Role of Basaltic Magmatism within Parnaíba Cratonic Basin of Northeast Brazil

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Abstract

Sedimentary fill of the Parnaíba basin in northeast Brazil is punctuated by two episodes of basaltic magmatism in Jurassic and Cretaceous times known as the Mosquito and Sardinha Formations, respectively. Quantitative understanding of depth and degree of melting that generated these basalts should provide useful constraints for the history of vertical motions in this cratonic basin. Here, we carry out forward and inverse modelling of major, trace and rare earth element compositions for primitive melts from both igneous provinces. We calculate melt fraction as a function of depth in order to determine melt volumes and mantle potential temperatures. Our results suggest that both episodes of magmatism result from shallow decompression melting within the asthenospheric mantle. Mosquito basalts record an excess temperature of 55–75°C and are probably related to the widespread Central Atlantic Magmatic Province associated with initial break-up of Gondwanaland. In contrast, Sardinha basalts record localised lithospheric thinning to less than half its original thickness with excess temperatures of 15–25°C. This younger activity is probably part of the Paraná-Etendeka Large Igneous Province, which is linked to rifting between South America and Africa. We suspect that these magmatic episodes are associated with phases of regional epeirogenic uplift affecting the Parnaíba basin.

Abbreviated title: Magmatism of Parnaíba Basin

The Parnaíba basin is a cratonic basin with an area of about 600,000 km² located in northeast Brazil. Its infill primarily consists of Palaeozoic and Mesozoic fluvial to shallow marine sedimentary rocks that can be divided into five Silurian-Cretaceous super-sequences (Figure 1; Vaz *et al.*, 2007). Two magmatic events are preserved within this basin: flood basalts of the Lower Jurassic Mosquito Formation and intrusive sills of the Lower Cretaceous Sardinha Formation (de Almeida, 1986). Both events are thought to be associated with extensive dyke and sill injection that has been penetrated by deep boreholes and is visible on seismic reflection profiles. These intrusive rocks may play some role in moderating source rock maturation (e.g. Thomaz-Filho *et al.*, 2008).

Minor Precambrian and Palaeozoic graben are exposed along the edges of the basin (e.g. Jaibaras trough) and potential field studies suggest that additional graben may occur beneath the basin fill itself (de Castro *et al.*, 2014). However, there is little evidence for younger extensional faulting and the inferred graben are probably insufficient to account for the cumulative tectonic subsidence of the basin (Daly *et al.*, 2014). Instead, it has been argued that tectonic subsidence may have an exclusively thermal origin. Thus, Mesozoic magmatic activity cannot be related in a simple way to rifting processes and basaltic magmas may have travelled significant distances from their source (Ernst & Baragar, 1992).

The stratigraphy of the Parnaíba basin is summarized in Figure 2. The basin contains a 3.5 km thick sequence of shallow marine and fluvial/coastal sedimentary rocks (Vaz *et al.*, 2007). The Mosquito Formation comprises abundant basaltic lava flows that are up to 175 m thick, covering an area of ~40,000 km² in the western and central parts of the basin. In contrast, the Sardinha Formation mainly consists of shallow-level dykes and sills in the eastern part of the basin (de Almeida, 1986; Bellieni *et al.*, 1990; Fodor *et al.*, 1990).

The older Mosquito Formation occurs within Lower Jurassic sedimentary rocks of the central western part of the Parnaíba Basin. It has been radiometrically dated at 199.7 ± 2.4 Ma using ⁴⁰Ar/³⁹Ar step heating (Merle *et al.*, 2011). This date suggests that its extrusion coincides with the appearance at ~200 Ma of the Central Atlantic Magmatic Province (CAMP) which is thought to be associated with initial break up of Gondwanaland in the central Atlantic Ocean. Even though the Mosquito Formation lies 2000 km further south than the generally accepted centre of the CAMP, activity in the Parnaíba basin has been linked to putative Brazilian occurrences of CAMP magmatism (Marzoli *et al.*, 1999; Milani & Zalan, 1999; Ernesto *et al.*, 2003; Merle *et al.*, 2011). The younger Sardinha Formation mostly consists of dykes and sills with very few preserved extrusive flows. This formation is intruded into sedimentary rocks of different ages within the central eastern part of the Parnaíba Basin. Sills reach a maximum thickness of 400 m (Bellieni *et al.*, 1990; Fodor *et al.*, 1990). Basaltic flows and sills of the Sardinha Formation have been dated at 129–124 Ma by ⁴⁰Ar/³⁹Ar step heating (Baksi & Archibald, 1997). They are regarded as being associated with the Early Cretaceous Paraná-Etendeka continental flood basalt province, that is linked with opening of the South Atlantic Ocean (de Min *et al.*, 2003).

Injection of significant volumes of melt during both Jurassic and Cretaceous igneous episodes is likely to have affected ongoing tectonic subsidence of the basin. Geomorphological observations, combined with field observations made on an 18 day excursion in February 2014, suggest that the region of the basin influenced by igneous activity sits ~200 m higher than the rest of the Parnaíba basin. In fact, Cretaceous sedimentary rocks are found north and

south of, but not within, this central region (Figure 1b). It is plausible that Jurassic and Cretaceous magmatism disrupted the subsidence pattern and influenced the loci of subsequent depositional centres, palaeoenvironments, and sedimentary facies.

Sample Composition

Both phases of igneous activity are dominated by tholeiitic basalts and basaltic andesites although the Sardinha Formation contains more alkaline and more evolved rocks. There is little evidence for significantly evolved compositions within either magmatic province and asthenospheric source regions have been proposed in both cases (de Min *et al.*, 2003). Both high- and low-TiO₂ suites have been identified where less evolved low-Ti basalts appear to have resulted from a higher percentage of melting but are derived from the same source region as more evolved high-Ti basalts (Fodor *et al.*, 1990; Bellieni *et al.*, 1990).

The basalts and basaltic andesites of the Mosquito Formation are aphyric to microporphyrific. They contain augite, plagioclase, Fe-Ti oxides phenocrysts, together with pigeonite or olivine that is often altered (Fodor *et al.*, 1990; Bellieni *et al.*, 1990; Figure 3b). 12 samples were identified and selected from Fodor *et al.* (1990), de Min *et al.* (2003) and Merle *et al.* (2011) based upon a combination of completeness of trace element analyses and a MgO content of >6.8 wt%. The use of samples with high MgO content enables petrological complications associated with melt fractionation to be reduced. The chosen samples are: 4A, 8, 10, 13 (Fodor *et al.*, 1990); M/5007, M/5023, M/5033 (de Min *et al.*, 2003); and M3, M6, M13, M19, M26 (Merle *et al.*, 2011). An additional sample was collected in February 2014 from the PHYSICAL (*sic*) quarry near Araguaina (7.418° S, 48.002° W; Figure 3a,b). This sample was analysed for Ba, Ce, Nd, Rb, Th, Pb together with rare earth elements on the PerkinElmer SCIEX Elan DRC II quadrupole inductively coupled plasma mass spectrometer (ICP-MS) at the Department of Earth Sciences, University of Cambridge. Major elements together with V, Y, Zr, Nb, Cr, Cu, Sc, Ni, Sr and Zn were analysed using X-ray fluorescence (XRF) on the Panalytical PW2404 wavelength-dispersive sequential X-ray spectrometer at the School of Geosciences, University of Edinburgh (Table 1).

Samples from the Sardinha Formation have a much coarser texture compared with the Mosquito Formation (Figure 3b,d). They show a mineral assemblage of principally clinopyroxene, plagioclase and Fe-Ti oxides with minor olivine (Fodor *et al.*, 1990; Ernesto *et al.*, 2003). They also have a higher alkali content than the Mosquito Formation so that their compositions are basaltic to trachy-andesitic. Outcrops of the Sardinha Formation are generally poorly exposed and often highly altered (Figure 3c,d). 14 samples were selected according to a MgO content of >6 wt%. Sample F2 is from Fodor *et al.* (1990), samples 8207, 8209, 8214 and 8218 are from Ernesto *et al.* (2003), and nine samples were chosen from da Silva *et al.* (2017). We excluded two other samples from da Silva *et al.* (2017) that had evidence of fluid alteration. An additional sample was collected by L. Menzies in November 2015 from a location ~100 km south of Presidente Dutra (6.142° S, 44.112° W; Figure 3c,d). However, it is too compositionally evolved to be used in this study. The chosen samples were divided into three groups. The five samples from Fodor *et al.* (1990) and Ernesto *et al.* (2003) have a TiO₂ content of <2 wt% and have been assigned Cretaceous ages by K-Ar dating and

by palaeomagnetic measurements. The nine samples from da Silva *et al.* (2017) have not been dated and therefore can be only tentatively assigned to the Sardinia Formation, based on their geographical distribution. These undated samples are sub-divided into low- and high-Ti groups: 1UN21/282.10m, 1UN21/272.70m, 1UN21/240.60m, and 1UN21/254.30m have $\text{TiO}_2 < 2$ wt%; 1UN06/176.20m, 1UN32/254.10m, 1UN32/233.10m, 1UN32/224.55m, and 1UN06/139.70m have $\text{TiO}_2 > 2$ wt%.

Inverse Modelling

Whole-rock geochemical analyses of basaltic samples can be used to determine the degree of melting as a function of depth provided that reasonable assumptions can be made about the composition of the source region. Trace elements, especially the rare earth elements (REEs), are very sensitive to the cumulative amount of melting and to the relative amount of melting that occurs in the garnet and spinel stability fields. This sensitivity is due to the way in which elements such as La, Sm and Yb partition into mineral phases. The ratio of light to medium REEs (e.g. La/Sm) is negatively correlated with melt fraction due to different degrees of incompatibility of the partitioning elements. For example, the smaller the melt fraction then the greater the difference in behaviour of the relatively more compatible Sm with respect to the more incompatible La (the partition coefficients for La and Sm in olivine are $K_D=0.0005$ and 0.001 , respectively; Salters & Stracke, 2004). Thus, large ratios of La/Sm are associated with small melt fractions. The depth of melting is determined relative to the stability field of garnet since heavy REEs such as Yb are compatible in garnet. High ratios of Sm/Yb, for example, indicate melting of garnet-peridotite, where a significant amount of Yb is retained within garnet in the solid phase ($K_D=0.25$ and 0.01 for Sm and $K_D=6.6$ and 0.01 for Yb in garnet and spinel, respectively McKenzie & O’Nions, 1991; Johnson, 1998).

Here, we exploit the **INVMEL** inverse model to calculate melt fraction as a function of depth for a specified source composition (McKenzie & O’Nions, 1991). This algorithm attempts to fit the average rare earth element composition by assuming isentropic decompression melting of a dry aluminous peridotite mantle source. The isentropic melt fractions for a given potential temperature and pressure are determined using the **Ridall** forward model (McKenzie & Bickle, 1988). The formulation is based on the solidus parametrisations of Katz *et al.* (2003). Temperature estimates for a given depth are 30–50 degrees lower than those used in the original parametrisation (McKenzie & O’Nions, 1991). However, the cumulative melt fractions are comparable in both cases. Sample compositions are corrected for olivine fractionation using MgO and FeO compositions of a primitive melt produced from the chosen source composition. No correction is made for contamination by lithospheric mantle or by the crust. Consequently, judicious sample selection is required before modelling is carried out. Melting is assumed to be fractional. The inverse model iteratively calculates melt compositions, which are compared to input compositions. The root mean square (RMS) misfit between observed and calculated REEs is minimised using Powell’s conjugate direction algorithm (Press *et al.*, 1992).

In this study, we ran **INVMEL v12.0** (D. McKenzie written communication, 2014). This version exploits partition coefficients that are determined using the approach of Blundy

& Wood (2003). Best-fitting melt fractions are calculated at 5 km depth intervals. The transition from spinel to garnet peridotite is set at 63–72 km (i.e. 21–24 kbar), in accordance with results of THERMOCALC modelling carried out by Jennings & Holland (2015). A broader depth range for the transition is used here in order to ensure stability of the inverse procedure. In comparison, Jennings & Holland (2015) computed a range that corresponds to <1 kbar. Average compositions for Mosquito and Sardinha Formations used for inverse modelling are reported in Table 1. Initially, we assumed that the asthenospheric mantle source is largely homogeneous beneath northeast Brazil. Models have also been run based on different proportions of primitive and depleted mantle sources (McKenzie & O’Nions, 1991). We also included 0–5% of enrichment with a small-fraction garnet-field melt to account for the presence of small-scale fusible heterogeneities. Throughout, the ambient asthenospheric potential temperature is assumed to be 1315°C.

Results

The results of inverse modelling of basalts from the Mosquito and Sardinha Formations are shown in Figure 4 and Figures 5 and 6, respectively. Tholeiitic basalts from the Mosquito Formation were fitted using 16.1% melting of a primitive mantle source (McKenzie & O’Nions, 1991). An additional 0.5% enrichment of this source was required to fit the light REEs. Melts are formed between 49–89 km depth, with the bulk of the melt volume generated within the spinel stability field. Major element compositions indicate that the basaltic magmas have undergone ~17% fractionation prior to eruption. The calculated melt fraction distribution as a function of depth is consistent with a potential temperature range of 1370–1390°C.

Inverse modelling yields an excellent fit to REE compositions with an RMS misfit <1. The composition of the remaining trace and major elements can then be calculated using a forward model approach. The resultant predictions accurately fit compositional data within error, with the exception of Rb, K, Pb, P, Co, Mg, Cr and Ni. The residual misfit for Co, Mg, Cr and Ni reflects the relatively large amount of fractionation undergone by the parental melt. We note that only olivine fractionation is corrected for, whereas we also observe clinopyroxene, plagioclase and oxide crystallisation in these samples. Growth of Ti oxides is the probable cause of the slightly lower Ti concentrations compared with the model prediction. There is no significant difference between melting conditions for low- or high-Ti basalts. Depletion of P may indicate some amount of apatite fractionation. The relative enrichment of mobile elements Rb, K, Pb compared to the model prediction suggests that a minor amount of contamination by lithospheric or crustal melts may have occurred. Nevertheless, we have generally been able to fit Mosquito sample compositions accurately with an inverse modelling approach.

Inverse modelling of low-Ti Sardinha samples that have reliable dates suggests that melting occurred at 50–75 km depth of a source composed of 43% primitive mantle and 57% depleted mantle (McKenzie & O’Nions, 1991). Final melt fractions of 9.0%, corresponding to a potential temperature of 1330–1340°C were obtained (Figure 5). These melts underwent 28% fractionation. The residual misfit is 3.0 which is significantly larger than that obtained for the Mosquito samples. This large value is probably a consequence of incomplete analyses, the small size of this database and uncertainties of 5–15% for the rare earth element

measurements (Fodor *et al.*, 1990; Ernesto *et al.*, 2003). Nevertheless, our principal result, that a smaller cumulative melt fraction was produced over a similar depth range compared with the Mosquito Formation, is probably robust and agrees well with the conclusions of both Bellieni *et al.* (1990) and Fodor *et al.* (1990). Despite their more evolved compositions with up to 50% fractionation, inverse modelling of dated high-Ti samples (not shown here) yields similar temperatures and depths, although melt fractions are <4%. This result is in agreement with Fodor *et al.* (1990).

Inverse modelling of samples chosen from da Silva *et al.* (2017) highlights significant differences between low- and high-Ti suites. Low-Ti samples suggest that melting of a primitive mantle source occurred at 51–81 km depth. Final melt fractions of 14.3% were obtained, corresponding to a potential temperature of 1350–1370°C. These melts underwent 31% fractionation. High-Ti samples record melting at 52–72 km depth of a source that is composed of 50% primitive and 50% depleted mantle (McKenzie & O’Nions, 1991). Final melt fractions of 9.0%, corresponding to a potential temperature range of 1330–1340°C, were obtained. These melts underwent 47% fractionation (Figure 6). The fit between observed and calculated REE compositions is excellent with RMS values <0.6 for both suites. The remaining trace elements are adequately matched with some obvious exceptions. For example, Ta, Nb, P, Mg, Cr, Ni are depleted while V and Fe are enriched in the compositions of both low- and high-Ti suites, compared with model predictions. In addition, Rb, K, Ba, and Th enrichment is observed in the low-Ti suite (Figure 6b,c). Depletion of Ta and Nb could indicate minor fractionation of hydrous phases. Significant misfits for Mg, Cr, Ni, V, and Fe are probably a consequence of significant fractionation, possibly of phases other than olivine and augite. The samples used here are somewhat evolved having undergone fractionation of between 30–50%. Discrepancies for other phases may result from similar processes to those inferred for generation of basalts from the Mosquito Formation.

Discussion

Our results are in agreement with recent interpretations of the origin of magmatic activity within the Parnaíba basin. The large melt fractions and requirement for excess asthenospheric temperature for the Mosquito tholeiites, as well as their ages, suggest a relation to CAMP and break-up of the central Atlantic Ocean (Marzoli *et al.*, 1999; Merle *et al.*, 2011). This correlation suggests that lateral melt movement within the crust or lithospheric mantle occurred over distances of up to 2000 km. A well-documented analogue is the Mackenzie dyke swarm of northern Canada (Ernst & Baragar, 1992). It is likely that CAMP magmatism does not involve a single mantle source but could include multiple distinct geochemical signatures (de Min *et al.*, 2003). However, a possible link between the Mosquito Formation and the Lavras da Mangabeira of the Borborema Province further east has also been suggested as part of Brazilian CAMP activity (Ernesto *et al.*, 2003).

Here, we have attempted to model Mosquito melt generation using a purely asthenospheric, albeit enriched, source composition. This approach contrasts with previous suggestions that include involvement of lithospheric melts and crustal contamination. For example, de Min *et al.* (2003) observe decoupling between Sr-Nd isotopic compositions and Rb/Sr and Sm/Nd

values, which they attribute to metasomatic enrichment of the mantle source. Merle *et al.* (2013) observe similar anomalies within CAMP rocks from eastern North America. Callegaro *et al.* (2014) find isotopic and trace element evidence for 5–20% of crustal recycling in southeastern North America. Bellieni *et al.* (1990) argue in favour of low-pressure crustal contamination since SiO₂, K₂O, Rb and Ba correlate with Sr-Nd isotopes. These observations match with the K and Rb anomalies that we obtained by inverse modelling. Thus our assumption of no significant contamination after melt formation is probably too simplistic. However, late-stage addition of enriched melts from the crust or lithospheric mantle would require the parental melt to be more depleted and of larger volume. In consequence, our pressure and temperature values can be regarded as minimum estimates for melt formation; a greater temperature anomaly and a smaller lithospheric thickness are possible.

Interpreting the results of inverse modelling of the Sardinha samples is not straightforward because these rocks have not all been radiometrically dated. Furthermore, they are divided into low- and high-Ti suites. The distribution of melt fraction as a function of depth obtained by inverting low-Ti, but undated, Sardinha samples from da Silva *et al.* (2017) is very similar to that obtained for Mosquito samples. In contrast, high-Ti, but undated, Sardinha samples from the same authors together with low-Ti, radiometrically dated, samples suggest a different source composition compared with that of the Mosquito Formation. Melting temperatures are also lower, yielding smaller cumulative melt fractions. These conclusions appear to contradict Fodor *et al.* (1990) and Bellieni *et al.* (1990) who, using isotope and trace element compositions, suggested that both the Sardinha and Mosquito Formations are generated by different amounts of melting of an identical peridotitic source. Critically, these two studies discriminate between Sardinha and Mosquito Formations according to TiO₂ content. A significant point of agreement is that inverse modelling also requires a sub-lithospheric mantle source region. This agreement contrasts with da Silva *et al.* (2017), who place an enriched source region within the sub-continental lithospheric mantle.

Thus, low- and high-Ti suites from the eastern side of the Parnaíba Basin are indicative of both different source compositions and different potential temperatures. These differences are difficult to reconcile with a simple rifting model that invokes passive decompression melting. However, as da Silva *et al.* (2017) point out, borehole samples from this side of the basin have not been dated and thus cannot be unequivocally associated with the Sardinha Formation. Since the low-Ti samples modelled here are from a borehole located close to outcrop distribution of the Mosquito Formation, it is conceivable that these eastern samples are indeed part of the Mosquito Formation. Significantly, there is evidence for low-Ti Cretaceous magmatism elsewhere (Baksi & Archibald, 1997; Ernesto *et al.*, 2003). Evolved low-Ti basalts from the northeastern part of the basin reported by da Silva *et al.* (2017) could belong to this group. We suggest that there is negligible difference between coeval low- and high-Ti suites. Minor compositional differences could result from changes in melt fraction and/or contamination by lithospheric material (Fodor *et al.*, 1990; Bellieni *et al.*, 1990).

Shallow depths of melt generation obtained by inverse modelling corroborate independent observations that associate the Sardinha Formation with Cretaceous rifting along the margins of the nascent South Atlantic Ocean (Sial, 1976; Thomaz-Filho *et al.*, 2000). Sardinha basaltic melting is coeval with, and of similar composition to, Paraná (Serra Geral) melting

from the southern part of Brazil (Bellieni *et al.*, 1990; Baksi & Archibald, 1997). A north-south alignment of outcrops from the Parnaíba basin might be caused by minor amounts of rifting parallel to the South Atlantic margin or by pre-existing weaknesses along the Transbrasiliano lineament (Figure 1a).

Elevated mantle potential temperatures beneath the lithospheric plate together with the emplacement of significant volumes of igneous intrusions probably caused regional epeirogenic uplift of the Parnaíba basin during Jurassic and Cretaceous times. These uplift events undoubtedly played a role in the development of basin-wide Triassic/Jurassic and Cretaceous unconformities (Figure 2).

Conclusions

Magmatic activity within the Parnaíba basin resulted from shallow melting of the asthenospheric mantle. The Jurassic Mosquito melts are probably associated with CAMP magmatism which means that they could have travelled up to 2000 km laterally within the crust or lithospheric mantle. These tholeiites record excess mantle potential temperatures of ~ 55 – 75°C . In contrast, magmatism of the Cretaceous Sardinha Formation is probably linked to an extensional event at the time of break up of the South Atlantic Ocean. Inverse modelling of these samples indicates localised thinning of the lithosphere to less than half its original thickness for excess mantle potential temperatures of ~ 15 – 25°C . Both these magmatic events will have had significant impact on the subsidence and maturation history of the Parnaíba basin.

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Table 1: Average chemical composition of selected samples from Mosquito and Sardinha Formations. Major elements are reported as oxide wt%, trace elements as ppm. PHYSICAL refers to sample acquired during field excursion.

	Mosquito* PHYSICAL	Mosquito ¹		Sardinha (low Ti) ²		undated (high Ti) ³		undated (low Ti) ³	
		avg (n=13)	σ	avg (n=5)	σ	avg (n=5)	σ	avg (n=4)	σ
SiO ₂ (wt%)	52.72	51.62	1.59	52.61	0.35	48.59	0.59	51.82	0.97
Al ₂ O ₃	14.61	14.62	0.72	14.75	0.65	13.60	0.26	14.02	0.15
Fe ₂ O ₃ ^T	10.48	10.92	1.04	10.75	0.28	14.29	0.50	11.84	0.19
MgO	7.17	7.46	0.51	7.36	0.24	6.50	0.22	6.36	0.20
CaO	10.44	10.26	0.37	10.95	0.62	10.49	0.11	10.32	0.40
Na ₂ O	2.02	2.19	0.16	2.54	0.86	2.62	0.38	2.17	0.05
K ₂ O	0.84	0.88	0.28	0.75	0.40	0.38	0.09	0.63	0.08
TiO ₂	1.00	1.12	0.28	1.03	0.08	2.22	0.08	1.41	0.05
MnO	0.18	0.17	0.03	0.17	0.01	0.20	0.02	0.18	0.01
P ₂ O ₅	0.13	0.14	0.02	0.11	0.02	0.20	0.02	0.15	0.01
Total	99.58	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a
LOI (%)	0.25	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a
Li (ppm)	11.39	11.39	-	-	-	-	-	-	-
Be	0.66	0.66	-	-	-	-	-	-	-
P	556.5	556.5	-	-	-	-	-	-	-
K	8010	8010	-	12203	-	-	-	-	-
Sc	36.20	23.37	6.91	-	-	41.00	0.71	39.00	0.82
Ti	6251	6251	-	6894	-	-	-	-	-
V	249.3	287.1	57.0	505.0	-	506.0	29.8	360.8	10.4
Cr	178.4	287.9	158.0	235.8	132.5	126.0	11.4	67.5	25.0
Mn	1355	1355	-	-	-	-	-	-	-
Co	44.28	48.38	4.88	-	-	74.00	16.09	74.00	14.33
Ni	70.50	86.58	18.31	93.20	19.20	100.00	7.07	87.50	15.00
Cu	32.00	32.00	-	-	-	-	-	-	-
Zn	71.90	80.78	13.37	85.00	-	-	-	-	-
Ga	17.14	17.14	-	-	-	-	-	-	-
Rb	34.78	24.75	9.25	21.20	13.41	10.00	3.39	15.25	2.50
Sr	181.1	203.3	30.4	215.2	37.0	227.2	31.3	215.8	12.8
Y	21.90	28.30	7.80	27.60	4.34	31.60	1.52	25.00	0.82
Zr	94.90	101.61	11.28	97.00	17.59	138.80	9.01	108.00	9.27
Nb	8.40	8.34	2.59	6.60	0.89	7.00	0.71	6.50	1.00
Sn	0.84	0.84	-	-	-	-	-	-	-
Cs	0.79	0.79	-	-	-	-	-	-	-
Ba	185.5	264.0	221.9	179.8	121.2	129.2	70.7	198.3	23.8
La	11.43	11.38	2.21	8.76	2.86	9.68	1.26	11.10	0.87
Ce	24.36	24.49	3.52	23.50	3.84	24.96	2.35	25.75	2.02
Pr	3.11	2.98	0.27	-	-	3.71	0.32	3.46	0.19
Nd	13.25	14.08	2.14	12.20	2.59	17.90	1.39	15.03	0.70
Sm	3.31	3.37	0.31	4.72	-	5.38	0.27	4.11	0.27
Eu	1.06	1.10	0.14	1.57	-	1.86	0.11	1.39	0.05
Gd	3.62	3.93	0.36	-	-	5.99	0.36	4.51	0.20
Tb	0.59	0.63	0.10	0.91	-	1.03	0.06	0.78	0.04
Dy	3.87	3.84	0.34	-	-	6.22	0.36	4.75	0.21
Ho	0.77	0.78	0.07	-	-	1.20	0.07	0.94	0.04
Er	2.24	2.25	0.18	-	-	3.39	0.20	2.65	0.10
Tm	0.32	0.31	0.02	-	-	0.47	0.03	0.38	0.01
Yb	1.98	1.96	0.13	2.40	-	2.98	0.21	2.43	0.10
Lu	0.31	0.29	0.02	0.49	-	0.45	0.03	0.37	0.02
Hf	2.43	2.53	0.27	-	-	3.50	0.22	2.90	0.22
Ta	0.51	0.51	-	-	-	-	-	-	-
Tl	0.1	0.1	-	-	-	-	-	-	-
Pb	3.74	2.95	1.06	-	-	-	-	-	-
Th	2.49	1.91	0.73	-	-	0.96	0.15	2.04	0.24
U	0.61	0.45	0.18	-	-	0.30	0.13	0.50	0.06

*Major elements, including V, Y, Zr, Nb, Cr, Cu, Sc, Ni, Sr and Zn, were analysed by XRF at University of Edinburgh, other trace elements by ICP-MS at University of Cambridge.

¹Average composition of samples 4A, 8, 10 and 13 from Fodor *et al.* (1990); M/5007, M/5023 and M/5033 from de Min *et al.* (2003); M3, M6, M13, M19 and M26 from Merle *et al.* (2011). M13 is only sample with TiO₂ > 2 wt%.

²Average composition of samples F2 from Fodor *et al.* (1990); 8207, 8209, 8214 and 8218 from Ernesto *et al.* (2003).

³Average composition of samples from da Silva *et al.* (2017): 1UN06/176.20m, 1UN32/254.10m, 1UN32/233.10m, 1UN32/224.55m, and 1UN06/139.70m (high-Ti suite); 1UN21/282.10m, 1UN21/272.70m, 1UN21/240.60m, and 1UN21/254.30m (low-Ti suite).

Figures

Fig. 1: Spatial distribution of volcanism within Parnaíba basin. (a) Sample locations and regional context. Solid circles show samples from Mosquito Formation used in this study (Fodor *et al.*, 1990; de Min *et al.*, 2003; Ernesto *et al.*, 2003); open circles show samples from Mosquito Formation that have insufficient trace element data or are too evolved for our purposes (Bellieni *et al.*, 1990; Fodor *et al.*, 1990; de Min *et al.*, 2003; Merle *et al.*, 2011); solid diamonds show samples from Sardinha Formation used in this study (Fodor *et al.*, 1990; Ernesto *et al.*, 2003; da Silva *et al.*, 2017); open diamonds show samples from Sardinha Formation that have insufficient trace element data or are too evolved for our purposes (Bellieni *et al.*, 1990; Fodor *et al.*, 1990; de Min *et al.*, 2003; Ernesto *et al.*, 2003); solid squares show samples from putative Sardinha Formation used in this study (da Silva *et al.*, 2017); solid/dashed line shows Transbrasiliano lineament (Bizzi *et al.*, 2003). (b) Geological map of Parnaíba Basin modified after Bizzi *et al.* (2003).

Fig. 2: Stratigraphic column of Parnaíba basin showing ages and spatial distribution of Mosquito and Sardinha Formations modified after Vaz *et al.* (2007) courtesy of V. Rodríguez Tribaldos. Upper panel shows topographic profile across basin.

Fig. 3: (a) Outcrop of Mosquito basalt at PHYSICAL (*sic*) quarry near Araguaina (7.418° S, 48.002° W). (b) Cross-polarised image of Mosquito flow sample showing microporphyritic texture with olivine, Fe-Ti oxide and plagioclase phenocrysts in groundmass consisting of plagioclase, clinopyroxene and glass (field of view is 4.3 mm). (c) Outcrop of Sardinha diabase located ~ 100 km S of Presidente Dutra (6.142° S, 44.112° W), courtesy of L. Menzies. (d) Cross-polarised image of Sardinha sample showing granular texture of intergrown plagioclase, clinopyroxene and oxides with minor olivine (note sericitization of plagioclase and alteration rims on pyroxenes, field of view 4.3 mm).

Fig. 4: REE inverse modelling of 13 samples from Mosquito Formation. (a) Average REE composition (symbols) and best-fit model (line). (b) and (c) trace element compositions (symbols) and forward model predictions (line). (d) Melt fraction as function of depth. Calculated melt fraction obtained by fitting average REE composition in (a) shown in solid line with symbols (corrected for 17.44% olivine fractionation); uncorrected melt fraction shown in dashed lines with symbols; isentropic curves calculated using Ridall algorithm shown as grey lines labelled with potential temperatures; phase boundaries for spinel and garnet shown as vertical dashed lines; T_P is mantle potential temperature; PM is primitive mantle (McKenzie & O'Nions, 1991).

Fig. 5: REE inverse modelling of five samples of Sardinha Formation (Fodor *et al.*, 1990; Ernesto *et al.*, 2003). (a) average REE composition (symbols) and best-fit model (line). (b) and (c) trace element compositions (symbols) and forward model predictions (line). (d) Melt fraction as function of depth. Calculated melt fraction obtained by fitting average REE composition in (a) shown in solid line with symbols (corrected for 28.87% olivine fractionation); uncorrected melt fraction shown in dashed lines with symbols; isentropic curves calculated using Ridall algorithm shown as grey lines labelled with potential temperatures; phase boundaries for spinel and garnet shown as vertical dashed lines; T_P is mantle potential temperature; PM is primitive mantle; DMM is depleted MORB mantle (McKenzie & O'Nions, 1991).

Fig. 6: REE inverse modelling of 4 low-Ti and 5 high-Ti samples from Sardinha Formation (da Silva *et al.*, 2017). (a) average REE composition (symbols) and best-fit model (line). (b) and (c) trace element compositions (symbols) and forward model predictions (line). (d) Melt fraction as function of depth. Calculated melt fraction obtained by fitting average REE composition in (a) shown in solid line with symbols (corrected for olivine fractionation as indicated); uncorrected melt fraction shown in dashed lines with symbols; isentropic curves calculated using Ridall algorithm shown as grey lines labelled with potential temperatures; phase boundaries for spinel and garnet shown as vertical dashed lines; T_P is mantle potential temperature; PM is primitive mantle; DMM is depleted MORB mantle (McKenzie & O'Nions, 1991).