Inversão Brasileira

As sequências Paranoá, Traíras e Araí foram invertidas durante a Orogenese Brasileira, no Neoproterozoico, evento que afetou também o embasamento. O estudo dessa inversão é importante para discriminar as estruturas que foram geradas durante a inversão daquelas que já existiam e foram reativadas. O tema é investigado por meio de dados gravimétricos, estruturais e estratigráficos na região da Chapada dos Veadeiros, onde a Saliência do Moquém está presente.

Apesar da inversão Brasileira ter sido investigada apenas localmente e não na Faixa Brasília como um todo, as conclusões do estudo permitiram avançar no conhecimento sobre alguns controles importantes que influenciaram de forma expressiva os processos de acomodação das tensões compressivas sofridas durante essa fase.

A dedução de que algumas porções do embasamento já se encontravam altas anteriormente ao evento compressivo Brasileiro, permite inferir que os processos de rifteamento anteriores deixaram heranças expressivas na arquitetura do embasamento. O fato de alguns blocos terem permanecido baixos apesar da compressão Brasileira em algumas áreas e em outras terem sido nivelados ou soerguidos acima dos blocos que eram altos, sugere que a intensidade da inversão pode ser setorizada.

O estudo mostra ainda que algumas falhas e blocos, herdados das fases extensionais, atuaram de forma a facilitar, impedir ou direcionar a propagação da deformação, dependendo da sua orientação em relação às tensões atuantes. Além disso, a temporalidade da evolução de feições estruturais regionais é inferida.

A sugestão da correlação causal do desenvolvimento da Saliência do Moquém com a exumação do corpo granulítico de Niquelândia é prematura, porém, é suportada pelo modelo de transporte divergente, que demanda uma fonte de tensão localizada no centro da curvatura, onde está localizado o referido corpo. Caso essa relação causal seja demonstrada em estudos futuros, a temporalidade da exumação do corpo granulítico poderá ser restringida como feição tardia do ciclo Brasileiro.

Os temas estudados na presente tese são considerados fundamentais para a compreensão dos processos tectônicos ocorridos durante o Éon Proterozoico nas escalas local, regional e global. Além de proporcionar algumas respostas que contribuem para o avanço do conhecimento geológico, o estudo aqui reportado gerou novas questões, que devem constituir temas de pesquisa futuras por parte dos pesquisadores envolvidos com cada tema de trabalho.