

This is the peer reviewed version of the following article:

Crust-mantle coupling during continental convergence and break-up: Constraints from peridotite xenoliths from the Borborema Province, northeast Brazil / Liu, Shiran; Tommasi, Andréa; Vauchez, Alain; Mazzucchelli, Maurizio. - In: TECTONOPHYSICS. - ISSN 0040-1951. - 766:(2019), pp. 249-269. [10.1016/j.tecto.2019.05.017]

Terms of use:

The terms and conditions for the reuse of this version of the manuscript are specified in the publishing policy. For all terms of use and more information see the publisher's website.

20/04/2024 01:58

(Article begins on next page)

Manuscript Number: TECTO13230R1

Title: Crust-mantle coupling during continental convergence and break-up: Constraints from peridotite xenoliths from the Borborema Province, northeast Brazil

Article Type: Research Paper

Keywords: Subcontinental mantle lithosphere
Crust-mantle coupling
Melt-rock interaction
Annealing
Olivine crystal preferred orientation
Seismic anisotropy

Corresponding Author: Dr. Andrea Tommasi, Dr

Corresponding Author's Institution: CNRS & Univ Montpellier

First Author: Shiran Liu

Order of Authors: Shiran Liu; Andrea Tommasi, Dr; Alain Vauchez; Maurizio Mazzucchelli

Abstract: We studied a suite of mantle xenoliths carried by Cenozoic volcanism in the Borborema Province, NE Brazil. These xenoliths sample a subcontinental lithospheric mantle affected by multiple continental convergence and rifting events since the Archean. Equilibrium temperatures indicate a rather hot geotherm, implying a ca. 80 km thick lithosphere. Most xenoliths have coarse-granular and coarse-porphyroclastic microstructures, recording variable degrees of annealing following deformation. The high annealing degree and equilibrated pyroxene shapes in coarse-granular peridotites equilibrated at $\sim 900^\circ\text{C}$ indicate that the last deformation event that affected these peridotites is several hundreds of Ma old. Coarse-porphyroclastic peridotites equilibrated at $950\text{--}1100^\circ\text{C}$ probably record younger (Cretaceous?) deformation in the deep lithospheric mantle. In addition, a few xenoliths show fine-porphyroclastic microstructures and equilibrium temperatures $\geq 1200^\circ\text{C}$, which imply a recent deformation, probably related to the dykes that fed the Cenozoic volcanism. Chemical and microstructural evidence for reactive percolation of melts is widespread. Variation in textural and chemical equilibrium among samples implies multiple melt percolation events well spaced in time (from Neoproterozoic or older to Cenozoic). Crystal preferred orientations of olivine and pyroxenes point to deformation controlled by dislocation creep with dominant activation of the $[100](010)$ and $[001]\{0kl\}$ slip systems in olivine and pyroxenes for all microstructures. Comparison of xenoliths' seismic properties to SKS splitting data in the nearby RCBR station together with the equilibrated microstructures in the low-temperature xenoliths point to coupled crust-mantle deformation in the Neoproterozoic (Brasiliano) continental-scale shear zones, which is still preserved in the shallow lithospheric mantle. This implies limited reworking of the lithospheric mantle in response to

extension during the opening of the Equatorial Atlantic in the Cretaceous, which in the present sampling is restricted to the base of the lithosphere.

Research Data Related to this Submission

There are no linked research data sets for this submission. The following reason is given:
Data will be made available on request

Dear Editor,

Pleased find enclosed the revised version of the ms. "Crust-mantle coupling during continental convergence and break-up: Constraints from peridotite xenoliths from the Borborema Province, northeast Brazil" (previous title "The lithospheric mantle in a continental domain submitted to multiple tectonic events: Peridotite xenoliths from the Borborema Province, northeast Brazil").

We have carefully revised the ms. implementing all suggestions from the two reviewers. In short, we fully reorganized the data presentation and the discussion as suggested by reviewer 1 and revised all figures to enhance their readability. We also reduced the number of figures and transferred the table with the chemical analyses data to the Supplementary on line material.

The result is a slightly shorter article (by 1 page and 1 figure), which we hope is easier to read and better presents the data and conveys our conclusions.

Sincerely yours,

Shiran Liu and Andréa Tommasi on the behalf of all co-authors

Crust-mantle coupling during continental convergence and break-up: Constraints from peridotite xenoliths from the Borborema Province, northeast Brazil

Shiran Liu^{1,2}, Andréa Tommasi¹, Alain Vauchez¹, Maurizio Mazzucchelli³

¹*Géosciences Montpellier, CNRS & Université de Montpellier, Montpellier, France.*

²*Key Laboratory of Orogenic Belts and Crustal Evolution, School of Earth and Space Sciences, Peking University, Beijing, China.*

³*Dipartimento di Scienze Chimiche e Geologiche, Università di Modena e Reggio Emilia, Modena, Italy.*

Abstract

We studied a suite of mantle xenoliths carried by Cenozoic volcanism in the Borborema Province, NE Brazil. These xenoliths sample a subcontinental lithospheric mantle affected by multiple continental convergence and rifting events since the Archean. Equilibrium temperatures indicate a rather hot geotherm, implying a ca. 80 km thick lithosphere. Most xenoliths have coarse-granular and coarse-porphroclastic microstructures, recording variable degrees of annealing following deformation. The high annealing degree and equilibrated pyroxene shapes in coarse-granular peridotites equilibrated at ~900°C indicate that the last deformation event that affected these peridotites is several hundreds of Ma old. Coarse-porphroclastic peridotites equilibrated at 950-1100°C probably record younger (Cretaceous?) deformation in the deep lithospheric mantle. In addition, a few xenoliths show fine-porphroclastic microstructures and equilibrium temperatures $\geq 1200^\circ\text{C}$, which imply a recent deformation, probably related to the dykes that fed the Cenozoic volcanism. Chemical and microstructural evidence for reactive percolation of melts is widespread. Variation in textural and chemical equilibrium among samples implies multiple melt percolation events well spaced in time (from Neoproterozoic or older to Cenozoic). Crystal preferred orientations of olivine and pyroxenes point to deformation controlled by dislocation creep with dominant activation of the [100](010) and [001]{0k1} slip systems in olivine and pyroxenes for all microstructures. Comparison of xenoliths' seismic properties to SKS splitting data in the nearby RCBR station together with the equilibrated microstructures in the low-temperature xenoliths point to coupled crust-mantle deformation in the Neoproterozoic (Brasiliano) continental-scale shear zones, which is still preserved in the shallow lithospheric mantle. This implies limited reworking of the lithospheric mantle in response to extension during the opening of the Equatorial Atlantic in the Cretaceous, which in the present sampling is restricted to the base of the lithosphere.

Dear editor,

Please find enclosed a point-by-point answers (in black) to the reviewers comments (in blue).

We have implemented all requested changes, with the exception of your suggestion of production of "a nice take-home message figure summarizing the results and allowing at once to capture the history (eg infiltrations, melting or deformation events), relative chronology and possible past location of the samples with respect to depth and major tectonic events (with of course appropriate question marks where needed)". We tried, but there is too much uncertainty in the timing of the different tectono-magmatic events in the mantle and in their correlation with the dated crustal record. We can clearly convey this uncertainty in the text. However, we feel that we cannot do it in a figure even if question marks were added (we would need to add question marks to all parts of the figure). It would also be difficult to indicate the variation in "robustness" of the various interpretations. A last problem in preparing such a figure is that, although we may infer the kinematics of the deformation frozen in the shallow lithospheric mantle based on the consistency between crustal deformation and SKS splitting data, we have no constraints on the kinematics of the possible Cretaceous deformation in the deep lithospheric mantle. The SKS splitting data is better explained if flow directions in the mantle are parallel to NNE over the entire lithospheric mantle and the asthenosphere. Yet such a direction is at high angle to all Cretaceous extensional structures... Therefore, we reworked the text in order to clarify our conclusions on the relative timing of the tectono-magmatic events in the mantle lithosphere and on the crust-mantle coupling during these events, but prefer not to produce a summary interpretative figure, which may lead the readers to imagine that these conclusions are more robust than they are really.

We fully reorganized the data presentation and the discussion as suggested by reviewer 1 and revised all figures to enhance their readability. The revision resulted in a slightly shorter article (by 1 page and 1 figure), which we hope is easier to read and better conveys our conclusions. All major changes are highlighted in the annotated version of the ms.

Comments by Sandra Piazzolo.

The manuscript "The lithospheric mantle in a continental domain submitted to multiple tectonic events: Peridotite xenoliths from the Borborema Province, northeast Brazil" by Liu et al. represents an extensive dataset on a suite of xenoliths from the subcontinental lithosphere of NE Brazil. It is without doubt a lovely dataset well worth publishing, the authors collected both orientation and chemical data and relate the data to existing geophysical datasets. The data is in general very nicely presented and thoroughly interrogated (mostly in form of bulk data), however at present the manuscript feels a bit like an amalgamation of data with not a clear hypothesis or research question to test in mind. This also means that there are lots of different strands of information followed in the manuscript and it leaves the reader a bit lost. So currently - after a fast read (what most reader do) currently the reader learns that xenoliths are varied and the lithospheric mantle is heterogeneous with different annealing and metasomatic modifications. But I think there is more you can get out of the data set. I believe that this manuscript could highly benefit from streamlining towards a bigger question: The authors provide the question in the first paragraph of their introduction (as well as the last part of their discussion) - is the deformation of crust and mantle coupled or not? Are there times when there is decoupling prevalent and other times were there is coupling seen. This is then a

large-scale big picture question of interest to the general geological community. And very few datasets allow such an evaluation.

This would mean that the authors would need to link more explicitly some of the microstructures to the geotectonic setting known from the crustal parts of the lithosphere. Maybe in some xenoliths you have different domains so you can develop a time sequence - it looks to me that you could do that? At present the authors do it for the last event - linking the high stress recrystallization to melt intrusion (nice), however can one make some links for the other 900 C and >1000 C structures?

Overall the paper will benefit from streamlining with the big paper in mind, please have a think if you need all figures. Maybe you can put some in supplementary (see suggestions). After the revisions the paper will be ideally reduced by 10-15%. Here some shuffling of section will help.

We sincerely thank S. Piazzolo for her throughout and very constructive review. We extensively reworked the presentation of the data and the discussion as suggested, so that we better streamline the presentation of the evidence and the discussion of the relative timing of the various tectono-magmatic events and their relation to the dated crustal structures. This allows to discuss the question of crust-mantle coupling during major tectonic events. In the present case, we propose that there is strong evidence for pervasive deformation of the entire lithosphere, with strong crust-mantle coupling, during the formation of the Borborema shear zone system in an intraplate setting during the late stages of the continental convergence that lead to the formation of the Gondwana supercontinent. In contrast, extension leading to continental break-up and formation of the Equatorial Atlantic Ocean in the Cretaceous seems to have resulted limited reworking of the lithospheric mantle, which in the present xenolith sampling is only recorded in a few xenoliths with equilibration temperatures >1000°C. Moreover, SKS splitting data does not support flow directions in the lithospheric mantle consistent with the extensional structures. Altogether, these observations point to decoupling between crustal and mantle deformation or to very localized deformation in response to extension in the lithospheric mantle.

Comments and suggestions:

1) Title: Have another look at it, can you provide a statement regarding coupling? Or deformation / metasomatism sequences you identify?

We revised the title as suggested. However, if the reviewers consider the new title too appealing, we are happy with the original one (The lithospheric mantle in a continental domain affected by multiple tectonic events: Peridotite xenoliths from the Borborema Province, northeast Brazil), which described precisely the observations presented in the article.

2) Abstract: needs some rephrasing to put up the question to be answered more clearly. Rework after defining more precisely which structures are probably related to which tectonic event. The current last two sentences are interesting but please add what that means - i.e. fabrics can be frozen in in the mantle part of the lithosphere over billions of years (if you argue for frozen in strike slip fabrics (actually not much about this in the current main body of text) - this is interesting because that would mean that such features would probably influence future tectonic events in orientation and deformation rate. In addition if that is true then there would be a clear direct coupling from continental crust to the continental lithosphere during strike slip - an important feature to know if one wants to understand large scale tectonics. Also what does it mean that there is a significant mismatch of calculated and measured seismic delay times? Does that mean that the xenolith only sample an old component and not

current deformation state? (again not really discussed in the main text - please rectify) (Please note, these questions need to be addressed in the discussion and depending on outcome the abstract needs to be modified).

We have reworked the abstract to better convey the conclusions of the article concerning the relative timing of the deformation and metasomatism in the mantle and crust-mantle coupling. We also rephrased the part presenting the comparison with seismic anisotropy data leaving to the discussion and conclusion of the article the discussion of the fact that the xenoliths only sample the lithospheric anisotropy, which cannot account alone for the very high delay times measured in station RCBR, and the implications of this to the present-day asthenospheric deformation.

2) Introduction: If you want to link crustal and lithospheric deformation, then it becomes important to make sure that the reader knows about the main events as known from the crust. Therefore I suggest to keep the regional part of the introduction shorter in the introduction- you only need to argue that the province you chose is a nice testing ground for the question asked. (Hence move some of the regional information into the Geological background - this then needs to be more explicit)

We added a phrase explaining why direct observations on the mantle structures and compositions are needed in addition to seismic anisotropy to discuss crust-mantle coupling during major tectonic episodes and shortened the regional part.

3) Chapter 2.1. To help the reader, I strongly recommend to provide a summary table of tectonic events with little comics of structural movement at different times and intrusion - like a table that has the columns time, major tectonic event (name and character (extensional, compression), scale of deformation (10s km, 100 km etc.), igneous activity (yes/no/ what type) - then you can easily refer to different events later - and even a reader not familiar with the area (most readers) will be able to follow your links. Hopefully it will also help to streamline this section. Concentrate on the main events. (modify figure accordingly - see also below).

We added to Figure 1 a time line of the major tectono-magmatic episodes recorded in the crust. We also revised the map and its legend to better convey the temporal evolution.

4) Chapter 2.2.: Is there a possibility to have a summary geophysical diagram that shows the essentials of what we know right now? Alternatively make the geophysical signal clearer in the map please.

We added to Figure 1 magnetic anomaly and seismic tomography maps, which image the structures in the middle crust and the seismic velocity anomalies in the shallow sublithospheric mantle, respectively. We also made the SKS splitting data clearer by enhancing the size of the symbols and presenting it onto both the geological map and on the magnetic anomaly map. This should allow the reader to visualize the relation between crustal structures and seismic anisotropy in the upper mantle.

4) Sampling: In Figure 1, can you make the sampling sites more easy to see - using the abbreviations. Then you can shorten the text and refer the reader to the figure and the table.

5) Chapter 3.1. should then be only half of what it is now, can be easily streamlined by putting GPS data into the table etc.

We enhanced the size of the symbols presenting the sampling locations in Figure 1 and revised and shortened the text of section 3.1 (lines 316-331).

- 6) Sections 3.1/3.2/3.3. are nicely written and all necessary data needed is there.
- 7) Why don't you first have the modal percentages and the mineral compositions combined - they belong together. This should also help to streamline the two sections as well.
- 9) 4.2. - Why not start with the section (current line 479 and following) with the definition of the microstructural types and then provide what makes them typical, again this will help streamline the text and make it easier for the reader. It is easy to get confused. In the whole results section it will be useful to use the microstructure types as anchor points - then the structure is similar in each results section, making it easier for the reader.

We reorganized the data presentation as suggested. It now starts with the presentation of the microstructures (section 4.1), which allow to define the microstructural groups from the start, and use this classification to present the CPO (section 4.2) and composition data (sections 4.3 and 4.4). As suggested by the reviewer, modal compositions are now presented just before the mineral composition data.

For the statistics - for a porphyroclastic microstructure you need to divide the structure and the bulk statistics into two parts the porphyroclasts/blasts and the matrix. This is not done and therefore the statistical values that are provided and plotted in figures are misleading - hence produce two datasets for these microstructure type please. And then replot, it should become much clearer what is going on.

For the fine-porphyroclastic peridotites, we now present the CPO data separately for porphyroclasts and neoblasts (new Figs. 5 and 6). The coarse-porphyroclastic peridotites present a continuous variation in grain sizes and some annealing, which render the distinction of porphyroclasts and neoblasts very difficult. This is now clearly stated in the text (lines 500-501).

Since you have already presented the modal percentages and chemical data, can you please when you describe the microstructures of each microstructural type, also link in the modal percentages and chemistry. This is done in some case, please put it in systematically (just needs a half sentence here and there) - please make sure you refer to the relevant figures

By reorganizing the text, we can now clearly describe the relation or not between microstructures and modal compositions or mineral chemistry. This is done in lines 615-620 and 655-656.

We also now clearly state that there are no relations between sampling site and microstructures, modal compositions, or mineral chemistry. All types are sampled in the Pico do Cabugi and also in the other sampled Macau volcanics (lines 461-467, 612-613, 657-658).

I would make the maps in Figure 2 larger and use some arrows you can then backrefer to in the text).

Figures were re-organized in consequence of the changes in the text and EBSD maps are now presented together with the microstructures in new Fig. 3. This allows to better refer to these maps, which clearly illustrate the olivine and pyroxene grain shapes, in the microstructures description section.

- 10) Section 4.3. CPOs (see comments on figure 7) nicely done, I notice that in figure 8 you have "recrystallized areas" - is that for the fine porphyroclastic ones? Please be explicit - they show a nice trend - so you have the data, show also in Fig. 7 please (did I miss it?) (Just a comment: I wonder how much of the CPO is inherited by metasomatic reactions? This might be its own contribution though - but would be good to know ...)

The CPO presentation figures (new Figs. 5 and 6) have been re-designed to clearly present olivine porphyroclasts and neoblasts data for the fine-porphyroclastic peridotites. The

neoblasts orientations are clearly inherited from the neoblasts ones (cf. pole figures in Fig. 6), but there is a strong dispersion, probably due to dynamic recrystallization. One cannot discard the influence of fluids or melts in this recrystallization, but there is no positive evidence for it, such crystallization of other mineral phases in the recrystallized domains (cf. Fig. 3g).

11) Discussion: 5.1. What would trigger the Cenozoic melt percolation? What is the geotectonic setting for that? I am a bit confused, on the one hand you say that the geotherm must be high (based on Figure 11, but then you say that the early equilibrium textures have survived and therefore high geotherms were prevalent in the Mesozoic ? So do the shallower sample - represent relatively recent high geotherm and the deep samples represent an earlier fabric, and an early high T geotherm? Please clarify. It would be good to streamline. Important for seismic interpretation as well ...

I think it would be better to have discussion point 5.2 first - as you can use the microstructures to argue for annealing/ long time at high T etc. - please switch around - hopefully it helps to streamline.

We completely reorganized the discussion following the reviewer suggestion. Now it starts by the discussion of the microstructures, then we discuss the thermal constraints and compare them to geophysical data and calculated melting conditions for the last two magmatic episodes, compare the seismic properties of the samples to seismic anisotropy data, and, at last try to use all this information to relate the mantle and crustal evolutions.

Concerning the questions above: the equilibrium temperatures and absence of garnet despite fertile compositions imply a rather hot geotherm in the Cenozoic, which is consistent with the calculated melting conditions for the most primitive Macau volcanics. Both datasets imply a hotter than normal sublithospheric mantle. Geophysical data points to a hotter than normal sublithospheric mantle nowadays. One may infer therefore that the present-day conditions persist since the Cenozoic. For the pre-Cenozoic thermal state, the only constraints we have are the calculated melting conditions for the most primitive Cretaceous basalts of the Ceara-Mirim dykes indicate still shallower partial melting. Based on these data, we suggest that the lithosphere-asthenosphere boundary might have been shallower in the Cretaceous. In conclusion, the shallower samples represent old lithospheric mantle, but the deeper ones might represent material accreted to the lithosphere by cooling since the Cretaceous. We have re-written this section of the discussion (lines 1073-1088) to clarify this point.

12) 5.2. I think it would be best to go through the three microstructural types and discuss their general features, with a bit of "outliers" - but really you want to emphasize the general trends. Many of the arguments you bring up in the microstructures part you can directly link to the geotherm - so switch 5.1. and 5.2. around. If you show the matrix versus porphyroclast data - then you can discuss here in more detail, it will help in setting up the evolution of these samples better.

Done, see answer to previous point

13) 5.3. I think you should take this part out, not needed - alternatively, shorten considerably - just point out the main difference relatively uniform versus high variations in your studied samples.

We have eliminated this section as suggested. However, we kept the oceanic samples in Fig. 2, since we use the comparison in the annealing degree between the FN samples, for which we can infer a maximum age for their accretion to the lithosphere and hence for the freezing of the microstructures, with the one measured in the Borborema peridotites to try to constrain the age of the deformation frozen in the coarse-granular and coarse-porphyroclastic peridotites from the continental domain (lines 1194-1199 & 1242-1257).

14) 5.4. Nice section - just rewrite a little bit keeping the big question in mind.
15) 5.5. this is the BIG story discussion, all other discussion sections (5.1.-5.4) should be written to feed into this one. Currently you do not "feed" off a) the metasomatic story, b) the geotherm story - If the geotherm seen is Mesozoic and cenozoic - then how is the Neoproterozoic structure preserved - are the hT only static with very little deformation? Maybe the only deformation signature is actually the melt percolation and annealing, but no CPO development? This section needs a major rehaul, there is a lot more in there, thinking about what can trigger mantle metasomatism - and when in the annealing and deformation history of the xenoliths that metasomatism is inflicted should provide some pinpoints for a relative time - melt percolation - deformation - annealing path. If you can construct these for the three main microstructure types then I think you have a nice story, you have done that half or 2/3rd of the way, but it needs to be clear and explicit. This discussion part will feed directly in rephrasing and restructuring the other discussion parts.

Neoproterozoic structures are preserved in the shallow part of the lithospheric mantle, which after the Neoproterozoic deformation have only been affected by annealing and minor (localized) reactive melt percolation leading to refertilisation - the coarse-granular peridotites. Neither of these processes produces major changes in the olivine CPO.

The few low-temperature coarse-porphyroclastic samples might record more recent deformation, but strains are probably small and have not significantly modified the CPO. This last interpretation is based on two observations: (1) the lack of evidence for any high-stress, low-temperature (<1000°C), high-strain deformation in the studied sampling (considering that the geotherm in the Mesozoic was only slightly hotter than the Cenozoic one, temperatures in the shallow section of the lithospheric mantle would not have been very different) and (2) the seismic anisotropy data derived from both SKS splitting and anisotropic receiver functions, which point to a fabric subparallel to the Neoproterozoic crustal structures in the lithospheric mantle.

This section of the discussion has been reorganized and extensively re-written. The points above are presented in the discussion lines 1223-1265.

16) Conclusions - largely fine, but needs some reworking after 5.5. is redone.
Done.

Figures:

Figure 1: Nice figure, but you need to add the tectonostratigraphic event table - Ideally you can link the two through numbers or colours (I think it will be worth the effort)

We added to Figure 1 a time line of the major tectono-magmatic episodes recorded in the crust. We also revised the map and its legend to better convey the temporal evolution. We also added maps of the geophysical data illustrating the continuation of the Neoproterozoic structures in the middle crust and seismic velocity anomalies which point to higher than normal temperatures in the sublithospheric mantle temperature beneath the study area.

Figure 2. I am not sure that you need both figure 2a and b, they show the same data twice, I would opt for 2b. Then you can have the legend bigger so that it is legible. The white writing on the maps is intelligible, please make maybe black writing with a white background

We have completely reorganized the order of figures to follow the new order of presentation of the data. The old Fig. 2, which is now Fig. 7 has been strongly simplified, since now the EBSD phase maps are presented together with the microstructures. We kept nevertheless the ternary diagram as an insert, since it is the most traditional way to present peridotites modes and as it is now, it does not use supplementary space.

Figure 3. No need in the manuscript (can be supplementary)
Deleted. Grain sizes are presented in Table 2.

Figure 4-5. Please use some extra arrows to point out specific features, e.g. irregular boundaries between different phases (CPX-OI) if there are subgrains (I assume you have the EBSD data for each of these datasets - then point out these out - why not make for each one a little inset showing an EBSD map with typical GOS and subgrain boundaries marked. This would help a lot in the discussion and the understanding of the reader what you are focussing on. Will also help to streamline the text.

These figures, which are now Fig. 3 and 4 have been extensively revised. For lack of space, we do not show intragranular misorientation maps for all microstructures. However, we do it in Fig. 3 for the fine-porphyroclastic microstructures, where the contrast in intragranular misorientations between porphyroclasts and neoblasts is well marked.

We also added arrows to mark the most important features, such as 120° triple junctions, subgrains, and pyroxene shapes, in the detail microphotographs in Fig. 4.

Fig. 7 Can you add to each sample the modal percentages of each phase (could be put for example below sample no. Note: If you divide up the porphyroclastic ones into two data sets each, you will have to plot them separately, I think this will be very informative and maybe a good way to look through the last deformation - so you would get a CPO history?! Would be worth checking out?!

We did not add the modal percentages to this figure (now Fig. 6) because it is already crowded. The modal percentages for all samples are presented in Table 1.

As required, we now present separately the olivine CPO for porphyroclasts (points) on top of the neoblasts CPO (contours) for the fine-porphyroclastic peridotites. This allows to clearly observe the orientation inheritance relations, but higher spread of the neoblasts orientations.

Figure 9. They are very small, I think you could put the spinel related figures into supplementary and make the other larger, then it is easier for the reader to see and relate to text.

Done. This is now Fig. 8 and spinel graphs are now Supplementary Material Fig. S1.

Figure 10. There is too much data on it - trends (or lack of) are hard to see. I suggest that you take our the literature as point data, but show them as fields that underlie your data. What are the open circles and squares? It becomes then much clearer that the fine porphyroclastic ones are distinct, while the other ones show a large range.

We now only present Macau peridotites data in this figure (now Fig. 9). Moreover, the use of colors, instead of gray level, allows as in the other figures, to better distinguish the present study data from the literature. In choosing the colors we paid attention that when printed in gray levels the symbols for the different microstructures could still be distinguished.

Hope this helps to make this manuscript even better,
Best wishes,
Sandra Piazzolo

28 Feb, 2019

Manuscript Number: TECTO13230

Title: The lithospheric mantle in a continental domain submitted to multiple tectonic events: Peridotite xenoliths from the Borborema Province, northeast Brazil

Authors: Liu et al.

This manuscript reported microtextures, CPO and major element compositions of mantle xenoliths from the Northern Borborema Province. The authors discussed P-T conditions and deformation-metasomatic history of the xenoliths, and finally discussed tectonic evolution of the studied mantle. The data quality is high, and interpretations and discussions are reasonably supported by their data coupled with previous data sets. The contents of the manuscript contain significant contributions to a link between petrological observations and geophysical interpretations. My decision for the present manuscript is, therefore, acceptable with minor revisions. To strengthen the manuscript, it might be better to discuss/describe two points before the authors go to discussions: (1) effects of infiltration of the host melt and (2) effects of heterogeneity of modal abundances for modal estimation.

1) Effects of infiltration of the host melt

Although the authors mentioned "Most xenoliths are small (on average 3 cm of diameter), but they are very fresh and display no macroscopic evidence for major interaction with the host lava. However, interstitial veins of altered glass, up to a few microns wide and often visible only by SEM, which could be sometimes followed to the contact with the basalt, were described in some xenoliths, such as AG6, AG7,"(307-311 lines), it is better to describe a bit more carefully what are fine-grained parts in deformed samples. According to Fig. 5., effects of infiltration of the host basalt are likely commonly observed (red arrows below).

The reviewer is 100% correct. The xenoliths are small and the effects of infiltration of the basalt, in particular close to the borders of the xenoliths, were understated in the submitted version of the ms. We have re-written this paragraph (lines 331-338) to clearly describe them and to state that the zones that displayed, under the microscope, any evidence for such reactions were carefully avoided in the chemical analyses and that this melt infiltration does not affect the microstructural data, because it concerns very small areas.

Especially, AG6, a deformed one, would be highly affected by infiltration of the host basalt. Please describe mineral assemblages in fine-grained part in deformed sample (AG6 for example), whether or not glass presents. Any hydrous minerals? If hydrous minerals are in the fine-grained parts, the discussions might be changed. I just want to make sure that the fine-grained parts were formed by deformation rather than melt-rock interactions. I almost believe that the former (deformation) would be a case for the studied samples.

The recrystallized domains in the fine-porphyroclastic peridotites are clearly distinct from the basalt infiltration ones. To better show this we added in the new Fig. 3 a detail EBSD phase map of such a domain in AG6, which clearly shows that the domain is composed solely by recrystallized olivine, with neither interstitial pyroxenes nor amphiboles that would have crystallized in response to melt rock interactions. Moreover, comparison of the olivine CPO of porphyroclasts and neoblasts, which are now plotted separately in the new Fig. 6 show that the two CPO are related, pointing to orientation inheritance, which is expected for recrystallization, but not for crystallization from a melt under static conditions.

2) Mineral mode estimation

Mineral mode is a key to estimate physical properties of the samples. However, because of large grain size and macroscopic heterogeneity in thin section scale, mineral modes in the manuscript is OK for the thin sections, but might need care to estimate representative mineral

mode for the mantle in the studied area. For instance, samples of CA14 and CA11 in Fig. 2 are very heterogeneous in mineral distributions. If we measure modal abundances of left half of the samples, mineral mode would be significantly different from the right half.

I would like to ask the authors to describe/discuss the effects of uncertainty in mineral mode for further discussions.

We agree that determination of mineral modes in such small samples is problematic, since mantle rocks are often compositionally heterogeneous. However, this is the best we can do given the size of xenoliths carried by these basalts. The only solution is to multiply the number of samples. Comparison with the present dataset with previous ones (which suffered from the same shortcomings) suggests, nevertheless, that the present sampling may be representative of the modal composition range in the lithospheric mantle in the region. Yet, because of the uncertainty in the determination of the modal composition of the individual samples, we only discuss the trends and use average samples for the calculation of seismic properties and comparison with the seismological data.

Minors:

266 lines typo

Significant digits for numbers in tables

Please consider significant digits for numbers in all tables. For instance, is it necessary to display the second decimal place for modal volume of minerals in Table 1?

Thanks for the remark. This was clearly our mistake. We have revised all tables to only show significant digits.

I hope these comments would improve the revised version of the manuscript.

Best regards, Tomo MORISHITA, Kanazawa University

Highlights:

- Mainly coarse-granular and porphyroclastic peridotites with [100](010) olivine textures
- Equilibrium temperatures consistent with a 80 km thick lithosphere in the Cenozoic
- Extensive annealing implies that deformation frozen in the shallow mantle is 100s Ma old
- Mantle fabric related to Neoproterozoic shear zones may partially explain SKS splitting
- Limited reworking of the lithospheric mantle during Cretaceous rifting

1 **Crust-mantle coupling during continental convergence and**
2 **break-up: Constraints from peridotite xenoliths from the**
3 **Borborema Province, northeast Brazil**

4

5

6

7

8 **Shiran Liu^{1,2}, Andréa Tommasi¹, Alain Vauchez¹, Maurizio Mazzucchelli³**

9

10

11

12 ¹Géosciences Montpellier, CNRS & Université de Montpellier, Montpellier,
13 France.

14 ²Key Laboratory of Orogenic Belts and Crustal Evolution, School of Earth and
15 Space Sciences, Peking University, Beijing, China.

16 ³Dipartimento di Scienze Chimiche e Geologiche, Università di Modena e
17 Reggio Emilia, Modena, Italy.

18

19 Corresponding author: Shiran Liu (hbhslsr@gmail.com), Andréa Tommasi
20 (andrea.tommasi@umontpellier.fr)

21

22

23

24

25 **Keywords:**
26 Subcontinental mantle lithosphere
27 Crust-mantle coupling
28 Melt-rock interaction
29 Olivine crystal preferred orientation
30 Seismic anisotropy

31

32 **Highlights:**

- 33 • Mainly coarse-granular and porphyroclastic peridotites with [100](010)
34 olivine textures
- 35 • Equilibrium temperatures consistent with a 80 km thick lithosphere in the
36 Cenozoic
- 37 • Extensive annealing implies that deformation frozen in the shallow mantle
38 is 100s Ma old
- 39 • Mantle fabric related to Neoproterozoic shear zones may partially explain
40 SKS splitting
- 41 • Limited reworking of the lithospheric mantle during Cretaceous rifting

42

43

44 **Abstract**

45 We studied a suite of mantle xenoliths carried by Cenozoic volcanism in the
46 Borborema Province, NE Brazil. These xenoliths sample a subcontinental
47 lithospheric mantle affected by multiple continental convergence and rifting
48 events since the Archean. Equilibrium temperatures indicate a rather hot
49 geotherm, implying a ca. 80 km thick lithosphere. Most xenoliths have
50 coarse-granular and coarse-porphyroclastic microstructures, recording
51 variable degrees of annealing following deformation. The high annealing
52 degree and equilibrated pyroxene shapes in coarse-granular peridotites
53 equilibrated at ~900°C indicate that the last deformation event that affected
54 these peridotites is several hundreds of Ma old. Coarse-porphyroclastic
55 peridotites equilibrated at 950-1100°C probably record younger (Cretaceous?)
56 deformation in the deep lithospheric mantle. In addition, a few xenoliths show
57 fine-porphyroclastic microstructures and equilibrium temperatures $\geq 1200^\circ\text{C}$,
58 which imply a recent deformation, probably related to the dykes that fed the
59 Cenozoic volcanism. Chemical and microstructural evidence for reactive
60 percolation of melts is widespread. Variation in textural and chemical
61 equilibrium among samples implies multiple melt percolation events well
62 spaced in time (from Neoproterozoic or older to Cenozoic). Crystal preferred
63 orientations of olivine and pyroxenes point to deformation controlled by
64 dislocation creep with dominant activation of the [100](010) and [001]{0kl} slip
65 systems in olivine and pyroxenes for all microstructures. Comparison of
66 xenoliths' seismic properties to SKS splitting data in the nearby RCBR station
67 together with the equilibrated microstructures in the low-temperature xenoliths

68 point to coupled crust-mantle deformation in the Neoproterozoic (Brasiliano)
69 continental-scale shear zones, which is still preserved in the shallow
70 lithospheric mantle. This implies limited reworking of the lithospheric mantle in
71 response to extension during the opening of the Equatorial Atlantic in the
72 Cretaceous, which in the present sampling is restricted to the base of the
73 lithosphere.

74

75 **1. Introduction**

76 Continental plates have long-lived histories. They are usually composed
77 by crustal domains with varied tectonic ages. Tectonic reworking is a common
78 feature in crustal exposures. Yet, major tectonic events such as convergence,
79 collision, and rifting involve the entire plate, not only the crust. By consequence,
80 large volumes of the subcontinental lithospheric mantle must have been
81 affected by a succession of tectono-thermal events (cf. reviews in Tommasi
82 and Vauchez, 2015; Vauchez et al., 2012). In addition, during orogenic events,
83 the strain regime frequently changes across the belt, with contiguous domains
84 accommodating transcurrent motions and others accommodating thrusting for
85 instance. However, how the deformation is accommodated in the lithospheric
86 mantle and the level of coupling between crust and mantle deformation during
87 these major tectonic episodes are still matter of debate.

88 Comparison between SKS splitting data and crustal structures points to
89 coherent kinematics, implying at least partial coupling between the crust and
90 the mantle in many orogenic belts (Tikoff et al., 2004). As recognized since the
91 early times of seismic anisotropy measurements (e.g., Vauchez et al., 1991),
92 crust-mantle coupling is well documented beneath large-scale strike slip faults
93 and transpressional belts, such as the San Andreas fault (Bonnin et al. 2012)
94 or the Pyrenees, the Appalachians, and the neoproterozoic Ribeira-Aracuai
95 belt of SE Brazil (cf. review in Vauchez et al., 2012). In contrast, in collisional
96 domains submitted to large amounts of thickening, like the Alps and the
97 Southern Tibet, or in active continental rifts, such as the East African rift
98 system, polarization directions of fast split SKS or SKKS waves are usually
99 parallel to the trend of the belts or rifts, but at high angle to the lineations
100 marking flow directions in the ductile crust and to the slip directions in active
101 faults (e.g., Barruol et al., 2011; Wu et al., 2015; Hammond et al., 2014). This
102 suggests at least partial decoupling between the crust and lithospheric mantle
103 (Tommasi et al., 1999).

104 However, seismic anisotropy data only offer indirect information on the
105 present-day structure of the upper mantle. Moreover, unless a variety of
106 methods based on waves with different incidence angles and propagation
107 directions are employed, localizing vertically the source of the anisotropy
108 remains difficult. Analysis of mantle xenoliths, which are mantle fragments
109 (peridotite xenoliths) carried to the surface by volcanic episodes, allows for
110 direct sampling of the lithospheric mantle. This sampling is imperfect: it is
111 discontinuous, non-oriented, and focused along the magmatic conduits.

112 Nevertheless, the characterization of the microstructures and associated
113 crystal preferred orientations, as well as of the chemical compositions and
114 mineralogy of the mantle xenoliths, allows determining the relative deformation,
115 annealing, and petrological (partial melting, reactive melt transport, solid-state
116 reequilibration) history, even if no dating is possible. Coupling petrophysical
117 analyses on mantle xenoliths to seismological data may allow better
118 constraining the deformation history of the lithospheric mantle and hence
119 discussing crust-mantle coupling during major tectonic events.

120 In this article, we present a petrostructural study of peridotite xenoliths
121 entrained by Cenozoic volcanism in NE Brazil. Based on these data and on
122 published SKS splitting measurements (Assumpção et al., 2011; Bastow et al.,
123 2015), we try to unravel the tectono-thermal evolution of the continental
124 lithospheric mantle in this region, which has been affected by multiple
125 collisional and extensional episodes since the Archean. Analysis of the
126 presently outcropping crustal structures highlights a series of
127 tectono-magmatic episodes, among which the most important and recent ones
128 are: (i) extensive intraplate deformation in a convergent setting, which
129 produced a continental-scale system of strike-slip shear zones and
130 transpressional belts accommodating lateral escape of this domain during the
131 formation of the Gondwana and (ii) extension, localized in a series of
132 intracontinental basins, during the early stages of the opening of the Equatorial
133 Atlantic.

134

135 **2. Geological and geophysical background**

136 *2.1 Evolution of the Northern Borborema Province (NBP)*

137 The analysis of the crustal rocks outcropping in the Northern Borborema
138 Province (NBP) points to a complex tectonic evolution, which probably started
139 in the Archean, as indicated by U-Pb ages ranging from 3.4 to 2.7 Ga recorded
140 in the São José Massif (Souza et al., 2016), ca. 100 km to the east of the study
141 area and in smaller nuclei elsewhere in the province. Between 2.1-2.4 Ga, a
142 major crust formation event produced ca. of the 50% of the present-day
143 outcropping rocks (Hollanda et al., 2011; Souza et al., 2016). The tectonic
144 setting for this extensive magmatic activity is discussed, but isotopic data
145 indicates significant recycling of pre-existing crustal material (Hollanda et al.,
146 2011).

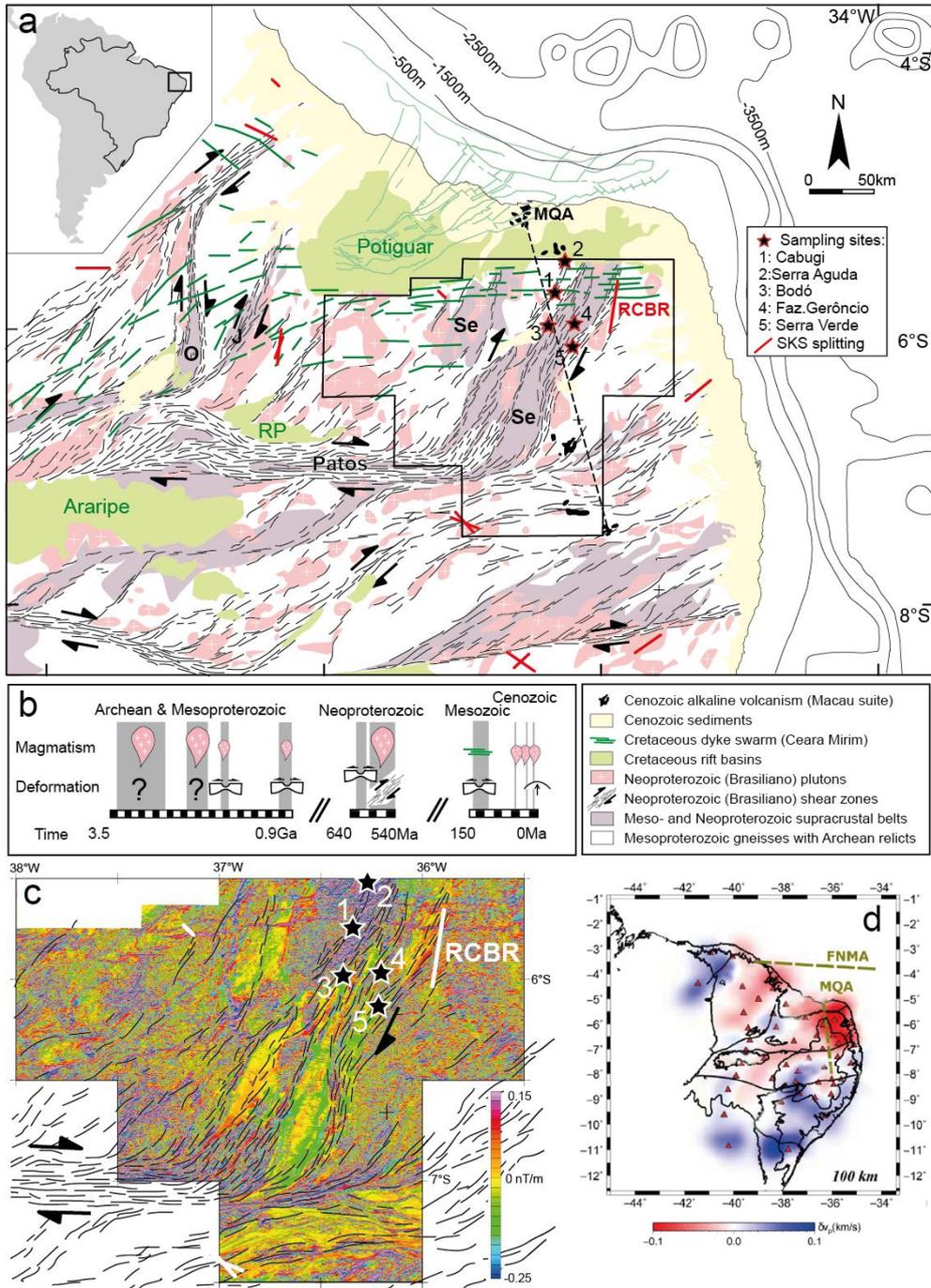
147 The Meso- to Neoproterozoic evolution of the Borborema Province is
148 characterized by a series of failed intracontinental rifting episodes. At 1.8-1.9
149 Ga, the Orós and Jaguaribe volcano-sedimentary basins formed to the west of
150 the study area (Fig. 1). Localized extension was accompanied by intrusion of
151 small volumes of alkaline granites and anorthosites elsewhere in the province,
152 which extended until 1.75 Ga (Sa et al., 1995; Hollanda et al., 2011). At 1-0.9
153 Ga, extension affected the central Borborema Province (Santos et al., 2010;
154 Van Schmus et al., 2008). The last Proterozoic extensional episode affected
155 most of the province and formed the Seridó basin within the study area (Fig. 1).

156 The youngest sedimentation in this basin occurred at ~630-615 Ma, but early
157 sedimentation may be as old as 1.8 Ga (Hollanda et al., 2015), suggesting an
158 early rifting episode simultaneous with the formation of the Orós and Jaguaribe
159 basins.

160 The major tectono-magmatic episode that affected the Borborema
161 Province, shaping its present-day geology, is, however, the compressional
162 Brasiliano event between 595-540 Ma (Neves, 2003; Archanjo et al. 2008;
163 Viegas et al., 2014). This event resulted in extensive tectono-thermal
164 reworking of the province through the formation of a continental-scale
165 transcurrent shear zone system – the Borborema shear zone system,
166 transpressional inversion of the Proterozoic metasedimentary basins, and
167 widespread emplacement of granitoids (Fig. 1). The granitoids geochemistry
168 indicates a dominant component of crustal reworking (Neves, 2003; Souza et
169 al., 2016), implying an intraplate setting for the Brasiliano deformation, with a
170 possible convergent plate boundary > 500 km to the NW from the study area
171 (e.g., Caby et al., 1991; Ancelmi et al., 2015). The complex network of
172 anastomosing E-W and NE-SW trending ductile dextral strike-slip shear zones,
173 some up to ~25 km wide, and transpressional belts deformed under high
174 temperature, low pressure conditions that compose the Borborema shear zone
175 system (Vauchez et al., 1995) is therefore the expression of strain localization
176 in an intraplate setting controlled by large-scale intraplate rheological
177 heterogeneities (the basins) in response to the convergence between São
178 Francisco, Amazonian, and West African cratons during the construction of the
179 Gondwana (Tommasi et al., 1995). This shear zone system allowed for
180 tectonic extrusion of the Borborema Province towards the NE (Ganade de
181 Araújo et al., 2014, 2016), that is, towards the Saharan province, which, like
182 Borborema, had been weakened by tectono-magmatic activity during the
183 Meso- and Neoproterozoic.

184 In the early Cretaceous, the Borborema province experienced regional
185 extension before rifting localized at the present location of the equatorial
186 Atlantic Ocean (Darros de Matos, 1999). Structures associated with these
187 early stages of rifting in the northern Borborema Province include several
188 intracontinental basins or aborted rifts and a large tholeiitic dyke swarm. The
189 basins formed as NW-SE trending asymmetric grabens (e.g., Potiguar, Rio do
190 Peixe, Araripe in Fig. 1) controlled by reactivation of the Brasiliano shear zones
191 (Darros de Matos, 1999; Castro et al., 2007, 2012; Marques et al., 2014). They
192 contain dominantly continental sediments of fluvial and lacustrine origin,
193 which attain thicknesses of 2000 m, deposited between 145-125 Ma. The
194 Potiguar basin is the best developed among these basins; its offshore domain
195 is the only one involved in the final opening of the Atlantic. The Ceará-Mirim
196 dyke swarm is a >350-km-long tholeiitic dyke swarm with a dominantly E-W
197 trend that rotates to NE-SW at the southern termination of the Potiguar basin
198 (Fig. 1). The dykes have ages between 135 Ma and 120 Ma (Hollanda et al.,
199 2018 and references therein). Their emplacement is therefore synchronous to

200 the formation of the intracontinental basins. Geochemical data points to
 201 primary magmas formed by melting of an enriched mantle ($EMI \pm FOZO$
 202 isotopic signatures) at shallow depths (<90 km; Ngonge et al., 2016).
 203



204
 205 Fig. 1. (a) Simplified geological map (after Vauchez et al., 1995 and Oliveira and Medeiros, 2018) displaying xenolith
 206 sampling locations. (b) Time line of the major tectono-magmatic events in the Borborema Province. (c) Aeromagnetic
 207 anomalies map (1st vertical derivative; Costa et al., 2016) imaging middle crust structures in and around the Seridó belt.
 208 (d) P-wave tomography model for the Borborema Province at 100 km depth (Simões Neto et al., 2018) characterized
 209 by a marked low velocity anomaly east of the Macau-Queimadas volcanic alignment (MQA) and north of the Patos
 210 shear zone. SKS splitting data in (a) and (c) from Assumpção et al. (2011) and Bastow et al. (2015). RP: Rio do Peixe

211

Cretaceous basin. Se: Seridó, O:Orós, and J: Jaguaribe supracrustal belts.

212 The Cenozoic is characterized by a long-lived, but small-volume alkaline
213 volcanism (both onshore and offshore), tectonic inversion of the Mesozoic
214 intracontinental basins, and uplift of the Borborema Plateau. The entire
215 Borborema Province sits nowadays on average at ~800m above the sea level
216 and the Araripe basin is topographically more than 500 m above the
217 Precambrian basement. Analysis of river profiles implies that most uplift
218 occurred within the last 30 Ma (Tribaldos et al., 2017), but earlier events may
219 have occurred (Marques et al., 2014; Nogueira et al., 2015). The present day
220 crustal thickness of 30-35 km implies some degree of sub-lithospheric support
221 for the uplift. The offshore Cenozoic volcanism forms an E-W volcanic
222 alignment that extends from the Ceará coast in northeast Brazil (the Mecejana
223 volcanic field) to the Fernando de Noronha archipelago. The Mecejana
224 volcanics yield K-Ar ages between 26 and 44 Ma (Mizusaki et al., 2002),
225 whereas volcanism in Fernando de Noronha has Ar-Ar ages between 12.5 and
226 6 Ma (Perlingeiro et al., 2013). The onshore Cenozoic volcanism is distributed
227 along a N-S trend, forming the Macau–Queimadas Alignment (MQA, Fig. 1).
228 Emplacement ages range between 52 Ma and 7 Ma, with two peaks around
229 26-29 Ma and 7-9 Ma and no clear age progression along the MQA (Souza et
230 al., 2003; Silveira 2006; Knesel et al., 2011).

231 The Macau volcanics that entrained the studied mantle xenoliths (Fig. 1)
232 erupted onto metasediments from the Neoproterozoic Seridó belt or
233 Mesoproterozoic gneisses that form their basement. In the study area, the
234 Brasiliano event is recorded by multiple NE-SW shear zones, which branch off
235 from the E-W Patos shear zone in the south, by transpressional structures in
236 the Seridó belt, and by intrusion of granitoids. The Cretaceous Ceará-Mirim
237 dyke swarm crosscuts the study area with an E-W trend and the Potiguar basin
238 outcrops to the north and west of it (Fig. 1).

239

240 *2.2 Geophysical data*

241 P-wave receiver functions, deep seismic refraction experiments, and
242 surface-wave dispersion data indicate that the crust is 30 ~ 32km thick in the
243 northern Borborema Province, but 36 ~ 38 km thick in the southern part of the
244 province (Oliveira and Medeiros, 2012; Almeida et al., 2015; Lima et al., 2015;
245 Luz et al., 2015). This variation in crustal thickness is consistent with gravity
246 data (Oliveira and Medeiros, 2018) and correlated with topography: elevated
247 regions show thicker crust. Receiver function data also imply an intra-crustal
248 discontinuity at 9-18 depth, most often observed in domains with thinner crust,
249 such as the area sampled by the studied xenoliths (Almeida et al., 2015).

250 Gravity and magnetic data illuminate the structuration of the deep crust in
251 the Borborema province (Oliveira and Medeiros, 2018). These data establish
252 the Patos, Pernambuco, and Jaguaribe shear zones as major structures,
253 which splay off from the Transbrasiliano Lineament, separating four crustal
254 blocks with different geophysical characteristics: the Southern, the Transversal,

255 the Ceará, and the Rio Grande do Norte domains. The studied xenoliths
256 sample the mantle beneath the Rio Grande do Norte domain, which is
257 delimited by the Jaguaribe and Patos shear zones to the west and south, and
258 by the Atlantic Ocean to the north and east. This domain is characterized by
259 strong magnetic contrasts with E-W or NNE trends, correlated with the
260 outcropping Brasiliano structures (Fig. 1c). Magnetic anomalies also clearly
261 define the Ceará-Mirim dykes (Fig. 1c). In addition, analysis of the Bouguer
262 gravity anomaly shows that most of the Rio Grande do Norte domain is
263 characterized by weak long wavelength positive anomalies, which follow the
264 trend of the coastline (cf. Fig. 7 of Oliveira and Medeiros, 2018), suggesting
265 that crustal thinning related to the Atlantic opening affected most of the
266 domain.

267 The first kilometers of the lithospheric mantle structure may be probed by
268 the analysis of Pn velocities, which decrease from 8.1 km/s to 7.9 km/s in a
269 NW-SE refraction profile across the Borborema Province to the south of the
270 study area (Lima et al., 2015). In the vicinity of the study area, Pn velocities are
271 ~8.0 km/s, which for an isotropic mantle with a spinel-lherzolite composition
272 would correspond to sub-Moho temperatures of 700-750°C (Schutt et al.,
273 2018).

274 No seismic velocity data is available for the deeper sections of the
275 lithospheric mantle, but a recent P-wave regional travel-time tomography
276 model (Simões Neto et al., 2019) imaged slower than average velocities at
277 depths <150 km in the northeastern Borborema province, just east of the study
278 area (Fig. 1d). This low velocity anomaly coincides with a local geoid anomaly
279 of +10 m (Ussami et al. 1999). Slower than average P-wave velocities in the
280 shallow mantle beneath the study area are coherent with a rather shallow
281 Lithosphere Asthenosphere boundary (LAB), estimated at 80 km depth based
282 on S receiver function data from the station RCBR (Heit et al., 2007). The
283 results of the regional P-wave tomography for the northeastern Borborema
284 Province are also consistent with global finite-frequency tomography models,
285 which image lower than average S-wave velocities in the mantle at 250-km
286 depth in this region (French et al. 2013).

287 SKS splitting data in the Borborema Province (Bastow et al., 2015) show a
288 complex pattern, with highly variable delay times and fast polarization
289 directions, which in some places closely follow the Neoproterozoic crustal
290 fabric, but are oblique to it in others (Fig. 1a,c). The station closest to the
291 xenolith sampling sites, RCBR (Fig. 1a,c), shows a NNE-oriented fast S-wave
292 polarization subparallel to the Brasiliano transpressive structures, which
293 structure the northern Seridó belt and its gneissic basement, and a high delay
294 time (1.9 ± 0.2 s) based on 9 individual measurements (Assumpção et al.,
295 2011).

296

297 2.3 Previous studies of Macau peridotite xenoliths

298 The petrology and the geochemistry of Macau mantle xenoliths were

299 studied by Comin-Chiaramonti et al. (1986), Princivalle et al. (1994), Fodor et
300 al. (2002), Rivalenti et al. (2000; 2007). These studies identified two
301 microstructural groups: protogranular (the dominant type) and (rare)
302 porphyroclastic (partially recrystallized) peridotites. Protogranular
303 microstructures are mainly observed in Iherzolites, whereas porphyroclastic
304 microstructures are usually observed in harzburgites. Protogranular peridotites
305 also yield lower two-pyroxene equilibrium temperatures ($825\pm 116^{\circ}\text{C}$) than
306 porphyroclastic ones ($>1150^{\circ}\text{C}$). Trace-element patterns and isotopic
307 compositions of the peridotites indicate variable degrees of metasomatism by
308 alkali-basaltic melts with EMI and minor EMII isotopic signatures, probably
309 during multiple metasomatic events (Rivalenti et al., 2000, 2007).

310

311 **3. Methods**

312 *3.1. Sampling*

313 We have studied 22 xenoliths from 5 different Macau volcanic sites, which
314 extend in a rough N-S trend, from the limit of the Potiguar basin to ~70 km
315 south of it (Fig. 1). Most xenoliths analyzed in the present study come from the
316 Pico do Cabugi basaltic neck ($5^{\circ}42.3'S$, $36^{\circ}19.4'W$). The Pico do Cabugi
317 basalts yield Ar-Ar ages 24.6 ± 0.8 Ma (Souza et al., 2003 and references
318 therein). Among the 16 xenoliths of the Pico do Cabugi in this study, 14
319 (marked as CA) have been collected by the authors in a field campaign in 2016
320 and two (marked as PC) were previously studied by Rivalenti et al. (2000).

321 We also analyzed 6 xenoliths previously studied by Rivalenti et al. (2007):
322 two from the Serra Aguda neck (AG, $5^{\circ}31'S$, $36^{\circ}17'W$), ~30 km north of the
323 Pico do Cabugi, at the border of the Potiguar basin (Fig. 1), one from the Serra
324 Preta do Bodó dykes (BO, $5^{\circ}58'S$, $36^{\circ}22'W$), ~20 km south of Pico do Cabugi,
325 and four from Fazenda Geroncio (GR, $5^{\circ}58'S$, $36^{\circ}14'W$) and Serra Verde
326 localities (SV, $6^{\circ}06'S$, $36^{\circ}12'W$), located ~30 and ~50 km south of Pico do
327 Cabugi (Fig. 1). There are no ages for the Fazenda Geroncio and Serra Verde
328 volcanics, but the Serra Aguda basalts were dated at 26 Ma (Silveira, 2006)
329 and those at Serra Preta do Bodó, at 7.1 Ma (Knesel et al, 2010).

330 Most xenoliths are small (on average 3 cm of diameter), but they are very
331 fresh. The samples display no macroscopic evidence for major interaction with
332 the host lava. However, small pockets of fine-grained clinopyroxene, spinel,
333 and olivine, as well as melt films along grain boundaries were observed under
334 the microscope, in particular close to the borders of many xenoliths. In addition,
335 some xenoliths have pyroxenes with spongy borders, indicative of limited
336 partial melting during extraction. These domains were carefully avoided during
337 microprobe analyses. They were sometimes included in the EBSD analyses,
338 but affect weakly the results because of their small area.

339

340 *3.1 Electron-backscattered diffraction (EBSD) data acquisition and treatment*

341 Petrostructural analyses were performed on all samples. Crystallographic
342 preferred orientations (CPOs) of olivine, pyroxenes, and spinel were measured

343 by indexing of electron-backscattered diffraction patterns in the SEM-EBSD
344 facility at Geosciences Montpellier, France. Data acquisition was performed
345 using a JEOL JSM 5600 scanning electron microscope with 17kV acceleration
346 voltage and 24mm working distance. We performed EBSD mapping over the
347 entire thin section (areas $\geq 2 \times 3 \text{ cm}^2$) with step sizes between 14 μm and 35 μm ,
348 depending on grain size. Indexation rates varied between 84% and 97%.
349 Non-indexed points correspond mainly to fractures. During post-acquisition
350 data treatment, we eliminated inaccurate indexing points ($\text{MAD} > 1.3^\circ$ and wild
351 spikes), corrected for rare olivine pseudo-symmetry misindexing, and filled up
352 non-indexed pixels with at least 6 coherent neighboring measurements with
353 the average of the neighbors' orientations.

354 The CPO data analysis, that is, the calculation of the orientation
355 distribution functions (ODF) and of the misorientations, the computation of the
356 strength and symmetry of the olivine CPO, the plotting of pole figures, and the
357 quantification of the microstructure (grains' size, shape, and orientation) was
358 performed using the MTEX toolbox in Matlab (<http://mtex-toolbox.github.io/>;
359 Hielscher and Schaeben, 2008; Bachmann et al., 2010; Bachmann et al.,
360 2011). The ODFs were calculated using a "de la Vallée Poussin" kernel
361 function with a half-width of 10° . The CPO is presented as one crystallographic
362 orientation per pixel in pole figures (lower hemispheric stereographic
363 projections). Thin sections were cut in random orientations, because of the
364 small size of the xenoliths. To facilitate comparison between samples, we
365 rotated the CPO of all samples into a common reference frame, in which the
366 maximum concentration of [100] of olivine is parallel to the E-W direction and
367 the maximum concentration of [010] axis of olivine is parallel to the N-S
368 direction of the pole figure. The choice of this reference frame can be justified
369 a-posteriori by the analysis of the CPO, which indicates that this reference
370 frame probably corresponds to the lineation and normal to the foliation
371 directions (cf. discussion).

372 The strength of the CPO was quantified using the dimensionless J-index
373 (Bunge, 1982). The J-index for a random CPO is 1; it can reach up to 20 in
374 natural peridotites, with a peak at 3-5 (Tommasi and Vauchez, 2015). We use
375 the dimensionless BA-index, based on the eigenvalues of the [100] and [010]
376 axes orientation distributions (Mainprice et al., 2014), to characterize the
377 olivine CPO symmetry. This index allows classifying the olivine CPO into three
378 types: (i) fiber-[010], characterized by a [010] point maxima and girdle
379 distributions of [100] and [001] (BA-index < 0.35), (ii) orthorhombic,
380 characterized by point maxima of [100], [010], and [001]
381 ($0.35 < \text{BA-index} < 0.65$) and (iii) fiber-[100], characterized by a point maxima
382 of [010] and girdle distributions of [010] and [001] (BA-index > 0.65). J- and
383 BA-indexes do not depend on the reference frame.

384 To analyze quantitatively the microstructure, we used the grain detection
385 method in MTEX (Bachmann et al., 2011) defining grains boundaries by
386 misorientation angles between adjacent pixels higher than 15° . Grains

387 composed by less than 10 pixels were not considered in the microstructural
388 analysis. The misorientation of each pixel relative to the mean orientation of
389 the grain (M2M) and the grain orientation spread (GOS) were calculated to
390 quantify the intracrystalline orientation gradients, which are a proxy of the
391 dislocation density. In addition, we characterized the sinuosity of the grain
392 boundaries by the shape factor, which is the ratio of the perimeter of the grain
393 over the perimeter of a circle with the same area. Both the intra-granular
394 misorientation (M2M and GOS) and the sinuosity of grain boundaries (shape
395 factor) should decrease in response to recrystallization.

396

397 *3.2 Mineral compositions and equilibrium temperatures*

398 Mineral compositions of olivine, orthopyroxene, clinopyroxene, and spinel
399 were analyzed in ten samples, selected based on their microstructure, so that
400 all microstructural types were represented. Measurements were performed in
401 a Cameca SX100 electron microprobe at the Microsonde Sud facility at the
402 University of Montpellier, France at a 20kV accelerating voltage and a 10nA
403 current. For each sample, we analyzed both rim and core composition in 3 to 4
404 grains for olivine, orthopyroxene, clinopyroxene, and spinel.

405 We calculated equilibrium temperatures based on the
406 clinopyroxene-orthopyroxene geothermometer (Taylor, 1998) and on the
407 Ca-in-opx geothermometer (Brey and Kohler, 1990; revised by Nimis and
408 Grutter, 2010). Average rim and core equilibrium temperatures were calculated
409 by averaging the temperatures calculated using 3-4 rim or core compositions
410 of individual opx grains or opx-cpx pairs for each sample. Since no reliable
411 barometers are available for spinel-facies peridotites, we chose arbitrarily 1.5
412 GPa (~46 km depth) as the pressure to calculate the two thermometers.
413 Changes in the assumed pressure produce a variation of ~10 °C per 0.5 GPa
414 (~11 km).

415

416 *3.3 Seismic properties*

417 Seismic properties of each sample were computed using the MTEX
418 toolbox (Mainprice et al., 2011), using Voigt-Reuss-Hill averaging based on the
419 CPOs and modal content of olivine, orthopyroxene, and clinopyroxene derived
420 from the EBSD maps and on the single crystal elastic constant tensors of the
421 three minerals and their temperature and pressure derivatives (Abramson et
422 al., 1997; Anderson et al., 1992; Chai et al., 1997; Isaak et al., 2006; Jackson
423 et al., 2007; Sang and Bass, 2014). Average seismic properties for the
424 lithospheric mantle beneath the study area were estimated by averaging the
425 elastic constant tensors of all samples with all CPO data rotated into a
426 common reference frame. The assumption is that the orientation of the
427 lineation and foliation is constant both laterally and vertically. **The average
428 seismic properties provide therefore an estimate of the maximum seismic
429 anisotropy that could be produced, if the xenolith sampling is representative of
430 the variability of compositions and CPO in the lithospheric mantle beneath the**

431 studied area.

432

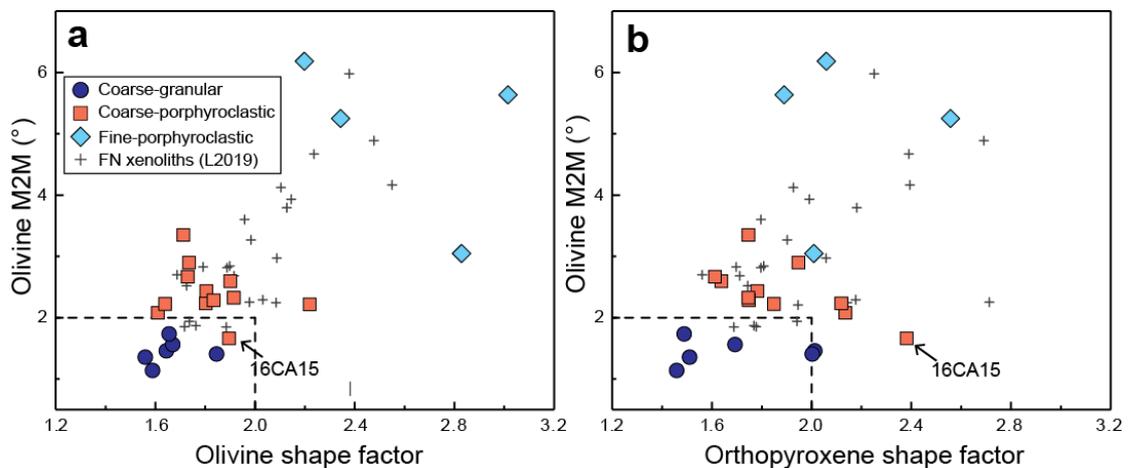
433 4. Data

434

435 4.1 Microstructures

436 Previous studies (Fodor et al., 2002; Rivalenti et al., 2000; Rivalenti et al.,
437 2007) have correlated optical observations of the microstructures with
438 chemical compositions and equilibrium temperatures to divide the Macau
439 peridotite xenoliths into two groups: protogranular and porphyroclastic. In the
440 present study, we associated to the optical observations quantitative analysis
441 of the microstructural data derived from EBSD mapping. Based on these data,
442 in particular the intragranular misorientation of olivine and the olivine and
443 orthopyroxene grain shapes (Fig. 2), we classify the Macau peridotites into
444 three microstructural groups (Table 1): coarse-granular (6),
445 coarse-porphyroclastic (12), and fine-porphyroclastic (4). The coarse-granular
446 and coarse-porphyroclastic groups correspond to a continuous variation in
447 microstructure (Figs. 2 and 3). They were described as protogranular in the
448 previous studies. The fine-porphyroclastic group is clearly different from the
449 other two groups (Figs. 2 and 3); it was described as porphyroclastic in the
450 previous studies.

451



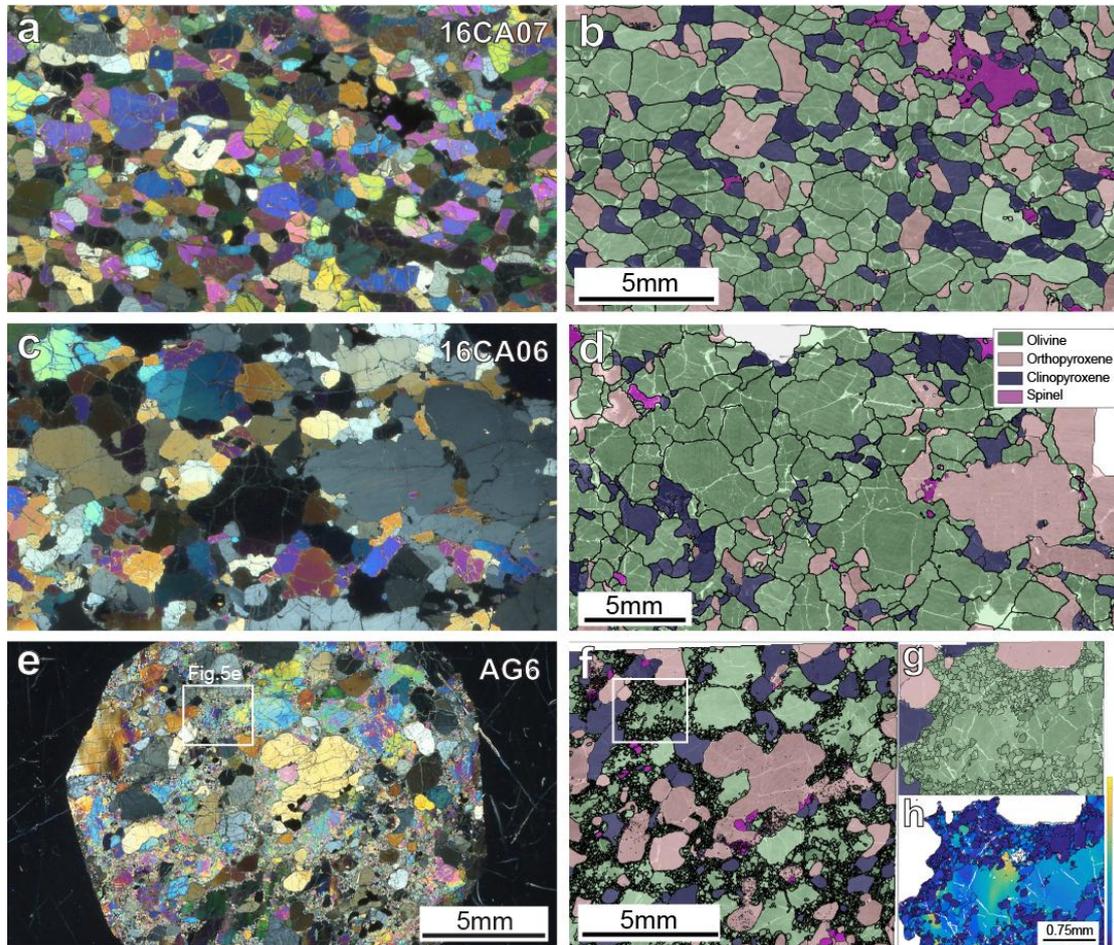
452

453 Fig. 2. Microstructural parameters calculated from EBSD data: (a) Olivine shape factor vs. intragranular misorientation
454 relative to the mean orientation of the grain (M2M); (b) Orthopyroxene shape factor vs. olivine M2M. All quantities are
455 average values at the sample scale weighted by the grains' area (cf. Table 2). Similar data for Fernando de Noronha
456 (FN) xenoliths, which sample an old domain of the equatorial Atlantic (Liu et al., 2019), are presented for comparison.

457

458 There is no relation between microstructure and sampling site. All three
459 microstructures are represented among the Pico do Cabugi peridotites, which
460 is the best-sampled site in the present study, and in at least one of the other
461 sampling localities (Table 1). This observation is corroborated by the previous
462 studies, which analyzed a larger number of peridotite xenoliths from other
463 Macau volcanic centers (Fodor et al. 2002; Rivalenti et al. 2007). Coarse

464 granular and coarse porphyroclastic microstructures predominate in all sites.
 465 Fine-porphyroclastic peridotites are always minor. They were only recovered in
 466 three sites, which are nevertheless >60 km apart (Fig. 1): Serra Aguda, Cabugi,
 467 and Serrote Preto (a small plug ca. 30 km to the SSE of Pico do Cabugi, Fodor
 468 et al. 2002).
 469



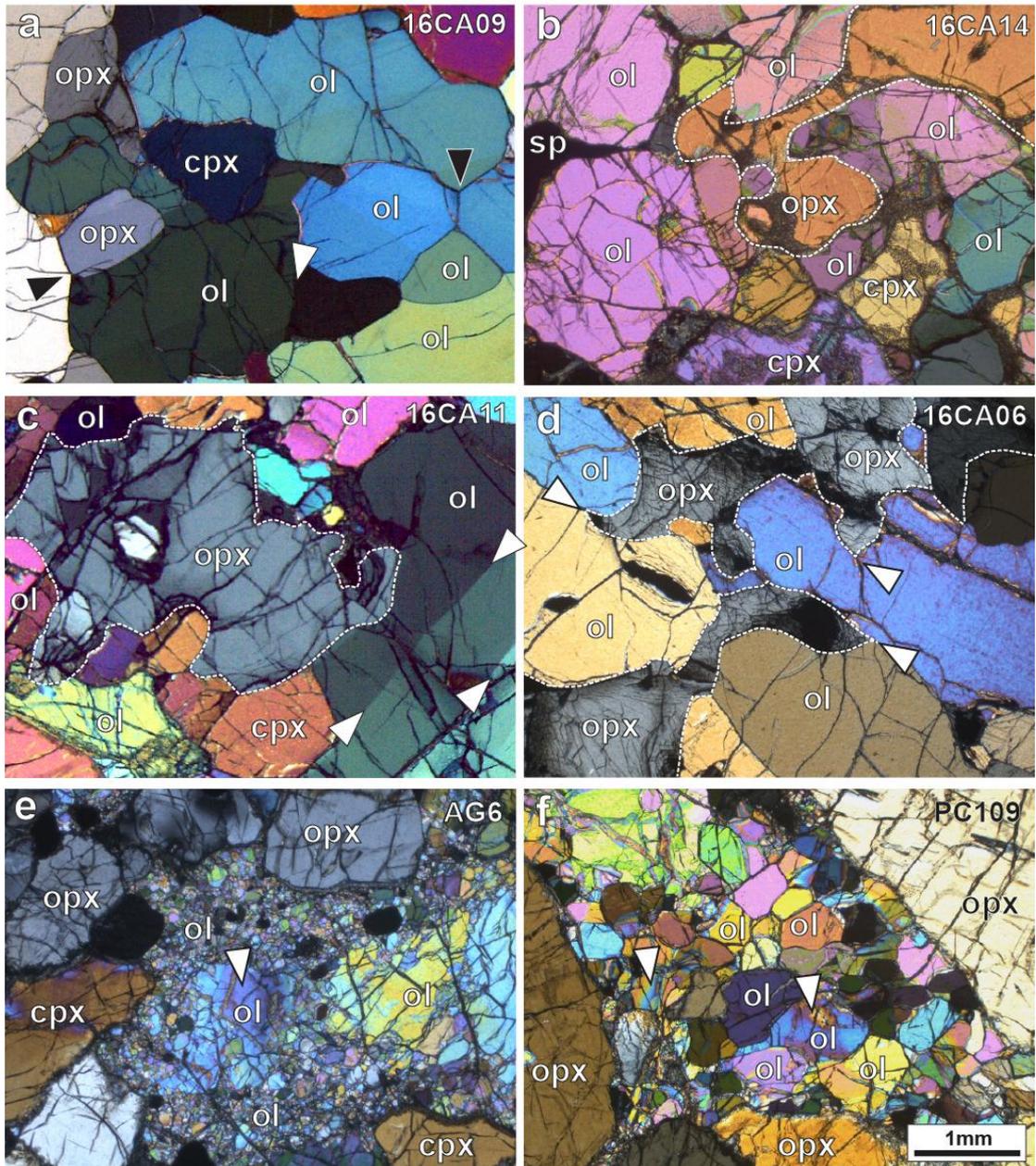
470
 471
 472 Fig. 3. Typical microstructures of Macau peridotite xenoliths: (a-b) coarse-granular, (c-d) coarse-porphyroclastic, and
 473 (e-f) fine-porphyroclastic. (a,c,e) Photomicrographs in cross-polarized light; scale bar is the same in all
 474 photomicrographs. (b,d,f,g) EBSD phase maps, which better illustrate the variation in shape of olivine and pyroxenes
 475 between the three microstructures. (g) Detail of a partially recrystallized domain in fine-porphyroclastic lherzolite AG6.
 476 (h) Misorientation relative to the mean orientation of the grain (M2M) map illustrating the difference in intragranular
 477 misorientation in olivine between porphyroclasts (high M2M) and neoblasts (low M2M) in the same domain.

478
 479 Coarse-granular peridotites are characterized by roughly equigranular
 480 microstructures (Fig. 3a-b). Olivine and orthopyroxene pyroxene grains are on
 481 average 1-3 mm, clinopyroxene slightly smaller: 0.8-1 mm (Table 2). Both
 482 olivine and pyroxenes have curvilinear to straight grain boundaries, evolving
 483 locally into polygonal shapes with 120° triple junctions (Fig. 4a), and very low
 484 densities of intracrystalline deformation features, such as undulose extinction,
 485 subgrains, or kinks (Fig. 3a and 4a). Quantitatively, these observations

486 translate into low shape factors ≤ 1.85 for olivine and ≤ 2 for orthopyroxene
487 and into low M2M values $\leq 2^\circ$ for olivine (Fig. 2). Lherzolite 16CA14 has an
488 intermediate microstructure characterized by polygonal olivine grains, but
489 pyroxenes with irregular shapes (Figs. 2 and 4b). Most coarse-granular
490 peridotites display a weak olivine shape-preferred orientation (SPO) (Fig. 3a),
491 but BO09 has higher olivine aspect ratios (Table 2) and a clear olivine SPO,
492 which results in a tabular microstructure. Pyroxene grains usually show no
493 exsolutions (Fig. 3a and 4a-b). Spinel grains have holy-leaf or interstitial
494 shapes. Coarse holy-leaf spinels may contain rounded inclusions of olivine or
495 pyroxenes (Fig. 3b).

496 Coarse-porphyroclastic microstructures are characterized by variable
497 grain sizes. In most cases, the grain size variation is continuous and it is
498 difficult to discriminate between porphyroclasts and neoblasts (Fig. 3c-d).
499 Olivine has more irregular grain shapes and clear evidence for intracrystalline
500 plastic deformation (Fig. 3c-d and 4c). Olivine grains have shape factors
501 between 1.6 and 2.2, consistently with the more sinuous grain boundaries, and
502 higher M2M values ranging from 2° to 3.4° (except 16CA15), consistent with
503 the higher frequency of undulose extinction and subgrain boundaries (Fig. 2).
504 Orthopyroxene grains also have irregular shapes (Fig. 3c-d and 4c-d), which
505 are associated with shape factors from 1.6 to 2.4 (Fig. 2b). Orthopyroxene
506 grain boundaries often show embayments or cusped shapes at the contact
507 with olivine grains (Fig. 4c-d). Clinopyroxene grains also show irregular or
508 interstitial shapes (Fig. 3c-d). In some samples, orthopyroxene grains display
509 exsolutions (Fig. 3c). Undulose extinction and kinks are observed locally.
510 Spinel usually has irregular shapes and occurs in association with ortho- or
511 clinopyroxene (Fig. 3d).

512 Fine-porphyroclastic peridotites show a well-developed bimodal olivine
513 grain size distribution, characterized by coexistence of coarse porphyroclasts
514 with serrated grain boundaries, undulose extinction, and closely-spaced
515 subgrain boundaries with a recrystallized matrix composed by polygonal
516 olivine neoblasts free of intracrystalline deformation (Figs. 3e-h and 4e-f). The
517 average size of the recrystallized grains varies between samples (Table 2): it is
518 $\sim 100 \mu\text{m}$ in AG6 (Fig. 4e), but $>300 \mu\text{m}$ in the other fine-porphyroclastic
519 peridotites (Fig. 4f, Table 2). It is noteworthy that the recrystallized olivine
520 grains are not organized in a planar mode, marking a foliation, but either form
521 irregular pockets or vein-like structures (Fig. 3e-f). Orthopyroxene is usually
522 coarser than clinopyroxene (Fig. 3f). Exsolutions were not observed in this
523 group. Shape factors of orthopyroxene vary between 1.89 and 2.56 (Fig. 2). In
524 AG6 and AG7, pyroxenes have rounded shapes (Fig. 3f), whereas in PC105
525 and PC109 fine-porphyroclastic samples, pyroxenes have irregular shapes,
526 with cusp-like terminations when in contact with olivine, similarly to pyroxenes
527 in coarse-porphyroclastic peridotites. Spinel occurs as fine rounded grains (Fig.
528 3f).



529

530

531

532

533

534

535

536

537

538

539

540

541

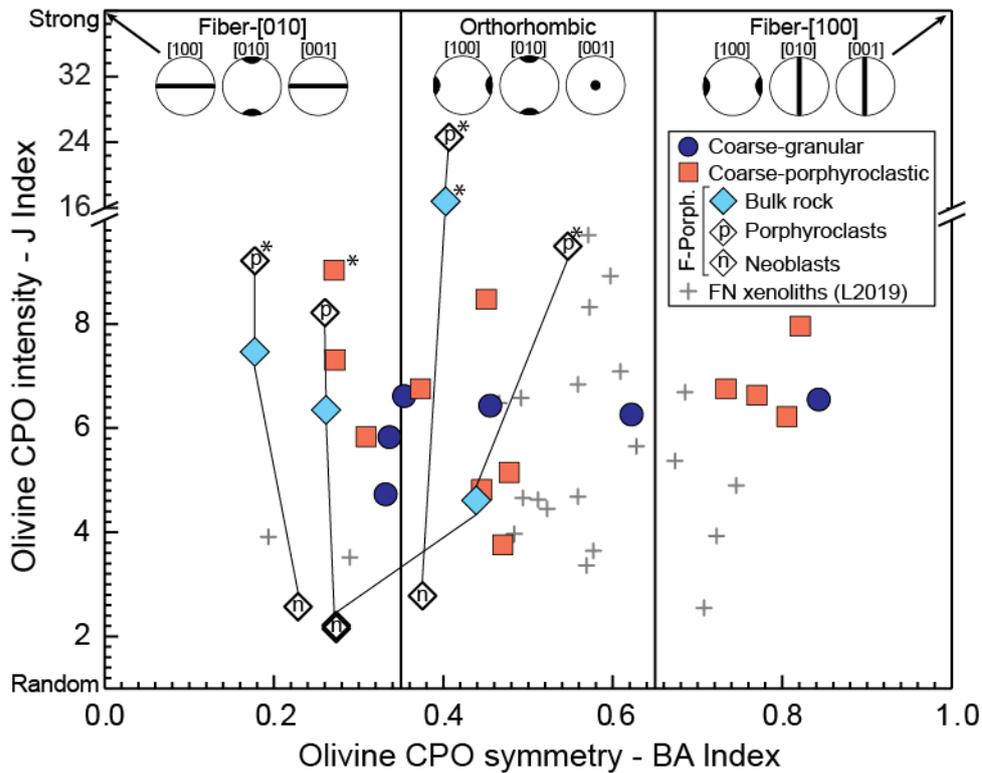
542

Fig. 4. Photomicrographs in cross-polarized light displaying typical features for the three microstructural groups. (a) Coarse granular lherzolite displaying polygonal grains (black arrow marks a 120° triple junction) almost free of intracrystalline deformation features, white arrow indicates a relict subgrain boundary in olivine (ol); (b) Coarse granular lherzolite 16CA14 displaying polygonal strain-free olivine grains, but irregularly-shaped orthopyroxene (opx) grains (highlighted by dashed line). (c-d) Coarse porphyroclastic lherzolites 16CA11 and 16CA06 showing irregularly shaped olivine, orthopyroxene (highlighted by dashed line), and clinopyroxene (cpx) grains with interpenetrating interphase boundaries, white arrows indicate subgrain boundaries in ol in (c) and mark cusp shaped opx in contact with ol in (d); (e) Fine-porphyroclastic lherzolite AG6 showing a matrix of very fine recrystallized olivine neoblasts surrounding a coarse olivine porphyroclast with undulose extinction and subgrains (white arrow); (f) Fine-porphyroclastic harzburgite PC109 showing tabular, strain-free olivine neoblasts as well as relicts of olivine porphyroclasts with undulose extinction (white arrow) within a strongly recrystallized domain between opx porphyroclasts. Scale bar is the same in all photomicrographs.

543 4.2 Crystallographic preferred orientations (CPO)

544 Quantitative data on the intensity and symmetry of the olivine CPO (J- and
 545 BA-indexes) as well as pole figures of the olivine, orthopyroxene, and
 546 clinopyroxene CPOs for all studied samples are presented in Figs. 5 and 6 and
 547 Table 2. For the samples in which a foliation was observed, the latter is
 548 displayed on the pole figures (Fig. 6). In these samples, the olivine [010]
 549 maximum is normal to the foliation.

550



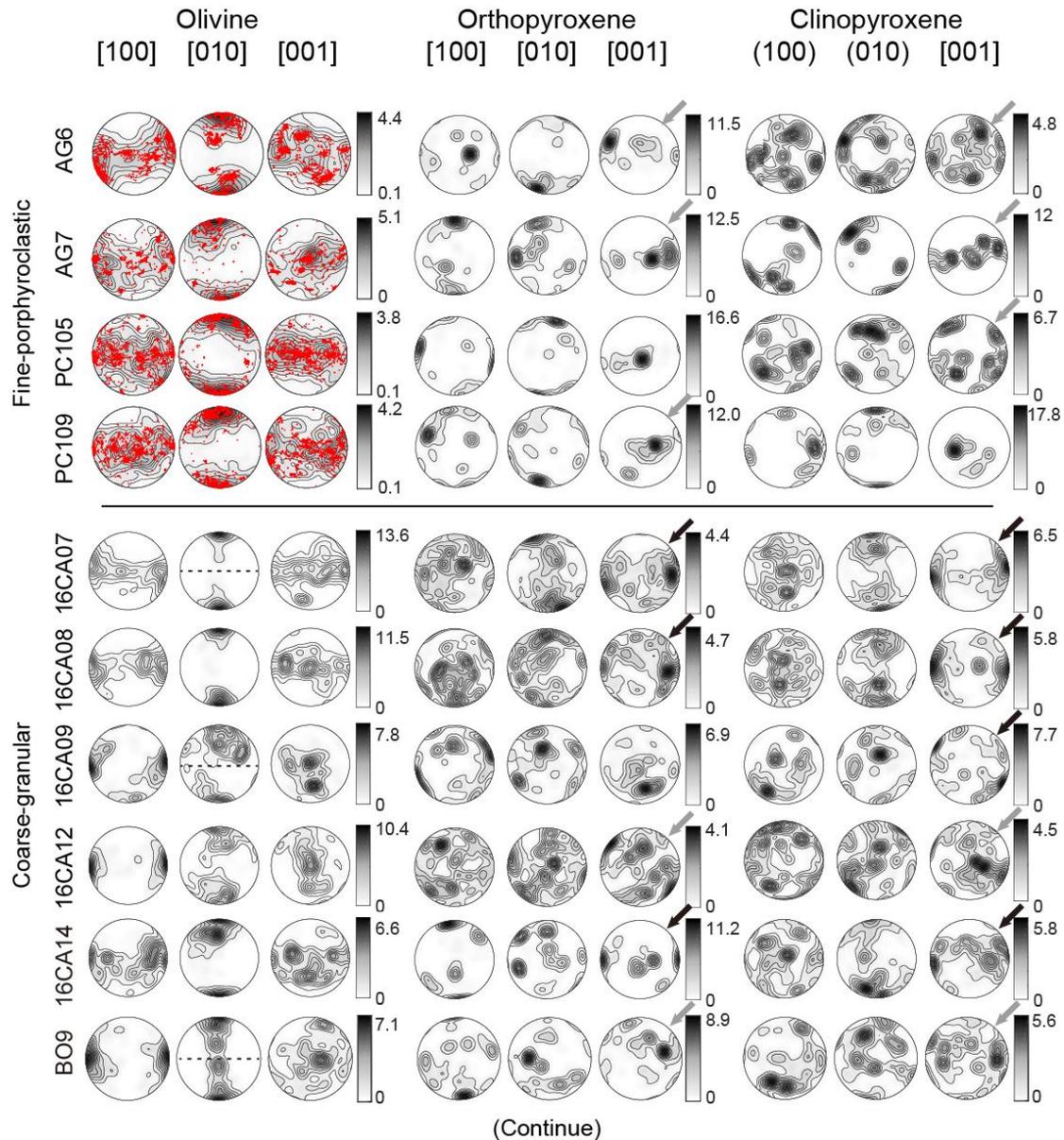
551

552 Fig. 5. Olivine CPO symmetry (BA-index) vs. CPO strength (J-index). The variation in olivine CPO pattern as a
 553 function of BA-index is illustrated at the top of the diagram. For fine-porphroclastic peridotites, we present data for the
 554 bulk rock and for porphyroclasts and neoblasts separately. Values are presented in Table 2. * indicates J-indexes that
 555 are probably overestimated, because <100 grains could be analyzed in the thin section. CPO data for Fernando de
 556 Noronha (FN) xenoliths (Liu et al., 2019) are displayed for comparison.

557

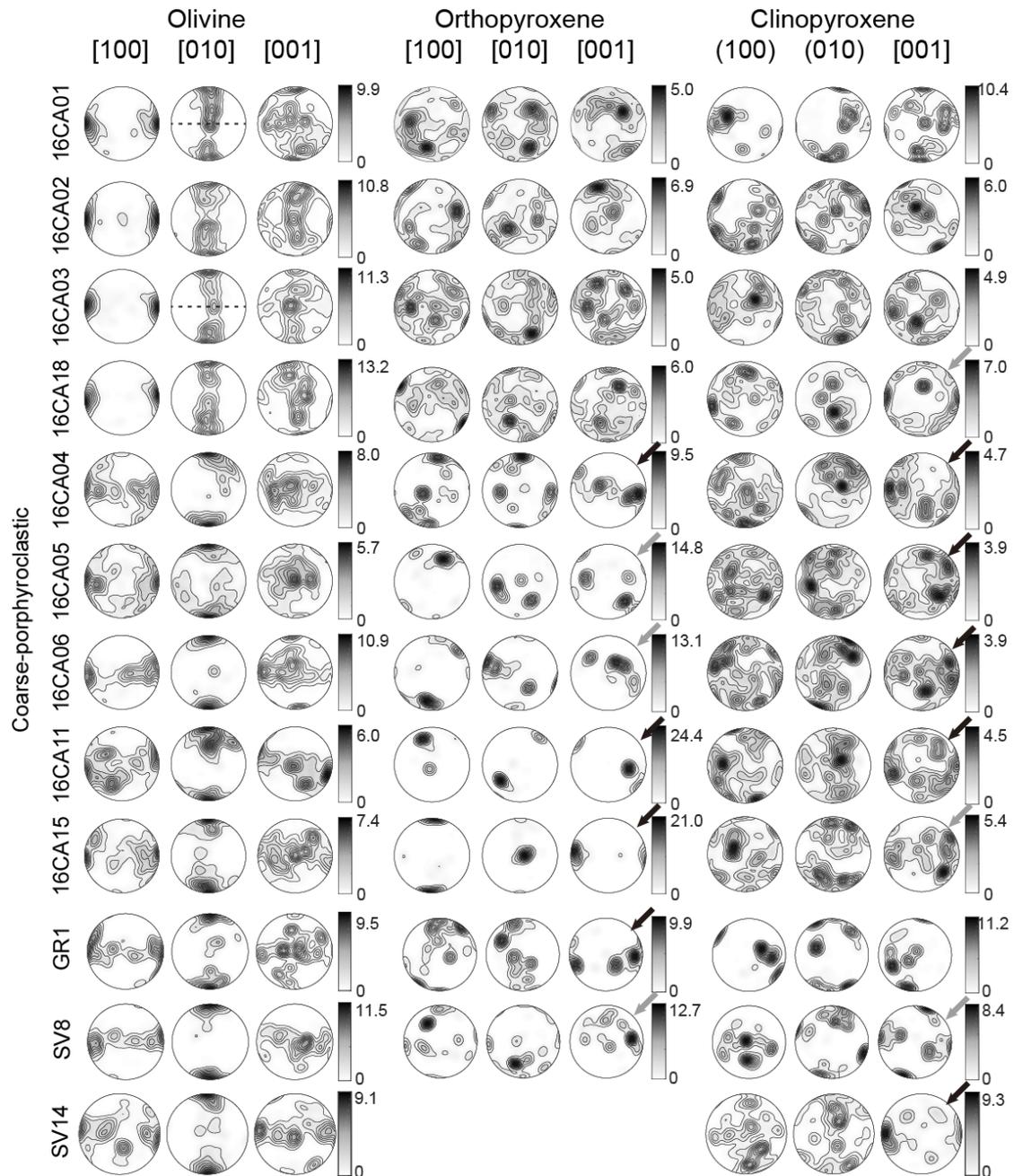
558 Olivine has moderate to strong CPO (Figs. 5 and 6). With exception of
 559 harzburgite AG7, which has a very high J-index of 16.6 due to a few coarse
 560 olivine porphyroclasts that occupy most of the thin section, the J-index varies
 561 between 3.8 and 8.5, with a mean of 6.4. The CPO symmetry varies between
 562 fiber-[010], orthorhombic, and fiber-[100] patterns. **Fine-porphroclastic
 563 peridotites display olivine CPO patterns intermediate between fiber-[010] and
 564 orthorhombic and the neoblasts have systematically weaker olivine CPO (Figs.
 565 5 and 6a).** There is no systematic variation of J-index or BA-index within the
 566 two other microstructural groups. Coarse-granular samples display dominantly
 567 orthorhombic olivine CPO patterns with J-indexes around 6 (Figs. 5 and 6a).
 568 Coarse-porphroclastic samples have more variable CPO patterns and

569 intensities (Fig. 5). Four coarse-porphyroclastic harzburgites (16CA01,
 570 16CA02, 16CA03, and 16CA18) have strong CPO with fiber-[100] patterns
 571 (Fig. 6b). The remainder coarse-porphyroclastic peridotites display variable
 572 CPO intensity orthorhombic patterns with a tendency towards fiber-[010],
 573 expressed as girdle of [100] with a maximum within it and a point maximum of
 574 [010] (Fig. 6b).
 575



576

577 Fig. 6. Crystal preferred orientations (CPO) of olivine, orthopyroxene, and clinopyroxene for all studied
 578 fine-porphyroclastic and coarse-granular peridotites and coarse-porphyroclastic peridotites. Lower hemisphere
 579 stereographic projections with contours at 1 multiple of a uniform distribution intervals. The three pole figures for each
 580 mineral are plotted using the same gray scale. For the fine-porphyroclastic peridotites, the CPO of olivine
 581 porphyroclasts is presented as points and the CPO of olivine neoblasts is presented as contours. Black arrows
 582 indicate pyroxene CPO that are well correlated with the olivine CPO and gray arrows indicate those partially correlated.
 583 When the sample shows a olivine shape preferred orientation marking a foliation, the latter is displayed as a dashed



585

586

587 Orthopyroxene CPO is consistent with the olivine CPO in the majority of
 588 the studied peridotites. It is characterized by $[001]_{\text{opx}}$ and $[100]_{\text{opx}}$ (or $[010]_{\text{opx}}$)
 589 maxima parallel to the $[100]_{\text{ol}}$ and $[010]_{\text{ol}}$ maxima, respectively, suggesting a
 590 common deformation history (arrows in Fig. 6). However,
 591 coarse-porphyroclastic harzburgites 16CA01, 16CA02, 16CA03, 16CA18 and
 592 coarse-granular lherzolite 16CA09 show weak orthopyroxene CPO that are not
 593 consistent with the olivine CPO, characterized by a weak $[001]_{\text{opx}}$ maximum at
 594 high angle to a well-developed $[100]_{\text{ol}}$ maximum.

595 Clinopyroxene shows more dispersed CPO, but which are, in many
 596 samples, consistent with both olivine and orthopyroxene CPO patterns,

597 characterized by $[001]_{\text{cpx}}$ and $[010]_{\text{cpx}}$ maxima subparallel to $[100]_{\text{ol}}$ and $[010]_{\text{ol}}$
598 maxima (arrows in Fig. 6). However, in most fine-porphyroclastic peridotites
599 and in coarse-granular lherzolite 16CA12, although the orthopyroxene CPO is
600 consistent with the olivine CPO, the clinopyroxene CPO is not. In
601 coarse-porphyroclastic harzburgites (16CA01, 16CA02, 16CA03, 16CA18)
602 neither the orthopyroxene nor the clinopyroxene CPO are correlated with the
603 olivine CPO.

604

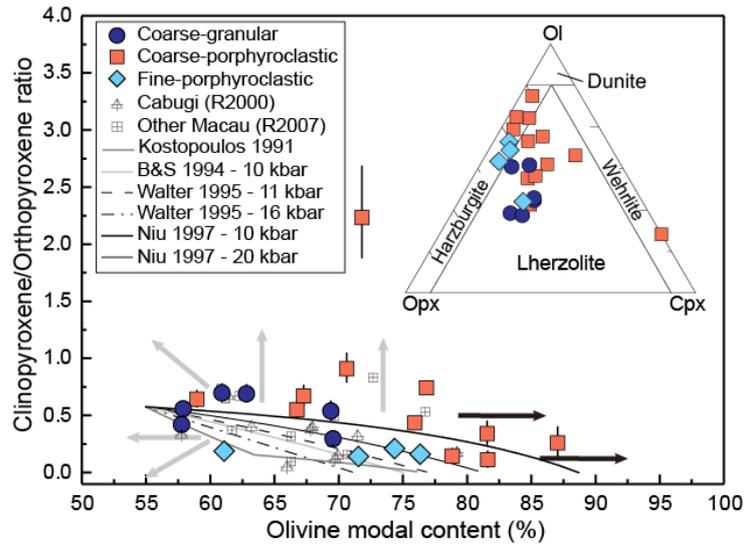
605 **4.3 Modal compositions**

606 Modal compositions were determined based on the area fraction of each
607 mineral in the EBSD maps (Fig. 7 and Table 1). All samples are spinel-facies
608 peridotites. Lherzolites predominate (14/22), but the sampling also includes 7
609 harzburgites and 1 wehrlite. **Predominance of fertile peridotites with no**
610 **dependence on sampling site** was also observed in previous sampling of
611 Macau xenoliths by Rivalenti et al. (2000, 2007; Fig. 7).

612 **The compositions of the two microstructural groups overlap, but**
613 **coarse-granular peridotites are on average more fertile (ol contents of 57-70%)**
614 **than coarse-porphyroclastic peridotites (ol contents between 67-87%, with one**
615 **outlier – 16CA04, which has 59% of olivine and an intermediate**
616 **microstructure). Fine-porphyroclastic peridotites display variable olivine**
617 **contents overlapping with the two other groups.**

618 Comparison of the modal compositions to those predicted by various
619 partial melting models highlights that roughly half of the studied Macau
620 xenoliths are enriched in clinopyroxene or olivine relatively to the models'
621 predictions (Fig. 7). Even for those samples plotting along the partial melting
622 trends, a large variability in partial melting degrees, from 1-20% melting, has to
623 be invoked to explain the full range of observed modal compositions. Such a
624 variation in partial melting is difficult to reconcile within the limited volume of
625 the mantle sampled by these volcanic eruptions. This suggests reactive melt
626 percolation leading to either crystallization of clinopyroxene and/or
627 orthopyroxene at the expenses of olivine (refertilization processes, e.g. Le
628 Roux et al., 2007; represented by gray arrows in Fig. 7) or crystallization of
629 olivine at the expense of pyroxenes (dunitization processes, e.g., Berger and
630 Vannier, 1984; Kelemen, 1990; black arrows in Fig. 7). The observed
631 deviations in modal compositions relatively to partial melting trends are
632 consistent with microstructural evidence for reactive melt percolation, such as
633 the sinuous olivine-pyroxene boundaries and the cusp-like shapes of
634 pyroxenes when in contact with two olivine grains (cf. Figs. 3 and 4).

635



636

637

Fig. 7. Modal compositions of the studied peridotites presented as clinopyroxene/orthopyroxene ratio vs. olivine modal content and as the traditional ol-opx-cpx ternary diagram (insert). Modal compositions of Cabugi and other Macau

638

volcanics peridotite xenoliths studied by Rivalenti et al. (2000) and (2007) are plotted for comparison. Black and gray curves represent the evolution of the modal composition predicted by different partial melting models for an initial

639

fertile modal composition of 55% ol, 28.5% opx, and 16.5% cpx (Baker and Stolper, 1994; Kostopoulos, 1991; Niu, 1997; Walter et al., 1995). Gray arrows indicate evolution trends associated with cpx and/or opx crystallization at the

640

expense of ol (refertilization reactions). Black arrow indicates evolution trends associated with ol crystallization at the expense of pyroxenes (dunitization reactions).

641

642

643

644

645

646

4.4. Mineral compositions

647

The chemical compositions of main mineral phases in the studied Macau peridotite xenoliths are listed in Supplementary Material Table S1.

648

Representative results are displayed in Fig. 8 together with previous data for Macau peridotite xenoliths (Supplementary Material Table S2). Comparison

649

between the present and previous datasets indicates that our sampling is representative of the variability in composition for these localities. Analysis of

650

the present dataset indicates that there is no simple correlation between microstructural types and chemical compositions (Fig. 8). There is also no

651

relation between mineral chemistry and sampling locality.

652

653

654

655

656

The Mg# number [$Mg\# = 100 \times Mg / (Fe + Mg)$, atomic ratio] of olivine, orthopyroxene, and clinopyroxene ranges between 89.1 – 91.6 (average 90.0),

657

89.1 - 91.7 (average 90.4), and 87.9 – 92 (average 90.9), respectively (Figs. 8a,b). Core-rim variations in Mg# are weak for all three minerals in most

658

samples. Olivine and orthopyroxene Mg# within each sample show low dispersion and are positively correlated, with the orthopyroxene being slightly

659

enriched in Mg relatively to olivine (all data points, except fine-porphyroclastic peridotites AG6 and AG7, plot slightly above the 1:1 line in Fig. 8b). Clinopyroxene Mg# shows much higher variability at the sample scale, in

660

particular within coarse-porphyroclastic harzburgites, indicating chemical disequilibrium at the mm-scale. Moreover, average clinopyroxene Mg# in

661

662

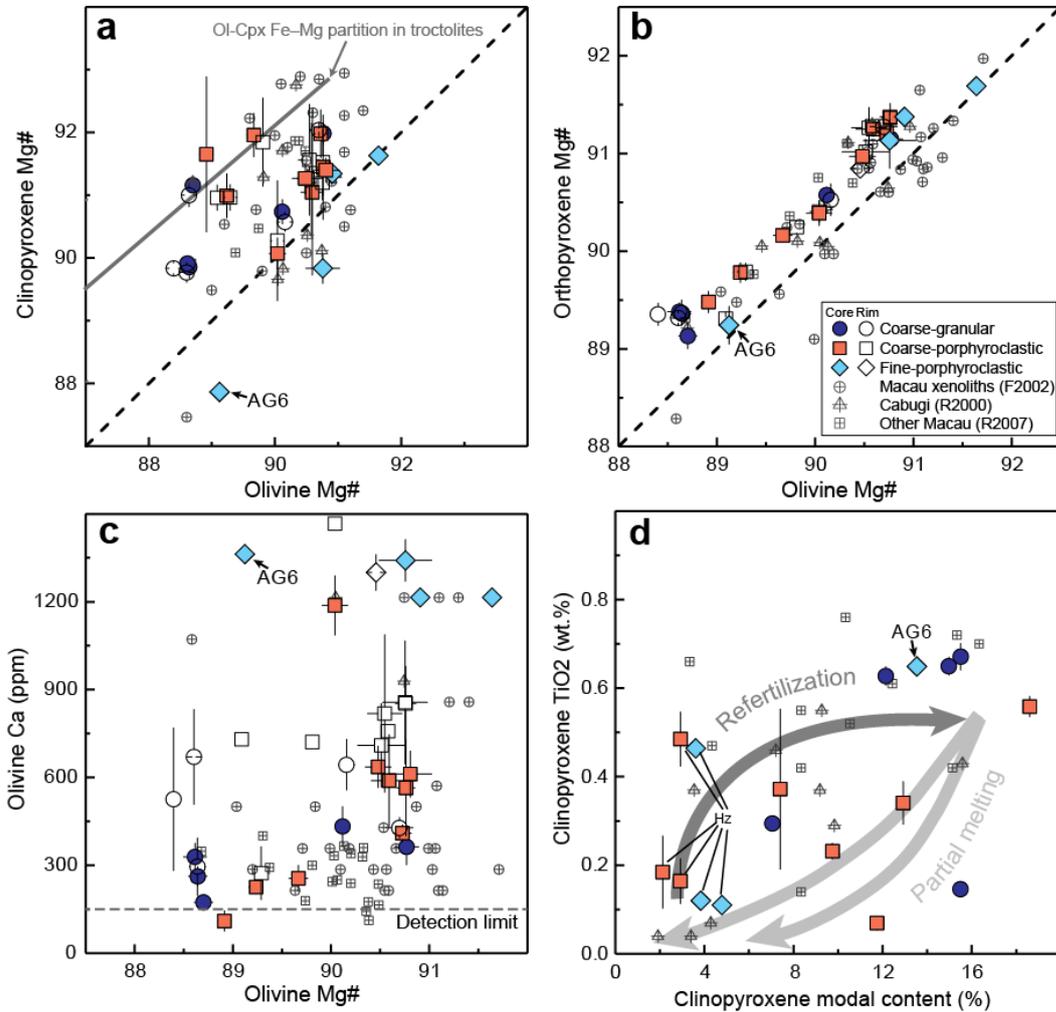
663

664

665

666

667 neither cores nor rims does not display a simple correlation with the olivine
 668 Mg# (Fig. 8a). Most measured Mg# spread between the 1:1 and the Fe-Mg
 669 partition trend between olivine, clinopyroxene, and melt in oceanic troctolites
 670 (Lissenberg and Dick 2008), but fine-porphyroclastic peridotites AG6 and AG7
 671 have clinopyroxenes enriched in Fe relatively to olivine.
 672



673
 674 **Fig. 8.** Chemical compositions for major rock-forming minerals: (a-c) Clinopyroxene Mg#, orthopyroxene Mg#, and
 675 olivine Ca content (ppm) vs. olivine Mg#, (d) Clinopyroxene TiO₂ content (wt.%) vs. modal content (%). Minerals
 676 chemical compositions for Macau peridotite xenoliths previously studied by Rivalenti et al. (2000), (2007), and Fodor
 677 et al. (2002) are plotted for comparison. The value of the depleted mantle (DM) from Workman and Hart (2005) is
 678 displayed in (a) and (b). Fe-Mg partition between olivine and clinopyroxenes in troctolites (Lissenberg and Dick, 2008)
 679 is plotted in (a). Partial melting and refertilization trends proposed by Le Roux et al. (2007) are plotted in (d). Hollow
 680 symbols represent rim compositions and solid symbols represent core ones. Error bars represent the chemical
 681 variation at the sample scale. Data is presented in Supplementary Material Table S1.

682
 683 Olivine cores show low Ca contents with a weak positive correlation with
 684 Mg#, (Fig. 8c), except for the fine-porphyroclastic peridotites and the
 685 coarse-porphyroclastic Iherzolite GR1, which have high Ca contents. Olivine
 686 rims in all studied peridotites are enriched in Ca and show a high variability at

687 the thin section (mm-cm) scale.

688 Clinopyroxene TiO₂ contents (wt.%) do not show the simple positive
689 correlation with cpx modal content (%) expected if partial melting controlled the
690 chemistry of the peridotites (Fig. 8d). Among the harzburgites, four have
691 Ti-poor clinopyroxene grains (TiO₂ contents < 0.2%), while the other two have
692 clinopyroxene grains with TiO₂ contents of 0.55%. All harzburgites show
693 marked variability in the Ti-content of clinopyroxene at the sample scale,
694 indicating disequilibrium. Half of the lherzolites have clinopyroxenes with
695 almost constant TiO₂ contents between 0.55-0.65%, independently of the
696 clinopyroxene modal content. The remaining lherzolites have rather Ti-poor
697 clinopyroxene grains.

698 Spinel Cr# [$Cr\# = 100 \times Cr / (Cr + Al)$] and Mg# contents correlate with the
699 modal composition. Harzburgites have spinel with Cr# ranging between
700 40-57.5 and Mg# ranging between 61-72, whereas lherzolites have spinel with
701 Cr# between 8-20 and Mg# between 73-84 (Supplementary Material Fig. S1a).
702 Yet, the lherzolites show a clear enrichment in Mg in spinel relatively to a
703 typical depleted mantle composition. Fine-porphyroclastic lherzolite AG6 has
704 the highest spinel Mg# (83) and lowest Cr# (11). TiO₂ contents (wt. %) in spinel
705 are low ($\leq 0.18\%$) in most lherzolites (Supplementary Material Fig. S1b). In
706 contrast, spinel in coarse-porphyroclastic harzburgites and lherzolite GR1 has
707 a wide range of average TiO₂ contents, which may reach 1.03%, with strong
708 variations within each sample, indicating disequilibrium. Spinel in
709 fine-porphyroclastic lherzolite AG6 also has a fairly high TiO₂ content of 0.45%.

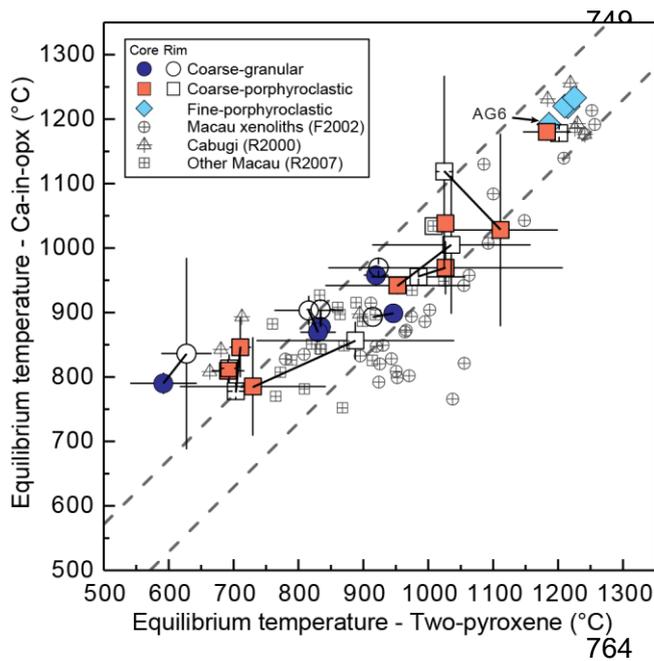
710

711 *4.5. Equilibrium temperatures and geotherm estimation*

712 Most equilibrium temperatures determined using two-pyroxenes
713 geothermometer of Taylor (1998; TTA98) and the Ca in orthopyroxene
714 (Ca-in-opx) of Brey and Kohler (1990) revised by Nimis and Grutter (2010)
715 agree within ± 70 °C (gray dashed lines in Fig. 9). This suggests that they are
716 reliable. However, peridotites with cpx-opx temperatures < 800 °C show larger
717 discrepancies (>90 °C), with systematically higher Ca-in-opx equilibrium
718 temperatures. This inconsistency between the predictions of the two
719 thermometers is observed for both core and rim temperatures and is not
720 correlated with the microstructure. It affects two coarse-porphyroclastic
721 peridotites (16CA06, 16CA11) and one coarse-granular lherzolite (16CA14).
722 The latter has the lowest temperature among all samples with strong
723 inconsistency between the predictions of the two thermometers (592°C and
724 790°C, Table 1).

725 Following Nimis and Grutter (2010), we use the temperature predictions
726 of TTA98 thermometer as the equilibrium temperatures. All
727 fine-porphyroclastic peridotites and the coarse-porphyroclastic peridotite GR1
728 have high equilibrium temperatures over 1200 °C. Coarse-porphyroclastic
729 peridotites have two ranges of equilibrium temperatures: harzburgites
730 (16CA01, 16CA03, 16CA18) as well as clinopyroxene-poor lherzolite SV8

731 have equilibrium temperatures between 900 and 1100 °C, whereas the
 732 clinopyroxene-rich coarse-porphyroclastic lherzolites (16CA06, 16CA11,
 733 16CA15) have lower equilibrium temperatures between 600 and 700 °C.
 734 Coarse-porphyroclastic peridotites 16CA03, 16CA15, and 16CA18 display
 735 core-rim variations in equilibrium temperature. In most cases, rim Ca-in-opx
 736 temperatures are higher than core ones, but there is no systematic trend. All
 737 samples that display core-rim variations also show significant dispersion in
 738 both core and rim temperatures at the sample scale (bars in Fig. 10).
 739 Coarse-granular peridotites 16CA07, 16CA08, 16CA09, and BO9 have
 740 intermediate equilibrium temperatures between 800 and 900 °C and less
 741 variations at sample scale as well as between cores and rims. Comparison
 742 with previous thermometry data on Macau peridotite xenoliths indicates that
 743 the present sampling is representative of the variability in this suite (Fig.9).
 744 Analysis of the full dataset also highlights that there is no relation between
 745 equilibrium temperatures and sampling site. The full range of equilibrium
 746 temperatures is observed both among Pico do Cabugi xenoliths, which is the
 747 best sampled site, and among xenoliths from other Macau volcanics.
 748



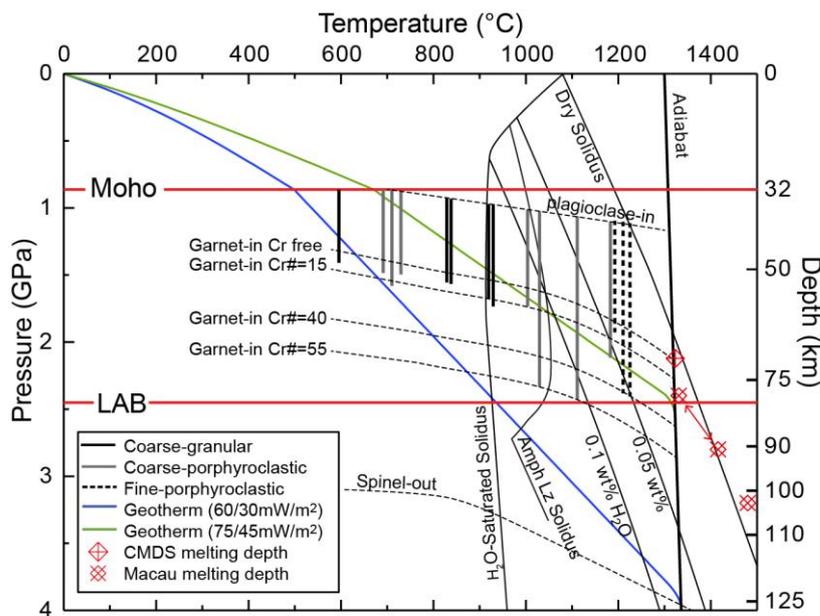
749 **Fig. 9.** Average core and rim equilibrium
 750 temperatures calculated using the cpx-opx
 751 thermometer (TTA98, Taylor, 1998) and the
 752 Ca-in-opx thermometer (Ca-in-opx, Brey and
 753 Kohler, 1990, revised by Nimis and Grutter,
 754 2010). Chemical compositions from Rivalenti
 755 et al. (2000) and (2007) and Fodor et al. (2002)
 756 were used to recalculate the equilibrium
 757 temperatures for their samples using the same
 758 thermometers (data is presented in
 759 Supplementary Material Tables S2 and S3).
 760 Error bars represent data spread within each
 761 sample. Gray dashed lines outline a $\pm 70^\circ\text{C}$
 762 variation between the two thermometers.

765 To estimate the Cenozoic geotherm in the northern Borborema Province,
 766 we plotted the TTA98 equilibrium temperatures of the studied xenoliths against
 767 the peridotite phase diagram in the upper mantle (spinel-out, plagioclase-out,
 768 and garnet-in curves for different Cr contents from O'Neill, 1981 and Klemme,
 769 2004), as well as equilibrium geotherms for different surface and reduced heat
 770 flows (Fig. 11). The geotherms are calculated assuming a surface heat flow of
 771 60 or 70 mW/m² (Hamza et al., 2018), a Moho depth of 32 km (Almeida et al.,
 772 2015), an exponential decrease of the radiogenic heat production with depth,
 773 with a characteristic depth of 10 km, and a reduced heat flow of either 30 or 45
 774 mW/m². For comparison, we also indicate the depth of the

775 lithosphere-asthenosphere boundary determined by P-S receiver functions for
 776 seismic station RCBR (Heit et al., 2007) and the partial melting conditions
 777 estimated based on the compositions of the most primitive melts of the
 778 Cenozoic Macau and Mesozoic Ceará-Mirim suites (Ngonge et al, 2015a,b).

779 P-T equilibrium conditions of the peridotites are displayed as a fixed
 780 temperature corresponding to the TTA98 thermometry prediction, but as a
 781 pressure range, because there are no reliable barometers for spinel-facies
 782 peridotites. The range of equilibrium pressure conditions is defined by the
 783 occurrence of spinel in absence of plagioclase or garnet for the actual Cr
 784 content of each peridotite (O'Neill, 1981; Klemme, 2004).

785 The wide range of equilibrium temperatures suggests that the studied
 786 peridotites represent a complete sampling of the subcontinental lithospheric
 787 mantle. The absence of garnet in the entire suite, which is largely composed of
 788 peridotites with Cr# < 15, constrains a maximum equilibrium depth of the
 789 studied peridotites shallower than 75 km (Fig. 10). By consequence, the
 790 equilibrium conditions of the Macau peridotite xenoliths, except 16CA14 that
 791 has the lowest and likely unreliable equilibrium temperature, are better fitted by
 792 the hotter geotherm (green line in Figure 10).



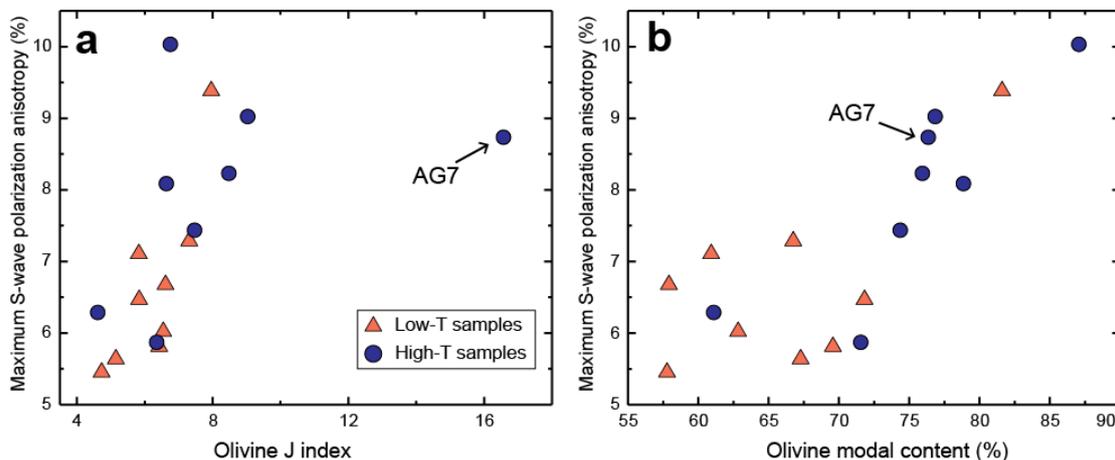
793
 794 Fig. 10. P-T diagram comparing the equilibration conditions estimated for the studied Macau xenoliths to two possible
 795 steady-state geotherms for the north Borborema province (surface / reduced heat flows indicated in parenthesis).
 796 Xenolith data is better fitted by the hotter geotherm (surface and reduced heat flows of 75 and 45mW/m², respectively).
 797 Since no barometers are available for spinel peridotites, equilibrium pressure ranges are presented as bars, based on
 798 the absence of plagioclase and of garnet in the studied peridotites, considering their Cr# (Supplementary Information
 799 Table S1) and the effect of Cr on the spinel-garnet phase transition (Klemme, 2004; O'Neill, 1981). Dry and wet solidus
 800 with variable H₂O contents after Ringwood (1975). Moho and Lithosphere-Asthenosphere Boundary (LAB) depths
 801 derived from S-P and P-S receiver function analyses (Almeida et al., 2015, Heit et al., 2007) as well as partial melting
 802 conditions derived from the primitive melt compositions of the Cretaceous Ceará-Mirim dyke system (CMDS) and
 803 Cenozoic Macau volcanics (Ngonge et al., 2016a,b) are displayed for comparison.

804 **4.6. Seismic properties**

805 For calculating seismic properties, we divided the xenoliths into two
 806 groups as a function of their TTA98 equilibrium temperature: low ($T < 1000\text{ }^{\circ}\text{C}$)
 807 and high-temperature ($T > 1000\text{ }^{\circ}\text{C}$). Then, we estimated the elastic tensors of
 808 the individual samples (Supplementary Material Table S3) for the pressure and
 809 temperature conditions predicted at 50 km (1.46 GPa and $916\text{ }^{\circ}\text{C}$) and 70 km
 810 depth (2.12 GPa and $1189\text{ }^{\circ}\text{C}$) for the geotherm that best fits the xenolith data
 811 ($Q_s=70\text{ mW/m}^2$ and $Q_m=45\text{ mW/m}^2$, green line in Figure 10).

812 Seismic anisotropy patterns of the individual samples vary slightly as a
 813 function of the olivine CPO symmetry. As usual, the seismic anisotropy
 814 intensity correlates positively with both the olivine CPO strength and olivine
 815 modal content (Figure 11). On average, the peridotites equilibrated at
 816 temperatures $<1000\text{ }^{\circ}\text{C}$ tend to display lower anisotropy, due to both more
 817 dispersed olivine CPO and lower olivine contents. However, seismic
 818 anisotropy intensity does not increase linearly with increasing J-index. By
 819 consequence, overestimation of the olivine CPO intensity for
 820 fine-porphroclastic harzburgite AG7 does not result in similar overestimation
 821 of its seismic anisotropy. The maximum S-wave polarization anisotropy varies
 822 from 5.5 to 7.3% in the low-temperature peridotites and from 5.9 to 10% in the
 823 high-temperature ones (Fig. 11 and Table 3). Similarly, the maximum P-wave
 824 propagation anisotropy varies from 7.4 to 10% and from 7.7 to 14.2% (Table 3).
 825 The maximum fast (S1) wave propagation anisotropy varies from 3.0 to 5.9%
 826 in low-temperature samples and from 4.1 to 7.2% in high-temperature ones,
 827 whereas the maximum slow (S2) wave propagation anisotropy varies from 3.4
 828 to 5.2% and from 1.9 to 7.3%, respectively (Table 3).

829



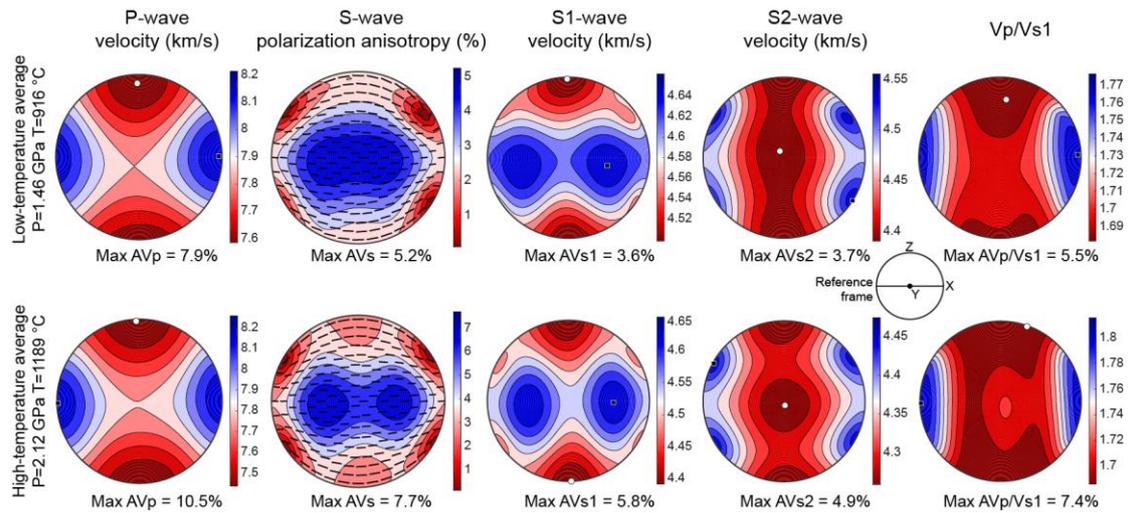
830

831 **Fig. 11.** Maximum S-wave polarization anisotropy (%) vs. (a) olivine CPO intensity (J index) and (b) olivine modal
 832 content (%) for all studied xenoliths. Seismic anisotropy data for individual samples are listed in Table 3.

833

834 Since seismic waves average elastic properties over large volumes, we
 835 calculated average elastic tensors for the low-temperature and
 836 high-temperature groups by assuming a coherent orientation of the past flow
 837 directions and planes (assumed as parallel to the [100] and [010] maxima of
 838 the olivine CPO, respectively – this assumption is justified by the analysis of

839 the CPO as discussed in the next section) over the entire lithosphere using a
 840 constant structural reference frame. These average seismic anisotropy
 841 patterns (Fig. 12) represent the maximum possible seismic anisotropy in the
 842 shallow (low-temperature group) and deep (high-temperature samples)
 843 sections of the lithospheric mantle, if the present sampling is representative of
 844 the volumes of the different microstructures and compositions in the
 845 lithospheric mantle in this region.
 846



847

848 **Fig. 12.** Average seismic properties for shallow and deep sections of the lithospheric mantle beneath the study area
 849 calculated by averaging the elastic constant tensors of the eight low-temperature samples calculated for 916°C,
 850 1.46GPa and of the nine high-temperature samples calculated for 1189°C, 2.12GPa. Lower hemisphere stereographic
 851 projections presenting the variation of the property as a function of the propagation direction relatively to the structural
 852 reference frame (flow direction (X) and normal to the flow plane (Z) shown by the insert). Black squares mark the
 853 highest values and white circles mark the lowest ones. Seismic properties data are listed in Table 3 and the elastic
 854 tensors, in Supplementary Material Table S4.

855

856 Seismic anisotropy patterns for the shallow and deep lithospheric mantle
 857 are similar, but as already discussed, the lower part of the lithosphere is more
 858 anisotropic (Fig. 12). The average maximum P-wave propagation anisotropy is
 859 10.5% with velocities ranging between 7.4 and 8.3 km/s, compared to 7.9% in
 860 the shallow lithospheric mantle, where P-wave velocities vary between 7.6
 861 km/s normal to the flow plane and 8.2 km/s parallel to the flow direction frozen
 862 in the lithospheric mantle (Table 3). The average maximum S-wave
 863 polarization anisotropy in the lower lithospheric mantle is 7.7% compared to
 864 5.2% in the shallower levels; in both cases it is observed for S-waves
 865 propagating within the flow plane but at $<45^\circ$ of the flow direction. Low S-wave
 866 polarization anisotropy is observed for all waves propagating at low angle to
 867 the XZ plane, that is, the plane that contains both the flow direction and the
 868 normal to the flow plane. For all propagation directions in which a significant
 869 polarization anisotropy is observed, the fast S-wave is polarized in the plane
 870 containing the flow direction. The average maximum S1-wave propagation

871 anisotropy increases from 3.6% to 5.8% from the shallow to the deep mantle
872 lithosphere. S1 velocity is maximum within the flow plane and minimum normal
873 to it. S1-waves propagating within this plane show a 90° periodicity in the
874 velocity variation. S2-waves average maximum propagation anisotropy
875 increases from 3.7% to 4.9% with depth. High S2 velocities are associated
876 with propagation within the XZ plane at ca. 45° to the flow direction (X) and low
877 velocities are observed for all propagation directions at high angle to X. The
878 maximum Vp/Vs1 ratio anisotropy increases from 5.5% to 7.4% with depth.
879 Highest Vp/Vs1 ratios (1.78-1.8) are observed for waves propagating parallel
880 to the flow direction and the lowest Vp/Vs1 ratios for waves propagating
881 normal to the flow plane.

882

883 **5. Discussion**

884

885 **5.1. Deformation, annealing, and reactive melt percolation**

886 Coarse-granular peridotites show polygonal grain shapes for both olivine
887 and pyroxenes and low intragranular misorientations (olivine M2M values < 2°,
888 Fig. 2). However, they have well-developed olivine CPO (Figs. 5 and 6). This
889 association supports that the deformation that produced the CPO was followed
890 by annealing (static recrystallization), which effectively reduced dislocation
891 densities and re-equilibrated grain shapes. The analysis of the relative
892 intensity of the orientation of [100], [010], and [001] of olivine, of the relations
893 between olivine and pyroxene CPOs, and of the relations between olivine CPO
894 and SPO (when a SPO was observed) points to deformation by dislocation
895 creep with dominant activation of the [100](010) slip system in olivine,
896 [001](100) in orthopyroxene, and [001](0kl) in clinopyroxene (Tommasi et al.
897 2000, Bascou et al. 2002). Coarse-granular peridotites have roughly constant
898 olivine CPO intensities, but variable symmetry, covering the entire range from
899 fiber-[100] to fiber-[010]. Fiber-[100] and orthorhombic olivine CPO are the
900 usual patterns produced by simple shear under dry, high to moderate
901 temperatures, and low pressure in the upper mantle (Tommasi et al., 1999;
902 2000; Hansen et al. 2014). Fiber-[010] patterns, on the other hand, require
903 particular conditions, which may be: transpression (Tommasi et al., 1999),
904 presence of melts during the deformation (Higgie and Tommasi, 2012, 2014;
905 Qi et al. 2018), changes in the olivine CPO by oriented growth during
906 recrystallization (e.g., Tommasi et al., 2008), or deformation under high stress
907 or high pressure conditions, which would lead to increased activation of [001]
908 glide (Durham and Goetze, 1977; Mainprice et al., 2005; Demouchy et al.
909 2013). The mineralogy and microstructures and CPO of the coarse-grained
910 peridotites falsify the last hypothesis. However, the present data does not allow
911 for discriminating among the other hypotheses. The different olivine CPO
912 symmetry may therefore result from variations in deformation regime, with
913 fiber-[010] CPO recording transpression (which is the dominant deformation
914 regime in the Seridó belt during the Brasiliano event), presence or not of melts

915 during the deformation (all deformation events were accompanied by
916 magmatism), or different CPO evolution during recrystallization.

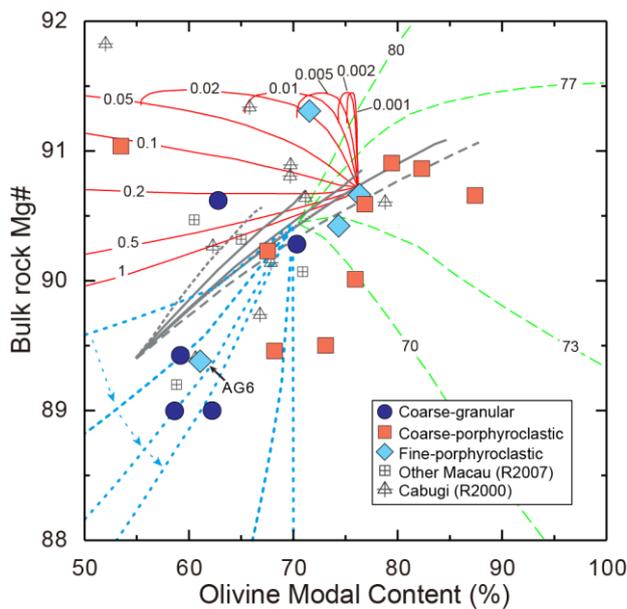
917 Analysis of the bulk rock Mg# vs. olivine modal content relation (Fig. 13)
918 shows that the compositions of most coarse-granular peridotites deviate from
919 partial melting trends, suggesting the occurrence of refertilization processes.
920 Yet, most coarse-granular peridotites have coherent olivine and pyroxene
921 CPOs (Fig. 6), which indicate co-deformation of the two phases. Thus, if
922 melt-rock reactions leading to refertilization occurred, they predated or were
923 concomitant to the deformation. Refertilization reactions produce sutured
924 pyroxene-olivine boundaries and irregular shapes for both minerals. The
925 equilibrated pyroxenes grain shapes in most coarse-granular peridotites
926 indicate therefore that reactive melt percolation producing refertilization also
927 predated annealing. It also implies that interphase grain boundary
928 rearrangements during annealing were effective. These rearrangements
929 depend on transport of ions along grain boundaries in a similar way to the
930 growth of porphyroblasts in a metamorphic rock, but with weaker driving forces
931 (grain boundary energy reduction opposed to chemical gradients). Most
932 coarse-granular peridotites equilibrated around 900°C (Fig. 9). At this
933 temperature, given the diffusivity of Si along olivine and enstatite grain
934 boundaries ($\leq 10^{-27} \text{ m}^3/\text{s}$; Fisler et al. 1997; Fei et al. 2016), grain boundary
935 equilibration at the 250 μm scale, which is the average amplitude of the
936 sinuosity of the olivine-pyroxene grain boundaries in the
937 coarse-porphyroclastic peridotites (Fig. 3c,d), will occur on time scale of
938 several hundreds of Ma. In contrast, irregular pyroxene grain shapes (high
939 orthopyroxene shape factor $\sim 2^\circ$, Fig. 2) in 16CA14 and BO9 point to a later
940 stage of melt percolation. Indeed, in both samples, part of the pyroxenes
941 shows crystal orientations not coherent with the olivine CPO (Fig. 6).

942 Coarse-porphyroclastic peridotites show less equilibrated
943 microstructures. All major phases show sinuous grain boundaries (higher
944 shape factors) as well as undulose extinction and subgrains, which translate
945 into higher M2M values (Fig. 2). This implies a less effective annealing due to
946 either a more recent deformation or lower post-deformation temperatures. The
947 former hypothesis may apply for the coarse-porphyroclastic peridotites with
948 high equilibrium temperatures ($\geq 1000^\circ\text{C}$, Fig. 9) and the latter, for those with
949 low equilibrium temperatures ($< 800^\circ\text{C}$, Fig. 9). Similarly to the coarse-granular
950 peridotites, olivine CPO in coarse-porphyroclastic peridotites is consistent with
951 deformation by dislocation creep with dominant activation of the [100](010)
952 system. This interpretation is corroborated by the high frequency of (100)
953 subgrain boundaries. However, the CPO patterns and intensities are more
954 varied than those of coarse-granular peridotites (Figs. 5 and 6).
955 Coarse-porphyroclastic harzburgites have strong fiber-[100] olivine CPOs,
956 typical of simple shear deformation (Tommasi et al., 1999; 2000; Bystricky et al,
957 2000; Hansen et al., 2014). Lherzolites have orthorhombic to fiber-[010] olivine
958 CPOs with variable strength. As for the coarse-granular peridotites, the

959 fiber-[010] olivine in these coarse-porphyroclastic peridotites may record either
 960 a component of transpression or the presence of melts during the deformation.

961 Although some coarse-porphyroclastic peridotites plot along partial
 962 melting trends in Fig. 13, evidence for reactive melt percolation leading to
 963 crystallization of pyroxenes or olivine is widespread. It encompasses: (i) the
 964 interpenetrating olivine-pyroxene grain boundaries and the locally interstitial
 965 shapes of pyroxenes (Figs. 3 and 4), which imply lack of microstructural
 966 equilibrium, (ii) the high variability of the olivine and pyroxenes chemical
 967 compositions both within grains and between grains in a sample (Fig. 8), which
 968 indicates absence of chemical equilibration, and (iii) lack or weak consistency
 969 between the olivine and the pyroxenes CPOs (Fig. 6). The latter feature is
 970 specific to the coarse-porphyroclastic harzburgites, which also show the
 971 highest equilibrium temperatures and least equilibrated mineral compositions
 972 among the coarse-porphyroclastic peridotites, implying that reactive melt
 973 percolation in these rocks, which sample the lower lithospheric mantle section
 974 in the province, postdates the deformation and is rather recent.

975



976

977 Fig. 13. Olivine modal content (%) vs. bulk rock Mg# in the studied xenoliths compared to evolutions predicted for
 978 partial melting and reactive melt percolation in the mantle. Gray lines represent the composition evolution predicted by
 979 various partial melting models using a source composition with 89.3 Mg# and 55% of olivine up to complete
 980 consumption of clinopyroxene (Bodinier & Godard, 2014). Colored lines represent different melt-rock reactions after
 981 Bodinier & Godard (2014). Red solid lines correspond to precipitation of clino- and orthopyroxene at the expense of
 982 olivine and melt with different mass ratio of crystallized minerals versus infiltrated melt (numbers of the top of the
 983 curves). Green dashed lines show olivine-forming reactions with melts with different Mg# (numbers of the top of the
 984 curves). Blue dotted lines represent multiple episodes of refertilization, starting with low Mg# melts ('primitive' melt
 985 Mg#=74.5), in which the peridotites successively reacted with the evolved melt resulting from previous infiltration
 986 stage (Bodinier et al., 2008). Compositions of Macau peridotite xenoliths previously studied by Rivalenti et al. (2000)
 987 and (2007) are plotted for comparison.

988

989 Fine-porphyroclastic peridotites show a bimodal olivine grain size
990 distribution and strong intragranular misorientations in the olivine
991 porphyroclasts characterized by dynamic recrystallization (Figs. 3 and 4). At
992 the high equilibrium temperatures recorded by these peridotites ($\geq 1200^\circ\text{C}$, Fig.
993 9), diffusion is fast. The low annealing level of the microstructures in these
994 peridotites, indicated by the high intragranular misorientations in the
995 porphyroclasts (Figs. 2 and 3), implies therefore that the deformation episode
996 that produced the recrystallization shortly predated their extraction from the
997 mantle by the Macau volcanism.

998 The variation in recrystallized grain sizes between the various
999 fine-porphyroclastic peridotites might record variations in stress (from ~ 75
1000 MPa in AG6 to ~ 10 MPa using the recrystallized grain size paleopiezometer of
1001 Van der Wal et al., 1993), but the coarser recrystallized grain sizes result more
1002 probably from partial annealing. The stresses estimated for lherzolite AG6 are
1003 high and, for the equilibrium temperature of 1200°C of these peridotites, imply
1004 extremely high strain rates of 10^{-6} to 10^{-9} s^{-1} based on usual olivine flow laws
1005 (Chopra and Paterson, 1981; Hirth and Kohlstedt, 2003; Gouriet et al. 2019).
1006 The microstructure of these peridotites is indeed very similar to the mosaic
1007 microstructure of deep sheared kimberlite-borne peridotites, which has been
1008 traditionally attributed to the initial stages of kimberlite dyke formation, due to
1009 the high stresses and high rates inferred based on the recrystallized grain
1010 sizes and equilibrium temperatures (e.g., Green and Gueguen, 1974; Boullier,
1011 1977; Skemer and Karato, 2008; Baptiste et al. 2012). Similar microstructures
1012 have been observed in the deepest mantle xenoliths from the Labait alkaline
1013 lavas in the Tanzanian craton (Vauchez et al., 2005) and from Malaita alnoites
1014 in the Ontong Java plateau (Tommasi and Ishikawa, 2014) with similar
1015 interpretations proposed.

1016 Fine-porphyroclastic peridotites with high equilibration temperatures have
1017 also been described in Cenozoic peridotite xenoliths, which sample the mantle
1018 beneath major Neoproterozoic shear zones at the border of the Hoggar swell,
1019 N Africa (Kourim et al., 2015). However, in the Hoggar peridotites, equilibration
1020 temperatures are lower ($1000\text{-}1100^\circ\text{C}$) and olivine recrystallization was
1021 associated with crystallization of elongated aggregates of clinopyroxene and
1022 amphibole. The fine-porphyroclastic microstructures were therefore interpreted
1023 as resulting from ductile reactivation and melt channeling in Neoproterozoic
1024 shear zones in response to the development of the Hoggar swell in the
1025 Cenozoic. However, in the fine-porphyroclastic peridotites from the Borborema
1026 province, there is no evidence for neocrystallization of pyroxenes or
1027 amphiboles within the recrystallized domains. Moreover, the fact that the
1028 recrystallized domains in the fine-porphyroclastic peridotites from the
1029 Borborema province do not align marking a foliation (Fig. 3e,f) and the lack of
1030 elongation of the pyroxenes suggests that the recrystallization was associated
1031 with high stresses, but low finite strains. This association, together with the
1032 equilibrium temperatures, which imply that these peridotites are derived from

1033 the base of the lithospheric mantle, is consistent with localized deformation
1034 associated with the formation of the dykes that fed the Cenozoic magmatism.

1035 Analysis of the olivine modal composition relative to the bulk rock Mg#
1036 implies that Iherzolites AG7 and AG6 were affected by refertilization processes
1037 (Fig. 13). The Fe-rich compositions of olivine and pyroxenes in AG6 (Fig. 8a,b)
1038 further point to high cumulated melt-rock ratios. Ortho- and clinopyroxenes in
1039 these two Iherzolites have unusual irregular, but rounded shapes, which clearly
1040 differ from those in coarse-porphroclastic peridotites (cf. EBSD phase maps
1041 in Fig. 3). Yet determining when this refertilization occurred is difficult. At the
1042 high temperatures at which these peridotites equilibrated, chemical diffusion is
1043 fast. Disequilibrium in mineral chemistry at the sample scale, which would
1044 point to melt-rock interaction shortly before extraction is only observed in AG7.
1045 Harzburgites PC105 and PC109 plot along the partial melting trends in Fig. 13
1046 and have higher Mg# in olivine and pyroxenes (Fig. 8a,b), but they also display
1047 chemical evidence for some melt-rock interaction, like enrichment in Ca in
1048 olivine (Fig. 8c).

1049

1050 *5.2. Cenozoic geotherm and thermal evolution of the NBP lithospheric mantle*

1051 We do have evidence in this study supporting a rather hot Cenozoic
1052 geotherm beneath the North Borborema Province. As illustrated in Fig. 10,
1053 equilibrium conditions of the xenoliths are consistent with the surface heat flow
1054 of 60-70 mW/m² measured in the Borborema Province (Hamza et al., 2018) if
1055 the heat flow from the convective mantle is rather high (45mW/m²). This
1056 implies a slightly hotter than average sublithospheric mantle beneath the
1057 Borborema Province, consistently with the low P-wave velocity anomaly
1058 imaged at 100 km depth beneath the Northern Borborema Province east of the
1059 Macau-Queimadas volcanic alignment (Simões Neto et al., 2019; Fig. 1) and
1060 with the weak low S-wave velocity anomaly imaged beneath this region in a
1061 recent global full-waveform tomography model (ca. -2%; French et al., 2013).
1062 The equilibrium geotherm that best fits the equilibrium temperature and
1063 pressure conditions of the Borborema Province is also consistent with the
1064 seismological constraints for the lithosphere-asthenosphere boundary (LAB)
1065 depth of 80km (Heit et al., 2007) and with the partial melting conditions
1066 calculated for the Macau most primitive basalts (1330-1415°C at 80-93km;
1067 Ngonge et al. 2015b; cf. Fig. 10). A hotter than average sublithospheric mantle
1068 may also account for the Cenozoic uplift of the Borborema Plateau (Almeida et
1069 al., 2015; Luz et al., 2015; Klöcking et al., 2018). The equilibrium temperatures
1070 of the studied xenoliths may therefore represent an equilibrium geotherm
1071 established in the Cenozoic in response to a slightly hotter than normal
1072 convective mantle temperature. Geophysical data imply that these conditions
1073 are still active today. Data on the xenoliths does not bring any constraints on
1074 the causes of the higher than average sublithospheric temperatures, which
1075 may result from a diffuse mantle upwelling, perturbation of the convective
1076 pattern by the São Francisco craton, or, as suggested by Simoes Neto et al.

1077 (2018), lateral channeling of hot material from a mantle plume upwelling to the
1078 SW of the Province.

1079 The present data also do not constrain the evolution through time of the
1080 upper mantle temperatures beneath the North Borborema Province. The
1081 estimated melting conditions for the most primitive basalts of the Cretaceous
1082 CMDS (ca. 1320 °C at 70 km depth; Ngonge et al., 2015a) are shallower than
1083 those for the Cenozoic Macau volcanics (Ngonge et al., 2015b, suggesting an
1084 even shallower LAB beneath the North Borborema Province in the Mesozoic.
1085 This suggests that the lithosphere beneath the NBP has probably cooled and
1086 thickened after the Mesozoic extension. The deepest xenoliths might therefore
1087 represent material accreted to the base of the lithosphere after the Mesozoic.

1088

1089 *5.3. Seismic anisotropy in the lithospheric mantle: comparison with SKS* 1090 *splitting data*

1091 Shear wave splitting data in the Borborema Province is highly
1092 heterogeneous and does not relate in a simple way to neither the outcropping
1093 geological structures nor the absolute plate motion of the South American
1094 plate (Bastow et al., 2015). However, the station RCBR, which is the closest to
1095 sampling sites, being located ca. 50 km east of the Pico do Cabugi, displays a
1096 NNE-oriented fast S-wave polarization parallel to the trend of the Seridó belt
1097 and of the main Brasiliano shear zones in the region, and a high delay time
1098 (1.9 s) based on 9 individual measurements (Assumpção et al., 2011).

1099 If we consider that at least part of the SKS splitting measured at RCBR is
1100 produced in the lithospheric mantle, the orientation of the fast polarization
1101 direction constrains the projection of the lineation on the horizontal plane to be
1102 oriented in the NNE direction ($9\pm 11^\circ$). However, there are no constraints on its
1103 plunge or on the dip of the foliation. To draw constraints on the orientation of
1104 the foliation and lineation in the lithospheric mantle, which would allow to
1105 discuss possible coupling between crustal and mantle structures, we estimate
1106 the contribution of the lithospheric mantle to the SKS splitting delay time (Δt)
1107 for three end-member orientations of the foliation and lineation in the
1108 lithospheric mantle, illustrated in Fig. 14, and compare these predictions to the
1109 observations at RCBR.

1110 Common conversion point (CCP) receiver function stacks support that the
1111 Moho is at ~32km (Almeida et al., 2015) and the LAB is at ~80km based on the
1112 S receiver function data by Heit et al. (2007) in the Northern Borborema
1113 Province. The thickness of the lithospheric mantle in this region is therefore 48
1114 km. Based on the thermobarometric data (Fig. 10), we divided the lithosphere
1115 into two layers and calculated the lithospheric mantle contribution to the
1116 measured delay time using the average seismic anisotropy of the low
1117 temperature peridotites (32 - 56km, < 1000 °C) and of the high-temperature
1118 ones (56 - 80 km, > 1000 °C).

1119

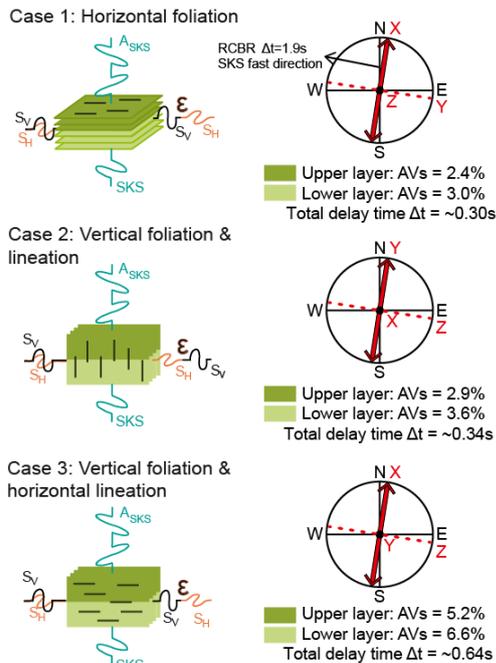


Fig. 14. Estimation of the maximum SKS splitting that may be produced in the lithospheric mantle for three end-member orientations of the flow directions and planes. Stereographic projections show the relation between the geographic (in black) and the structural (in red) reference frames in the three cases. The vertical is in all cases at the center of the diagram. The orientation of the fast SKS polarization (thick red arrow) is based on SKS splitting data for station RCBR.

1136

1137 If the foliation is horizontal and the lineation NE-SW (case 1), the S-wave
 1138 polarization anisotropy is 2.4% for upper lithospheric mantle and 3.0% for
 1139 lower lithospheric mantle. The cumulated SKS delay time is only $\sim 0.30\text{s}$. If
 1140 both the foliation and lineation are vertical (case 2), the fast SKS polarization
 1141 constrains the direction of the foliation, and SKS waves propagate parallel to
 1142 the lineation. The S-wave polarization anisotropy in this direction is low, like in
 1143 case 1. It is 2.9% for upper layer and 3.6% for lower one and the delay time
 1144 that may be cumulated in the lithospheric mantle is $\sim 0.34\text{s}$. If the foliation is
 1145 vertical but the lineation is horizontal and parallel to the fast SKS polarization
 1146 direction (case 3), the S-wave polarization anisotropy is higher. It is 5.2% in the
 1147 upper lithospheric mantle and 6.6% for lower lithospheric mantle, leading to a
 1148 cumulate delay time in the lithospheric mantle of 0.64s .

1149 In case 3, the lithospheric mantle has a fabric consistent with the crustal
 1150 deformation around station RCBR, which is dominated by dextral strike-slip in
 1151 transpressional structures in the Seridó belt and in the shear zones that border
 1152 it. This case would therefore imply a structuration in the lithospheric mantle
 1153 coherent with the crustal deformation in the Brasiliano event. Seismic
 1154 anisotropy in the lithospheric mantle, with anisotropy directions parallel or
 1155 subparallel to the main nearby Neoproterozoic shear zones is also required by
 1156 receiver function analysis on most stations in the Borborema Province
 1157 (Lamarque and Julià, 2019). At station RCBR, this study proposes horizontal
 1158 anisotropy axis trending NNE in the lithospheric mantle, consistent with case 3.

1159 It is important to note that even for case 3, the lithospheric mantle can
 1160 contribute to $< 1/3$ of the SKS delay time measured at the RCBR station ($1.9 \pm$
 1161 0.2 s , Assumpção et al., 2011). Thus a large part of the SKS signal at RCBR
 1162 has to be produced in the asthenosphere and, for the contributions of the
 1163 lithosphere and asthenosphere to add up, asthenospheric flow directions

1164 should not deviate much from NNE. Present-day absolute plate motion (APM)
1165 directions for NE Brazil do not follow this direction. Hotspot reference frame
1166 models, such as HS3-Nuvel-1A, predict an ENE direction, whereas no-net
1167 rotation models predict a NNW direction (Gripp and Gordon, 2002). However,
1168 a NNE flow direction in the sublithospheric mantle beneath the Borborema
1169 province is predicted by models in which the present-day mantle flow is
1170 calculated based on a density anomaly distribution derived from global seismic
1171 tomography models (cf. Fig. 8 in Assumpção et al., 2011).

1172

1173 *5.4. Relations between the mantle structure and the geodynamical evolution of* 1174 *the Borborema Province*

1175 When did the deformation, annealing, and reactive melt percolation
1176 processes recorded by the xenoliths happen? How do they relate to the
1177 geodynamical evolution of NE Brazil? There are no ways of dating deformation
1178 processes in the mantle. However, we may try to use the data discussed in the
1179 previous sections to constrain the imprint of the different tectonic events that
1180 affected the Borborema Province in the lithospheric mantle.

1181 The coarse-granular microstructures, with their well-equilibrated
1182 pyroxene-olivine grain boundaries, imply effective annealing, which, at the
1183 equilibrium temperatures of these xenoliths, require very long time delays, on
1184 the order of several hundreds of Ma. Although the temperature may have
1185 varied since the deformation of the xenoliths, to re-equilibrate the
1186 pyroxene-olivine grain boundaries in <100 Ma, temperatures $\geq 1100^{\circ}\text{C}$, that
1187 is $>200^{\circ}\text{C}$ above those recorded by the xenoliths at the time of extraction, are
1188 needed. Coarse-granular microstructures are well expressed among the
1189 low-equilibrium temperature xenoliths, suggesting that the shallow part of the
1190 lithospheric mantle beneath the Macau volcanics records essentially old
1191 tectonic events. The seismic anisotropy data in the region imply that this
1192 lithospheric structure results from Neoproterozoic dextral strike-slip and
1193 transpressional intraplate deformation. A pre-Mesozoic origin of this mantle
1194 fabric is also suggested by the higher annealing degree of the microstructures
1195 of the Borborema xenoliths relative to the Fernando de Noronha ones (Fig. 2).
1196 Indeed, Fernando de Noronha mantle xenoliths record deformation frozen up
1197 in the oceanic lithosphere by plate cooling, which post-dates the opening of the
1198 Atlantic (Liu et al., 2019).

1199 Some coarse-porphroclastic Iherzolites have lower equilibrium
1200 temperatures ($\leq 800^{\circ}\text{C}$) than most coarse-granular peridotites. This is not a
1201 sampling location effect, since both types of peridotites were sampled in Pico
1202 do Cabugi. The higher olivine intragranular misorientations and less
1203 equilibrated olivine and pyroxene grain shapes indicate less effective
1204 annealing, pointing to either slower annealing due to cooler temperatures or a
1205 more recent deformation event. In the first case, the deformation recorded by
1206 these xenoliths may also be associated with the Brasiliano event. In the
1207 second, these peridotite may record Cretaceous or even younger deformation

1208 preserved in the shallow lithospheric mantle. In any case, the studied xenolith
1209 suite exhibits no evidence of strong deformation under low temperature
1210 conditions. There are no xenoliths displaying mylonitic microstructures, with
1211 marked grain size reduction of olivine and strong elongation of olivine and
1212 orthopyroxene porphyroclasts, which are usually observed in extensional
1213 shear zones developed in peridotite massifs under low temperature conditions
1214 ($\leq 1000^{\circ}\text{C}$; e.g., Drury et al., 1991; Frets et al., 2014; Kaczmarek and Tommasi,
1215 2011). Although annealing could have partially obliterated the olivine
1216 deformation microstructure, it cannot change the aspect ratios of the
1217 orthopyroxenes in a time scale of 100 Ma. Thus if the low temperature
1218 coarse-granular peridotites correspond to sections of the lithospheric mantle
1219 deformed during the Mesozoic, these are low strain zones and their olivine
1220 CPO and seismic anisotropy may still preserve orientations produced by the
1221 previous deformation episodes.

1222 Based on the microstructures of the low equilibration temperature
1223 peridotites and on the seismic anisotropy data, we conclude therefore that the
1224 shallow sections of the lithospheric mantle of the Borborema province records
1225 coupled crust and mantle deformation during the formation of the Borborema
1226 shear zone system in the Neoproterozoic. The present dataset has no
1227 evidence for extensive reworking of the shallow lithospheric mantle by the
1228 extensional deformation in the Cretaceous or during the Cenozoic uplift of the
1229 province. Yet, the absence of (annealed or not) low-temperature mylonites in
1230 the xenolith sampling does not imply that shear zones accommodating a
1231 Mesozoic extension did not form in the shallow lithospheric mantle of the
1232 Borborema Province, since our sampling is punctual and such a deformation
1233 would be by nature heterogeneous.

1234 The coarse-porphyroclastic microstructures that characterize the lower
1235 part of the lithospheric mantle (equilibrium temperatures $> 1000^{\circ}\text{C}$) are more
1236 difficult to relate to a given tectonic episode. Partial melting conditions
1237 estimated for the Ceará-Mirim basalts suggest that the
1238 lithosphere-asthenosphere boundary was shallower in the Cretaceous than in
1239 the Cenozoic (cf. Fig. 10 and discussion section 5.2). This would imply that the
1240 deep lithospheric mantle beneath Borborema might be composed by material
1241 accreted by cooling since the Cretaceous. Comparison between the
1242 microstructures of these coarse-porphyroclastic peridotites with those of
1243 peridotite xenoliths from the nearby Fernando de Noronha (FN) archipelago,
1244 formed in response to Cenozoic volcanism onto 100-105 Ma old crust in the
1245 equatorial Atlantic (Liu et al., 2019) also favors a Mesozoic age for the
1246 deformation of these peridotites. Indeed, the olivine M2M and olivine and
1247 pyroxene shape factors of the Borborema peridotites overlap with the lower
1248 range of olivine M2M and shape factors, that is, with the most annealed
1249 microstructures of the FN peridotites (Fig. 2). The FN peridotites record an
1250 asthenospheric deformation, which has been frozen in the oceanic lithosphere
1251 by cooling and evolved by annealing since then (Liu et al., 2019). Their

1252 deformation is therefore younger than the opening of the Equatorial Atlantic.
1253 The similar to slightly stronger annealing degree of the coarse-porphyroclastic
1254 Macau peridotites suggests that the deformation that produced these
1255 microstructures in the deep section of the lithospheric mantle beneath the
1256 Borborema province might be related to the Cretaceous extension. Thus,
1257 although this event did not result in widespread deformation of the shallow
1258 levels of the lithospheric mantle, it might have reworked the base of the
1259 lithosphere (bottom-up lithospheric thinning?). Yet the high SKS delay times in
1260 station RCBR do not favor strong deviations in flow direction from NNE across
1261 the entire lithosphere-asthenosphere section, implying that either strains were
1262 small and did not change the Neoproterozoic CPO orientations or that flow
1263 directions during extension were at high angle to the extensional structures in
1264 the Cretaceous basins and to the Equatorial Atlantic spreading directions.

1265 Finally, the very low annealing degree and high equilibration temperatures
1266 of the fine-grained porphyroclastic microstructures necessarily imply
1267 deformation close in time to the Cenozoic Macau volcanism. Based on the
1268 microstructural evidence for high stresses, but low finite strains for this
1269 deformation, we propose that it is related to the formation of the dykes that
1270 brought the Macau volcanics to the surface. The equilibration temperatures of
1271 the xenoliths imply a rather hot geotherm, leading to an 80-km thick
1272 lithosphere in the Cenozoic. This is consistent with melting depths inferred for
1273 the Macau basalts (Ngonge et al., 2015b). It is also coherent with the
1274 present-day thermal state of the lithospheric and sublithospheric mantle as
1275 imaged by geophysical data .

1276 Dating the reactive melt percolation events is even more difficult than the
1277 deformation ones. The equilibrated microstructures and the coherent olivine
1278 and pyroxene CPOs in the coarse-granular peridotites undoubtedly point to a
1279 Brasiliano or older reactive melt percolation event. Coarse-porphyroclastic
1280 peridotites record younger melt percolation events, which in some cases
1281 predated or were synchronous to the main deformation recorded by the
1282 samples. However, in other cases, like in the coarse-porphyroclastic
1283 harzburgites, melt percolation post-dated the main deformation. Finally, the
1284 non-equilibrated chemical compositions at the sample scale observed in many
1285 samples, in particular the high-temperature coarse-porphyroclastic
1286 harzburgites, point to a last event of melt percolation that shortly predated the
1287 extraction of the peridotites. In summary, this peridotite suite recorded multiple
1288 reactive melt percolation events, probably well separated in time and related to
1289 the different magmatic episodes recorded in the crust.

1290

1291 **6. Conclusion**

1292 Integrated analysis of microstructures, crystal preferred orientations,
1293 mineral chemical compositions, and equilibrium temperatures in a suite of 22
1294 peridotite xenoliths reveals that the lithospheric mantle beneath the Northern
1295 Borborema Province preserves microstructures related to different deformation

1296 episodes since at least the Neoproterozoic. In all cases, olivine CPO points to
1297 deformation by dislocation creep with dominant activation of [100](010) slip
1298 system. However, the deformation microstructures were modified by variable
1299 degrees of annealing. The analysis of the extent of the annealing considering
1300 the equilibration temperatures allows rough 'dating' of the deformation
1301 episodes and relating them to the major deformation events recorded in the
1302 crust. The well-annealed olivine microstructures and pyroxene shapes in
1303 coarse-granular peridotites equilibrated at ca. 900°C indicate that the last
1304 deformation event that affected these peridotites is several hundreds of Ma old.
1305 In contrast, the fine-porphroclastic peridotites, which have equilibrium
1306 temperatures $\geq 1200^\circ\text{C}$, have suffered a high stress deformation, which shortly
1307 predated their extraction, probably related to the dykes that fed the Cenozoic
1308 volcanism. The coarse-porphroclastic microstructures, which are observed
1309 both in the shallow and deep lithospheric mantle are more difficult to relate to a
1310 given tectonic episode. Yet comparison between the microstructures of these
1311 peridotites and those of peridotite xenoliths from nearby Fernando de Noronha
1312 island, which sample the oceanic mantle lithosphere of an old domain of the
1313 Equatorial Atlantic, suggest that the high-temperature coarse-porphroclastic
1314 peridotites may record deformation related to the Cretaceous extension.
1315 Multiple reactive melt percolation events, probably well spaced in time, may
1316 also be inferred based on the microstructures, modal, and mineral
1317 compositions of the xenoliths.

1318 Comparison of the computed seismic anisotropy of the lithospheric mantle
1319 based on the xenolith data to SKS splitting in nearby RCBR station supports
1320 that the strongest contribution of the lithospheric mantle to the measured
1321 anisotropy would correspond to a frozen strike-slip fabric parallel to the major
1322 NNE-NE Neoproterozoic shear zones in the region. A shallow lithospheric
1323 mantle fabric parallel to the Neoproterozoic shear zones is also suggested by
1324 anisotropic receiver functions (Lamarque and Julia, 2019). These observations
1325 corroborate the conclusion that the shallow lithospheric mantle in the Northern
1326 Borborema province still preserves a structure acquired by coupled
1327 crust-mantle deformation during the formation of the Borborema shear zone
1328 system in the Neoproterozoic. It also suggests that Cretaceous extension,
1329 which seems to be recorded in the deeper sections of the lithosphere, did not
1330 produce pervasive reworking of the shallow lithospheric mantle, pointing to
1331 'partial' or total crust-mantle decoupling during this event. However, even if the
1332 entire lithospheric mantle has a frozen strike-slip fabric parallel to the major
1333 NNE-NE Neoproterozoic shear zones in the region, it can produce $<1/3$ of the
1334 measured delay time of 1.9s in station RCBR. Thus most of the measured SKS
1335 splitting in RCBR should record flow in the sublithospheric mantle, which also
1336 has a NNE orientation, which is not parallel to the APM, but is consistent with
1337 predictions of mantle circulation models for this region.

1338 Finally, equilibrium temperatures and petrological compositions of the
1339 xenoliths indicate a rather hot Cenozoic geotherm, implying a ca. 80 km thick

1340 lithosphere. This estimate is consistent with the melting conditions estimated
1341 for the formation of the Macau basalts (Ngonge et al. 2015). It is also coherent
1342 with geophysical data that point to a present-day 80-km thick lithosphere (Heit
1343 et al. 2007) and hotter than average sublithospheric mantle beneath this region
1344 (French et al., 2013; Simões Neto et al., 2019).

1345

1346 **7. Acknowledgments**

1347 S. Liu was funded during his 14-month stay in Geosciences Montpellier
1348 by the China Scholarship Council. M.M. was supported by projects MIUR-PRIN
1349 2005-2005049821 and 2017-20178LPCPW. C. Archanjo and M.H.B.M.
1350 Hollanda are thanked for assistance during field work. D. Delmas and C.
1351 Nevado prepared high-quality polished thin sections for EBSD measurements,
1352 which were performed with the assistance of F. Barou at the EBSD-SEM INSU
1353 national facility at Géosciences Montpellier (CNRS & Université de
1354 Montpellier). Electron microprobe analyses were carried out with the
1355 assistance of B. Boyer at the Service Microsonde Sud, Université de
1356 Montpellier. S. Piazzolo and T. Morishita are warmly thanked for their
1357 constructive reviews.

1358 The data used in this article are presented in the figures, tables, and
1359 supporting material. The raw EBSD data are available from the corresponding
1360 author upon request.

1361

1362

1363 **8. References**

- 1364 Abramson, E.H., Brown, J.M., Slutsky, L.J., Zaug, J.M., 1997. The elastic
1365 constants of San Carlos olivine to 17 GPa. *J Geophys Res-Sol Ea.* 102,
1366 12253-12263.
- 1367 Almeida, Y.B., Julià, J., Frassetto, A., 2015. Crustal architecture of the
1368 Borborema Province, NE Brazil, from receiver function CCP stacks:
1369 Implications for Mesozoic stretching and Cenozoic uplift. *Tectonophys.* 649,
1370 68-80.
- 1371 Ancelmi, M.F., Santos, T.J.S.d., Amaral, W.d.S., Fuck, R.A., Dantas, E.L.,
1372 Zincone, S.A., 2015. Provenance of metasedimentary rocks from the Ceará
1373 Central Domain of Borborema Province, NE Brazil: implications for the
1374 significance of associated retrograded eclogites. *J. S. Am. Earth Sci.* 58,
1375 82-99.
- 1376 Anderson, O.L., Isaak, D., Oda, H., 1992. High-Temperature Elastic-Constant
1377 Data on Minerals Relevant to Geophysics. *Rev. Geophys.* 30, 57-90.
- 1378 Archanjo, C. J., Hollanda, M. H. B., Rodrigues, S. W., Neves, B. B., &
1379 Armstrong, R., 2008. Fabrics of pre-and syntectonic granite plutons and
1380 chronology of shear zones in the Eastern Borborema Province, NE Brazil. *J.*
1381 *Struct. Geol.* 30, 310-326
- 1382 Archanjo, C.J., Viegas, L.G.F., Hollanda, M.H.B.M., Souza, L.C., Liu, D., 2013.
1383 Timing of the HT/LP transpression in the Neoproterozoic Seridó Belt

1384 (Borborema Province, Brazil): Constraints from UPb (SHRIMP)
1385 geochronology and implications for the connections between NE Brazil and
1386 West Africa. *Gondwana Res.* 23, 701-714.

1387 Assumpção, M., Guarido, M., van der Lee, S., Dourado, J.C., 2011.
1388 Upper-mantle seismic anisotropy from SKS splitting in the South American
1389 stable platform: A test of asthenospheric flow models beneath the
1390 lithosphere. *Lithosphere* 3, 173-180.

1391 Bachmann, F., Hielscher, R., Schaeben, H., 2010. Texture Analysis with MTEX
1392 – Free and Open Source Software Toolbox. *Solid State Phenomena* 160,
1393 63-68.

1394 Bachmann, F., Hielscher, R., Schaeben, H., 2011. Grain detection from 2d and
1395 3d EBSD data - Specification of the MTEX algorithm. *Ultramicroscopy* 111,
1396 1720-1733.

1397 Baker, M.B., Stolper, E.M., 1994. Determining the Composition of
1398 High-Pressure Mantle Melts Using Diamond Aggregates. *Geochim.*
1399 *Cosmochim. Acta* 58, 2811-2827.

1400 Baptiste V., Tommasi, A., Demouchy S., 2012. Deformation and hydration of
1401 the lithospheric mantle beneath the Kaapval craton. *Lithos* 149, 31-50.

1402 Barruol, G., Bonnin, M., Pedersen, H., Bokelmann, G.H.R. & Tiberi, C., 2011.
1403 Belt-parallel mantle flow beneath a halted continental collision: the Western
1404 Alps. *Earth Planet. Sci. Lett.* 302, 429-438 .

1405 Bascou, J., Tommasi, A., Mainprice, D. 2002. Plastic deformation and
1406 development of clinopyroxene lattice preferred orientations in eclogites. *J.*
1407 *Struct. Geol.*, **24**: 1357-1368

1408 Bastow, I.D., Julia, J., do Nascimento, A.F., Fuck, R.A., Buckthorp, T.L.,
1409 McClellan, J.J., 2015. Upper mantle anisotropy of the Borborema Province,
1410 NE Brazil: Implications for intra-plate deformation and sub-cratonic
1411 asthenospheric flow. *Tectonophys.* 657, 81-93.

1412 Berger, E.T., Vannier, M., 1984. Dunites in Alkali Basalts from Oceanic
1413 Volcanic Islands - Petrological Approach. *Bull. Mineral.* 107, 649-663.

1414 Bodinier, J.L., Garrido, C.J., Chanefo, I., Bruguier, O., Gervilla, F., 2008. Origin
1415 of Pyroxenite-Peridotite Veined Mantle by Refertilization Reactions:
1416 Evidence from the Ronda Peridotite (Southern Spain). *J. Petrol.* 49,
1417 999-1025.

1418 Bodinier, J.L., Godard, M., 2014. Orogenic, Ophiolitic, and Abyssal Peridotites,
1419 in: Holland, H.D., Turekian, K.K. (Eds.), *Treatise on Geochemistry* (Second
1420 Edition). Elsevier, Oxford, pp. 103-167.

1421 Boullier, A.M., 1977. Structure des péridotites en enclaves dans les kimberlites
1422 d'Afrique du Sud. Conséquences sur la constitution du manteau supérieur.
1423 *Bull. Soc. Française Minéral. Cristall.* 100, 214–229.

1424 Bonnin, M., Tommasi, A., Hassani, R., Chévrot, S., Wookey, J., Barruol, G.,
1425 2011. Numerical modeling of upper mantle anisotropy beneath a migrating
1426 strike-slip plate boundary : the San Andreas fault system. *Geophys. J.*
1427 *Intern.* 191, 436-458

1428 Brey, G.P., Kohler, T., 1990. Geothermobarometry in 4-Phase Lherzolites .2.
1429 New Thermobarometers, and Practical Assessment of Existing
1430 Thermobarometers. *J. Petrol.* 31, 1353-1378.

1431 Bystricky, M., Kunze, K., Burlini, L., Burg, J. P.; 2000. High shear strain of
1432 olivine aggregates: Rheological and seismic consequences. *Science* 290,
1433 1564-1567

1434 Caby, R., Sial, A., Arthaud, M., Vauchez, A., 1991. Crustal evolution and the
1435 Brasiliano orogeny in Northeast Brazil. In: *The West African orogens and*
1436 *Circum Atlantic correlatives.* Springer- Verlag, pp. 373-397.

1437 Castro, D.L., Bezerra, F.H.R., Sousa, M.O.L., Fuck, R.A., 2012. Influence of
1438 Neoproterozoic tectonic fabric on the origin of the Potiguar Basin,
1439 northeastern Brazil and its links with West Africa based on gravity and
1440 magnetic data. *J. Geodyn.* 54, 29-42

1441 Castro, D.L., Oliveira, D.C., Gomes Castelo Branco, R.M., 2007. On the
1442 tectonics of the Neocomian Rio do Peixe Rift Basin, NE Brazil: Lessons
1443 from gravity, magnetics, and radiometric data. *J. S. Am. Earth Sci.* 24,
1444 184-202.

1445 Chai, M., Brown, J.M., Slutsky, L.J., 1997. The elastic constants of an
1446 aluminous orthopyroxene to 12.5 GPa. *J Geophys Res-Sol Ea.* 102,
1447 14779-14785.

1448 Chopra, P.N., Paterson, M.S., 1981. The Experimental Deformation of Dunite.
1449 *Tectonophys.* 78, 453-473.

1450 Comin-Chiaramonti, P., Demarchi, G., Girardi, V.A.V., Princivalle, F., Sinigoi, S.,
1451 1986. Evidence of mantle metasomatism and heterogeneity from peridotite
1452 inclusions of northeastern Brazil and Paraguay. *Earth Planet. Sci. Lett.* 77,
1453 203-217.

1454 Costa, A.P., Cunha, A.L.C., Cavalcante, R., Medeiros, V.C., Spicily, A.L.,
1455 Dantas, A.R., 2016. Carta Geologica Integrada de Areas de Relevante
1456 Interesse Mineral: Area Serido, Projeto Metalogenia das Provincias
1457 Mineraias do Brasil, Escala 1:350000, Companhia de Pesquisas de
1458 Recursos Mineraias, Brazil.

1459 Darros de Matos, R.M., 1999. History of the northeast Brazilian rift system:
1460 kinematic implications for the break-up between Brazil and West Africa.
1461 *Geological Society, London, Special Publications.* 153, 55-73.

1462 Drury, M.R., Vissers, R.L.M., Vanderwal, D., Strating, E.H.H., 1991. Shear
1463 Localization in Upper Mantle Peridotites. *Pure Appl Geophys.* 137, 439-460.

1464 Durham, W.B. and Goetze, C., 1977. Plastic flow of oriented single crystals of
1465 olivine. 1.Mechanical data. *J. Geophys. Res.* 82: doi:
1466 10.1029/JB082i036p05737

1467 Fei, H., Koizumi, S., Sakamoto, N., Hashiguchi, M., Yurimoto, H., Marquardt,
1468 K., Katsura, T., 2016. New constraints on upper mantle creep mechanism
1469 inferred from silicon grain-boundary diffusion rates. *Earth Planet. Sci. Lett.*
1470 433, 350-359.

1471 Fisler, D. K., Mackwell, S. J., Petsch, S., 1997. Grain boundary diffusion in

1472 enstatite. *Physics and Chemistry of Minerals*. 24, 264-273.

1473 Fodor, R.V., Sial, A.N., Gandhok, G., 2002. Petrology of spinel peridotite
1474 xenoliths from northeastern Brazil: lithosphere with a high geothermal
1475 gradient imparted by Fernando de Noronha plume. *J. S. Am. Earth Sci.* 15,
1476 199-214.

1477 French, S., Lekic, V., Romanowicz, B., 2013. Waveform Tomography Reveals
1478 Channeled Flow at the Base of the Oceanic Asthenosphere. *Science*. 342,
1479 227-230.

1480 Frets, E.C., Tommasi, A., Garrido, C.J., Vauchez, A., Mainprice, D., Targuisti,
1481 K., Amri, I., 2014. The Beni Bousera Peridotite (Rif Belt, Morocco): an
1482 Oblique-slip Low-angle Shear Zone Thinning the Subcontinental Mantle
1483 Lithosphere. *J. Petrol.* 55, 283-313.

1484 Ganade de Araujo, C.E., Weinberg, R.F. and Cordani, U.G., 2014. Extruding
1485 the Borborema Province (NE-Brazil): a two-stage Neoproterozoic collision
1486 process. *Terra Nova* 26, 157-168.

1487 Ganade de Araujo, C.E., Cordani, U.G., Agbossoumounde, Y., Caby, R., Basei,
1488 M.A.S., Weinberg, R.F. and Sato, K., 2016. Tightening-up NE Brazil and NW
1489 Africa connections: New U–Pb/Lu–Hf zircon data of a complete plate
1490 tectonic cycle in the Dahomey belt of the West Gondwana Orogen in Togo
1491 and Benin. *Precambrian Res.* 276, 24-42.

1492 Gourié, K., Cordier, P., Garel, F., Thoraval, C., Demouchy, S., Tommasi, A.,
1493 Carrez, P., 2019. Dislocation dynamics modelling of the power-law
1494 breakdown in olivine single crystals: Toward a unified creep law for the
1495 upper mantle. *Earth Planet. Sci. Lett.* 506, 282-291.

1496 Green, H.W., Gueguen, Y., 1974. Origin of Kimberlite Pipes by Diapiric
1497 Upwelling in Upper Mantle. *Nature*. 249, 617-620.

1498 Gripp, A., Gordon, R., 2002. Young tracks of hotspots and current plate
1499 velocities. *Geophys. J. Int.* 150, 321–361.

1500 Hammond, J.O.S., Kendall, J.M., Wookey, J., Stuart, G.W., Keir, D., Ayele, A.,
1501 2014. Differentiating flow, melt, or fossil seismic anisotropy beneath Ethiopia,
1502 *Geochem. Geophys. Geosyst.* (15) DOI:10.1002/2013GC005185

1503 Hamza, V.M., Vieira, F.P., Silva, R.T.A., 2018. Anomalous heat flow belt along
1504 the continental margin of Brazil. *Int J Earth Sci.* 107, 19-33.

1505 Hansen, L. N., Zhao, Y. H., Zimmerman, M. E., Kohlstedt, D. L., 2014.
1506 Protracted fabric evolution in olivine: Implications for the relationship among
1507 strain, crystallographic fabric, and seismic anisotropy. *Earth Planet. Sci. Lett.*
1508 387, 157-168

1509 Heine, C., Zoethout, J., Muller, R.D., 2013. Kinematics of the South Atlantic rift.
1510 *Solid Earth* 4, 215-253.

1511 Heit, B., Sodoudi, F., Yuan, X., Bianchi, M., Kind, R., 2007. An S receiver
1512 function analysis of the lithospheric structure in South America. *Geophys.*
1513 *Res. Lett.* 34.

1514 Hielscher, R., Schaeben, H., 2008. A novel pole figure inversion method:
1515 specification of the MTEX algorithm. *J. Appl. Crystallogr.* 41, 1024-1037.

1516 Higgle, K., Tommasi, A., 2012. Feedbacks between deformation and melt
1517 distribution in the crust–mantle transition zone of the Oman ophiolite. *Earth*
1518 *Planet. Sci. Lett.* 359, 61-72.

1519 Higgle, K., Tommasi, A., 2014. Deformation in a partially molten mantle:
1520 Constraints from plagioclase Iherzolites from Lanzo, western Alps.
1521 *Tectonophys.* 615, 167-181.

1522 Hirth, G., Kohlstedt, D., 2003. Rheology of the upper mantle and the mantle
1523 wedge: A view from the experimentalists. *Inside the subduction Factory*, 138,
1524 83-105.

1525 Hollanda, M.H.B.M., Archanjo, C.J., Bautista, J.R., Souza, L.C., 2015. Detrital
1526 zircon ages and Nd isotope compositions of the Seridó and Lavras da
1527 Mangabeira basins (Borborema Province, NE Brazil): Evidence for
1528 exhumation and recycling associated with a major shift in sedimentary
1529 provenance. *Precambrian Res.* 258, 186-207.

1530 Hollanda, M.H.B.M., Archanjo, C.J., Souza, L.C., Liu, D.Y., Armstrong, R.,
1531 2011. Long-lived Paleoproterozoic granitic magmatism in the
1532 Serido-Jaguaribe domain, Borborema Province-NE Brazil. *J. S. Am. Earth*
1533 *Sci.* 32, 287-300.

1534 Hollanda, M.H.B.M., Archanjo, C.J., Macedo Filho, A.A., Fossen, H., Ernst,
1535 R.E., Castro, D.L.d., Melo, A.C. and Oliveira, A.L., 2018. The Mesozoic
1536 Equatorial Atlantic Magmatic Province (EQUAMP), in: *Dyke Swarms of the*
1537 *World: A Modern Perspective*. Springer Singapore, pp. 87-110.

1538 Isaak, D.G., Ohno, I., Lee, P.C., 2006. The elastic constants of monoclinic
1539 single-crystal chrome-diopside to 1,300 K. *Phys Chem Miner.* 32, 691-699.

1540 Jackson, J.M., Sinogeikin, S.V., Bass, J.D., 2007. Sound velocities and
1541 single-crystal elasticity of orthoenstatite to 1073 K at ambient pressure.
1542 *Phys. Earth Planet. Inter.* 161, 1-12.

1543 Kaczmarek, M.A., Tommasi, A., 2011. Anatomy of an extensional shear zone
1544 in the mantle, Lanzo massif, Italy. *Geochem. Geophys. Geosyst.* 12,
1545 Q0AG06, doi:10.1029/2011GC003627.

1546 Kelemen, P.B., 1990. Reaction between Ultramafic Rock and Fractionating
1547 Basaltic Magma .1. Phase-Relations, the Origin of Calc-Alkaline Magma
1548 Series, and the Formation of Discordant Dunite. *J. Petrol.* 31, 51-98.

1549 Klemme, S., 2004. The influence of Cr on the garnet-spinel transition in the
1550 Earth's mantle: experiments in the system MgO-Cr₂O₃-SiO₂ and
1551 thermodynamic modelling. *Lithos* 77, 639-646.

1552 Knesel, K.M., Souza, Z.S., Vasconcelos, P.M., Cohen, B.E., Silveira, F.V., 2011.
1553 Young volcanism in the Borborema Province, NE Brazil, shows no evidence
1554 for a trace of the Fernando de Noronha plume on the continent. *Earth*
1555 *Planet. Sci. Lett.* 302, 38-50.

1556 Kostopoulos, D.K., 1991. Melting of the Shallow Upper Mantle - a New
1557 Perspective. *J. Petrol.* 32, 671-699.

1558 Kourim, F., Vauchez, A., Bodinier, J. L., Alard, O., Bendaoud, A., 2015.
1559 Subcontinental lithosphere reactivation beneath the Hoggar swell (Algeria):

1560 Localized deformation, melt channeling and heat advection. *Tectonophys.*
1561 650, 18-33.

1562 Lamarque, G., Julià, J., 2019. Lithospheric and sub-lithospheric deformation
1563 under the Borborema Province of NE Brazil from receiver function harmonic
1564 stripping. *Solid Earth Discuss.* <https://doi.org/10.5194/se-2019-41>.

1565 Le Roux, V., Bodinier, J.L., Tommasi, A., Alard, O., Dautria, J.M., Vauchez, A.,
1566 Riches, A.J.V., 2007. The Lherz spinel Iherzolite: Refertilized rather than
1567 pristine mantle. *Earth Planet. Sci. Lett.* 259, 599-612.

1568 Lima, M.V.A.G.d., Berrocal, J., Soares, J.E.P., Fuck, R.A., 2015. Deep seismic
1569 refraction experiment in northeast Brazil: New constraints for Borborema
1570 province evolution. *J. S. Am. Earth Sci.* 58, 335-349.

1571 Lissenberg, C.J., Dick, H.J.B., 2008. Melt–rock reaction in the lower oceanic
1572 crust and its implications for the genesis of mid-ocean ridge basalt. *Earth*
1573 *Planet. Sci. Lett.* 271, 311-325.

1574 Liu, S., Tommasi, A., Vauchez, A., Mazzucchelli, M. 2019. Deformation,
1575 annealing, melt-rock interaction, and seismic properties of an old domain of
1576 the equatorial Atlantic lithospheric mantle. *Tectonics* 38, doi:
1577 1029/2018TC005373.

1578 Luz, R.M.N., Julià, J., do Nascimento, A.F., 2015. Bulk crustal properties of the
1579 Borborema Province, NE Brazil, from P-wave receiver functions:
1580 Implications for models of intraplate Cenozoic uplift. *Tectonophys.* 644-645,
1581 81-91.

1582 Mainprice, D., Bachmann, F., Hielscher, R., Schaeben, H., 2014. Descriptive
1583 tools for the analysis of texture projects with large datasets using MTEX:
1584 strength, symmetry and components. *Geol. Soc. London Spec. Publ.* 409,
1585 251-271.

1586 Mainprice, D., Hielscher, R., Schaeben, H., 2011. Calculating anisotropic
1587 physical properties from texture data using the MTEX open-source package.
1588 *Geol. Soc. London Spec. Publ.* 360, 175-192.

1589 Marques, F.O., Nogueira, F.C.C., Bezerra, F.H.R., Castro, D.L., 2014. The
1590 Araripe Basin in NE Brazil: An intracontinental graben inverted to a
1591 high-standing horst. *Tectonophys.* 630, 251-264.

1592 Klöcking, M., Tribaldos, V.R., Hoggard, M., White, N., Maclennan, J., Rezende
1593 Guimarães, A., 2018. Linking uplift and volcanism of the Borborema
1594 Province, northeast Brazil. *EGU General Assembly Conference Abstracts*
1595 20, p. 9966.

1596 Mercier, J.C.C., Nicolas, A., 1975. Textures and Fabrics of Upper-Mantle
1597 Peridotites as Illustrated by Xenoliths from Basalts. *J. Petrol.* 16, 454-487.

1598 Mizusaki, A.M.P., Thomaz-Filho, A., Milani, E.J., Césero, P., 2002. Mesozoic
1599 and Cenozoic igneous activity and its tectonic control in northeastern Brazil.
1600 *J. S. Am. Earth Sci.* 15, 183-198.

1601 Moulin, M., Aslanian, D., Unternehr, P., 2010. A new starting point for the South
1602 and Equatorial Atlantic Ocean. *Earth-Sci Rev.* 98, 1-37.

1603 Neves, S.P., 2003. Proterozoic history of the Borborema province (NE Brazil):

1604 Correlations with neighboring cratons and Pan-African belts and
1605 implications for the evolution of western Gondwana. *Tectonics*. 22.

1606 Ngonge, E.D., de Hollanda, M.H.B.M., Archanjo, C.J., de Oliveira, D.C.,
1607 Vasconcelos, P.M.P., Muñoz, P.R.M., 2016a. Petrology of continental
1608 tholeiitic magmas forming a 350-km-long Mesozoic dyke swarm in NE Brazil:
1609 Constraints of geochemical and isotopic data. *Lithos*. 258-259, 228-252.

1610 Ngonge, E.D., de Hollanda, M.H.B.M., Pimentel, M.M., de Oliveira, D.C.,
1611 2016b. Petrology of the alkaline rocks of the Macau Volcanic Field, NE
1612 Brazil. *Lithos*. 266-267, 453-470.

1613 Nimis, P., Grutter, H., 2010. Internally consistent geothermometers for garnet
1614 peridotites and pyroxenites. *Contrib Mineral Petr.* 159, 411-427.

1615 Niu, Y.L., 1997. Mantle melting and melt extraction processes beneath ocean
1616 ridges: Evidence from abyssal peridotites. *J. Petrol.* 38, 1047-1074.

1617 Nogueira, F.C.C., Marques, F.O., Bezerra, F.H.R., Castro, D.L., Fuck, R.A.,
1618 2015. Cretaceous intracontinental rifting and post-rift inversion in NE Brazil:
1619 Insights from the Rio do Peixe Basin. *Tectonophysics* 644-645, 92-107.

1620 O'Neill, H.S.C., 1981. The transition between spinel lherzolite and garnet
1621 lherzolite, and its use as a geobarometer. *Contrib. Mineral. Petr.* 77,
1622 185-194.

1623 Oliveira, R.G., Medeiros, W.E., 2012. Evidences of buried loads in the base of
1624 the crust of Borborema Plateau (NE Brazil) from Bouguer admittance
1625 estimates. *J. S. Am. Earth Sci.* 37, 60-76.

1626 Perlingeiro, G., Vasconcelos, P.M., Knesel, K.M., Thiede, D.S., Cordani, U.G.,
1627 2013. Ar-40/Ar-39 geochronology of the Fernando de Noronha Archipelago
1628 and implications for the origin of alkaline volcanism in the NE Brazil. *J.*
1629 *Volcanol. Geotherm. Res.* 249, 140-154.

1630 Princivalle, F., Salviulo, G., Fabro, C., Demarchi, G., 1994. Inter- and
1631 intra-crystalline temperature and pressure estimates on pyroxenes from NE
1632 Brazil mantle xenoliths. *Contrib. Mineral. Petr.* 116, 1-6.

1633 Qi, C., Hansen, L. N., Wallis, D., Holtzman, B. K., Kohlstedt, D. L., 2018.
1634 Crystallographic Preferred Orientation of Olivine in Sheared Partially Molten
1635 Rocks: The Source of the "a-c Switch". *Geochem. Geophys. Geosyst.* 19,
1636 316-336

1637 Rivalenti, G., Mazzucchelli, M., Girardi, V.A.V., Vannucci, R., Barbieri, M.A.,
1638 Zanetti, A., Goldstein, S.L., 2000. Composition and processes of the mantle
1639 lithosphere in northeastern Brazil and Fernando de Noronha: evidence from
1640 mantle xenoliths. *Contrib. Mineral. Petr.* 138, 308-325.

1641 Rivalenti, G., Zanetti, A., Girardi, V.A.V., Mazzucchelli, M., Tassinari, C.C.G.,
1642 Bertotto, G.W., 2007. The effect of the Fernando de Noronha plume on the
1643 mantle lithosphere in north-eastern Brazil. *Lithos* 94, 111-131.

1644 Tribaldos, V. R., White, N.J., Roberts, G.G., Hoggard, M.J., 2017. Spatial and
1645 temporal uplift history of South America from calibrated drainage analysis.
1646 *Geochem. Geophys. Geosyst.* 18, 2321-2353.

1647 Sa, J.M., McReath, I., Leterrier, J., 1995. Petrology, Geochemistry and

1648 Geodynamic Setting of Proterozoic Igneous Suites of the Oros Fold Belt
1649 (Borborema Province, Northeast Brazil). *J. S. Am. Earth Sci.* 8, 299-314.
1650 Sang, L.Q., Bass, J.D., 2014. Single-crystal elasticity of diopside to 14 GPa by
1651 Brillouin scattering. *Phys. Earth Planet. Inter.* 228, 75-79.
1652 Santos, E.J., Schmus, W.R.V., Kozuch, M., Neves, B.B.d.B., 2010. The Cariris
1653 Velhos tectonic event in Northeast Brazil. *J. S. Am. Earth Sci.* 29, 61-76.
1654 Schutt, D.L., Lowry, A.R., Buehler, J.S., 2018. Moho temperature and mobility of lower
1655 crust in the western United States. *Geology* 46, 219-222.
1656 Silveira, F.V., 2006. Magmatismo cenozóico da porção central do Rio Grande
1657 do Norte, NE do Brasil. Ph.D. Thesis, Univ. Federal do Rio Grande do Norte,
1658 UFRN, Brazil.
1659 Simões Neto, F.L., Julià, J., Schimmel, M., 2019. Upper-mantle structure of the
1660 Borborema Province, NE Brazil, from *P*-wave tomography: implications for
1661 rheology and volcanism, *Geophys. J. Intern.* 216, 231-250.
1662 Skemer, P., Karato, S.I., 2008. Sheared Iherzolite xenoliths revisited. *J.*
1663 *Geophys. Res.* 113, B07205.
1664 Souza, Z.S., Vasconcelos, P., Nascimento, M., V. Silveira, F., S. Paiva, H., Dias,
1665 L.G., Thiede, D., Carmo, I., 2003. 40Ar/39Ar geochronology of Mesozoic
1666 and Cenozoic magmatism in NE Brazil. In Proceedings of IV South
1667 American Symposium on Isotope Geology, 691-694.
1668 Souza, Z.S., Kalsbeek, F., Deng, X.-D., Frei, R., Kokfelt, T.F., Dantas, E.L., Li,
1669 J.-W., Pimentel, M.M., Galindo, A.C., 2016. Generation of continental crust
1670 in the northern part of the Borborema Province, northeastern Brazil, from
1671 Archaean to Neoproterozoic. *J. S. Am. Earth Sci.* 68, 68-96.
1672 Taylor, W.R., 1998. An experimental test of some geothermometer and
1673 geobarometer formulations for upper mantle peridotites with application to
1674 the thermobarometry of fertile Iherzolite and garnet websterite. *Neues Jb.*
1675 *Miner. Abh.* 172, 381-408.
1676 Tikoff, B., Russo, R., Teyssier, C., Tommasi, A. 2004. Mantle-driven
1677 deformation of orogenic zones and clutch tectonics. In: J. Grocott, K.
1678 McCaffrey, G. Taylor & B. Tikoff (eds) Vertical and horizontal decoupling in
1679 the lithosphere. *Geol. Soc. London Spec. Publ.* 227, 41-64.
1680 Tommasi, A., Ishikawa, A., 2014. Microstructures, composition, and seismic
1681 properties of the Ontong Java Plateau mantle root. *Geochem. Geophys.*
1682 *Geosyst.*, 15, doi: 10.1002/2014GC005452
1683 Tommasi, A., Vauchez, A., 2015. Heterogeneity and anisotropy in the
1684 lithospheric mantle. *Tectonophys.* 661, 11-37.
1685 Tommasi, A., Vauchez, A., Daudré, B., 1995. Initiation and propagation of
1686 shear zones in a heterogeneous continental lithosphere. *J. Geophys. Res.*
1687 100, 22083-22101.
1688 Tommasi, A., Mainprice, D., Canova, G., Chastel, Y. 2000. Viscoplastic
1689 self-consistent and equilibrium-based modeling of olivine lattice preferred

1690 orientations. Implications for the upper mantle seismic anisotropy. *J.*
1691 *Geophys. Res.* 105, 7893-7908

1692 Tommasi, A., Tikoff, B., Vauchez, A. 1999. Upper mantle tectonics:
1693 Three-dimensional deformation, olivine crystallographic fabrics and seismic
1694 properties. *Earth Planet. Sci. Lett.* 168, 173-186

1695 Tommasi, A., Vauchez, A., Ionov, D. A., 2008. Deformation, static
1696 recrystallization, and reactive melt transport in shallow subcontinental
1697 mantle xenoliths (Tok Cenozoic volcanic field, SE Siberia). *Earth Planet. Sci.*
1698 *Lett.* 272, 65-77

1699 Ussami, N., Molina, E.C., Medeiros, W.E., 1999. Novos vinculos sobre a
1700 Evolução Térmica da Margem Continental Leste do Brasil, *In VII National*
1701 *Symposium on Tectonic Studies, Extended Abstracts*: Sociedade Brasileira
1702 de Geologia, 7, 20–23

1703 Van der Wal, D., Chopra, P., Drury, M., Gerald, J.F., 1993. Relationships
1704 between dynamically recrystallized grain size and deformation conditions in
1705 experimentally deformed olivine rocks. *Geophys. Res. Lett.* 20, 1479-1482.

1706 van Schmus, W.R., Oliveira, E.P., da Silva Filho, A.F., Toteu, S.F., Penaye, J.,
1707 Guimarães, I.P., 2008. Proterozoic links between the Borborema Province,
1708 NE Brazil, and the Central African Fold Belt. *Geol. Soc. London Spec. Publ.*
1709 294, 69-99.

1710 Vauchez, A., Nicolas, A., 1991. Mountain building: strike-parallel motion and
1711 mantle anisotropy. *Tectonophys.* 185, 183-201

1712 Vauchez, A., Dineur, F., Rudnick, R., 2005. Microstructure, texture and seismic
1713 anisotropy of the lithospheric mantle above a mantle plume: insights from
1714 the Labait volcano xenoliths (Tanzania). *Earth and Planetary Science*
1715 *Letters* 232, 295-314.

1716 Vauchez, A., Neves, S., Caby, R., Corsini, M., Egydio-Silva, M., Arthaud, M.,
1717 Amaro, V., 1995. The Borborema Shear Zone System, NE Brazil. *J. S. Am.*
1718 *Earth Sci.* 8, 247-266.

1719 Vauchez, A., Tommasi, A., Mainprice, D., 2012. Faults (shear zones) in the
1720 Earth's mantle. *Tectonophys.* 558, 1-27.

1721 Viegas, L.G.F., Archanjo, C.J., Hollanda, M.H.B.M. and Vauchez, A., 2014.
1722 Microfabrics and zircon U/Pb (SHRIMP) chronology of mylonites from the
1723 Patos shear zone (Borborema Province, NE Brazil). *Precambrian Research*
1724 243, 1-17.

1725 Walter, M.J., Sisson, T.W., Presnall, D.C., 1995. A Mass Proportion Method for
1726 Calculating Melting Reactions and Application to Melting of Model
1727 Upper-Mantle Lherzolite. *Earth Planet. Sci. Lett.* 135, 77-90.

1728 Workman, R.K., Hart, S.R., 2005. Major and trace element composition of the
1729 depleted MORB mantle (DMM). *Earth Planet. Sci. Lett.* 231, 53-72.

1730 Wu, J., Zhang, Z., Kong, F., Yang, B.B., Yu, Y., Liu, K.H., Gao, S.S., 2015.
1731 Complex seismic anisotropy beneath western Tibet and its geodynamic
1732 implications. *Earth Planet. Sci. Lett.* 413, 167-175.

1 **Crust-mantle coupling during continental convergence and**
2 **break-up: Constraints from peridotite xenoliths from the**
3 **Borborema Province, northeast Brazil**

4

5

6

7

8 **Shiran Liu^{1,2}, Andréa Tommasi¹, Alain Vauchez¹, Maurizio Mazzucchelli³**

9

10

11

12 ¹Géosciences Montpellier, CNRS & Université de Montpellier, Montpellier,
13 France.

14 ²Key Laboratory of Orogenic Belts and Crustal Evolution, School of Earth and
15 Space Sciences, Peking University, Beijing, China.

16 ³Dipartimento di Scienze Chimiche e Geologiche, Università di Modena e
17 Reggio Emilia, Modena, Italy.

18

19 Corresponding author: Shiran Liu (hbhslsr@gmail.com), Andréa Tommasi
20 (andrea.tommasi@umontpellier.fr)

21

22

23

24

25 **Keywords:**
26 Subcontinental mantle lithosphere
27 Crust-mantle coupling
28 Melt-rock interaction
29 Olivine crystal preferred orientation
30 Seismic anisotropy

31

32 **Highlights:**

- 33 • Mainly coarse-granular and porphyroclastic peridotites with [100](010)
34 olivine textures
- 35 • Equilibrium temperatures consistent with a 80 km thick lithosphere in the
36 Cenozoic
- 37 • Extensive annealing implies that deformation frozen in the shallow mantle
38 is 100s Ma old
- 39 • Mantle fabric related to Neoproterozoic shear zones may partially explain
40 SKS splitting
- 41 • Limited reworking of the lithospheric mantle during Cretaceous rifting

42

43

44 **Abstract**

45 We studied a suite of mantle xenoliths carried by Cenozoic volcanism in the
46 Borborema Province, NE Brazil. These xenoliths sample a subcontinental
47 lithospheric mantle affected by multiple continental convergence and rifting
48 events since the Archean. Equilibrium temperatures indicate a rather hot
49 geotherm, implying a ca. 80 km thick lithosphere. Most xenoliths have
50 coarse-granular and coarse-porphyroclastic microstructures, recording
51 variable degrees of annealing following deformation. The high annealing
52 degree and equilibrated pyroxene shapes in coarse-granular peridotites
53 equilibrated at ~900°C indicate that the last deformation event that affected
54 these peridotites is several hundreds of Ma old. Coarse-porphyroclastic
55 peridotites equilibrated at 950-1100°C probably record younger (Cretaceous?)
56 deformation in the deep lithospheric mantle. In addition, a few xenoliths show
57 fine-porphyroclastic microstructures and equilibrium temperatures $\geq 1200^\circ\text{C}$,
58 which imply a recent deformation, probably related to the dykes that fed the
59 Cenozoic volcanism. Chemical and microstructural evidence for reactive
60 percolation of melts is widespread. Variation in textural and chemical
61 equilibrium among samples implies multiple melt percolation events well
62 spaced in time (from Neoproterozoic or older to Cenozoic). Crystal preferred
63 orientations of olivine and pyroxenes point to deformation controlled by
64 dislocation creep with dominant activation of the [100](010) and [001]{0kl} slip
65 systems in olivine and pyroxenes for all microstructures. Comparison of
66 xenoliths' seismic properties to SKS splitting data in the nearby RCBR station
67 together with the equilibrated microstructures in the low-temperature xenoliths

68 point to coupled crust-mantle deformation in the Neoproterozoic (Brasiliano)
69 continental-scale shear zones, which is still preserved in the shallow
70 lithospheric mantle. This implies limited reworking of the lithospheric mantle in
71 response to extension during the opening of the Equatorial Atlantic in the
72 Cretaceous, which in the present sampling is restricted to the base of the
73 lithosphere.

74

75 **1. Introduction**

76 Continental plates have long-lived histories. They are usually composed
77 by crustal domains with varied tectonic ages. Tectonic reworking is a common
78 feature in crustal exposures. Yet, major tectonic events such as convergence,
79 collision, and rifting involve the entire plate, not only the crust. By consequence,
80 large volumes of the subcontinental lithospheric mantle must have been
81 affected by a succession of tectono-thermal events (cf. reviews in Tommasi
82 and Vauchez, 2015; Vauchez et al., 2012). In addition, during orogenic events,
83 the strain regime frequently changes across the belt, with contiguous domains
84 accommodating transcurrent motions and others accommodating thrusting for
85 instance. However, how the deformation is accommodated in the lithospheric
86 mantle and the level of coupling between crust and mantle deformation during
87 these major tectonic episodes are still matter of debate.

88 Comparison between SKS splitting data and crustal structures points to
89 coherent kinematics, implying at least partial coupling between the crust and
90 the mantle in many orogenic belts (Tikoff et al., 2004). As recognized since the
91 early times of seismic anisotropy measurements (e.g., Vauchez et al., 1991),
92 crust-mantle coupling is well documented beneath large-scale strike slip faults
93 and transpressional belts, such as the San Andreas fault (Bonnin et al. 2012)
94 or the Pyrenees, the Appalachians, and the neoproterozoic Ribeira-Aracuai
95 belt of SE Brazil (cf. review in Vauchez et al., 2012). In contrast, in collisional
96 domains submitted to large amounts of thickening, like the Alps and the
97 Southern Tibet, or in active continental rifts, such as the East African rift
98 system, polarization directions of fast split SKS or SKKS waves are usually
99 parallel to the trend of the belts or rifts, but at high angle to the lineations
100 marking flow directions in the ductile crust and to the slip directions in active
101 faults (e.g., Barruol et al., 2011; Wu et al., 2015; Hammond et al., 2014). This
102 suggests at least partial decoupling between the crust and lithospheric mantle
103 (Tommasi et al., 1999).

104 However, seismic anisotropy data only offer indirect information on the
105 present-day structure of the upper mantle. Moreover, unless a variety of
106 methods based on waves with different incidence angles and propagation
107 directions are employed, localizing vertically the source of the anisotropy
108 remains difficult. Analysis of mantle xenoliths, which are mantle fragments
109 (peridotite xenoliths) carried to the surface by volcanic episodes, allows for
110 direct sampling of the lithospheric mantle. This sampling is imperfect: it is
111 discontinuous, non-oriented, and focused along the magmatic conduits.

112 Nevertheless, the characterization of the microstructures and associated
113 crystal preferred orientations, as well as of the chemical compositions and
114 mineralogy of the mantle xenoliths, allows determining the relative deformation,
115 annealing, and petrological (partial melting, reactive melt transport, solid-state
116 reequilibration) history, even if no dating is possible. Coupling petrophysical
117 analyses on mantle xenoliths to seismological data may allow better
118 constraining the deformation history of the lithospheric mantle and hence
119 discussing crust-mantle coupling during major tectonic events.

120 In this article, we present a petrostructural study of peridotite xenoliths
121 entrained by Cenozoic volcanism in NE Brazil. Based on these data and on
122 published SKS splitting measurements (Assumpção et al., 2011; Bastow et al.,
123 2015), we try to unravel the tectono-thermal evolution of the continental
124 lithospheric mantle in this region, which has been affected by multiple
125 collisional and extensional episodes since the Archean. Analysis of the
126 presently outcropping crustal structures highlights a series of
127 tectono-magmatic episodes, among which the most important and recent ones
128 are: (i) extensive intraplate deformation in a convergent setting, which
129 produced a continental-scale system of strike-slip shear zones and
130 transpressional belts accommodating lateral escape of this domain during the
131 formation of the Gondwana and (ii) extension, localized in a series of
132 intracontinental basins, during the early stages of the opening of the Equatorial
133 Atlantic.

134

135 **2. Geological and geophysical background**

136 *2.1 Evolution of the Northern Borborema Province (NBP)*

137 The analysis of the crustal rocks outcropping in the Northern Borborema
138 Province (NBP) points to a complex tectonic evolution, which probably started
139 in the Archean, as indicated by U-Pb ages ranging from 3.4 to 2.7 Ga recorded
140 in the São José Massif (Souza et al., 2016), ca. 100 km to the east of the study
141 area and in smaller nuclei elsewhere in the province. Between 2.1-2.4 Ga, a
142 major crust formation event produced ca. of the 50% of the present-day
143 outcropping rocks (Hollanda et al., 2011; Souza et al., 2016). The tectonic
144 setting for this extensive magmatic activity is discussed, but isotopic data
145 indicates significant recycling of pre-existing crustal material (Hollanda et al.,
146 2011).

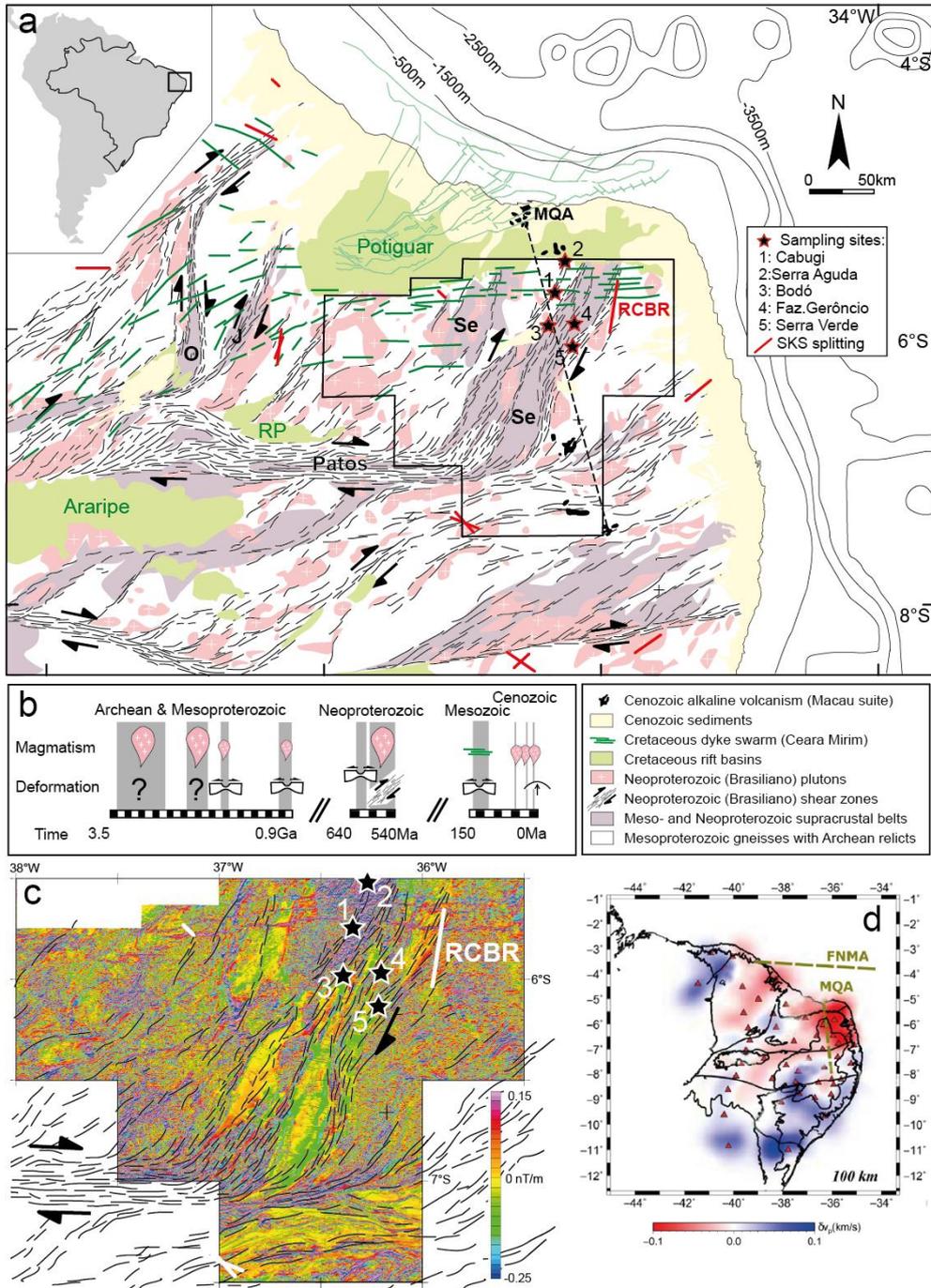
147 The Meso- to Neoproterozoic evolution of the Borborema Province is
148 characterized by a series of failed intracontinental rifting episodes. At 1.8-1.9
149 Ga, the Orós and Jaguaribe volcano-sedimentary basins formed to the west of
150 the study area (Fig. 1). Localized extension was accompanied by intrusion of
151 small volumes of alkaline granites and anorthosites elsewhere in the province,
152 which extended until 1.75 Ga (Sa et al., 1995; Hollanda et al., 2011). At 1-0.9
153 Ga, extension affected the central Borborema Province (Santos et al., 2010;
154 Van Schmus et al., 2008). The last Proterozoic extensional episode affected
155 most of the province and formed the Seridó basin within the study area (Fig. 1).

156 The youngest sedimentation in this basin occurred at ~630-615 Ma, but early
157 sedimentation may be as old as 1.8 Ga (Hollanda et al., 2015), suggesting an
158 early rifting episode simultaneous with the formation of the Orós and Jaguaribe
159 basins.

160 The major tectono-magmatic episode that affected the Borborema
161 Province, shaping its present-day geology, is, however, the compressional
162 Brasiliano event between 595-540 Ma (Neves, 2003; Archanjo et al. 2008;
163 Viegas et al., 2014). This event resulted in extensive tectono-thermal
164 reworking of the province through the formation of a continental-scale
165 transcurrent shear zone system – the Borborema shear zone system,
166 transpressional inversion of the Proterozoic metasedimentary basins, and
167 widespread emplacement of granitoids (Fig. 1). The granitoids geochemistry
168 indicates a dominant component of crustal reworking (Neves, 2003; Souza et
169 al., 2016), implying an intraplate setting for the Brasiliano deformation, with a
170 possible convergent plate boundary > 500 km to the NW from the study area
171 (e.g., Caby et al., 1991; Ancelmi et al., 2015). The complex network of
172 anastomosing E-W and NE-SW trending ductile dextral strike-slip shear zones,
173 some up to ~25 km wide, and transpressional belts deformed under high
174 temperature, low pressure conditions that compose the Borborema shear zone
175 system (Vauchez et al., 1995) is therefore the expression of strain localization
176 in an intraplate setting controlled by large-scale intraplate rheological
177 heterogeneities (the basins) in response to the convergence between São
178 Francisco, Amazonian, and West African cratons during the construction of the
179 Gondwana (Tommasi et al., 1995). This shear zone system allowed for
180 tectonic extrusion of the Borborema Province towards the NE (Ganade de
181 Araújo et al., 2014, 2016), that is, towards the Saharan province, which, like
182 Borborema, had been weakened by tectono-magmatic activity during the
183 Meso- and Neoproterozoic.

184 In the early Cretaceous, the Borborema province experienced regional
185 extension before rifting localized at the present location of the equatorial
186 Atlantic Ocean (Darros de Matos, 1999). Structures associated with these
187 early stages of rifting in the northern Borborema Province include several
188 intracontinental basins or aborted rifts and a large tholeiitic dyke swarm. The
189 basins formed as NW-SE trending asymmetric grabens (e.g., Potiguar, Rio do
190 Peixe, Araripe in Fig. 1) controlled by reactivation of the Brasiliano shear zones
191 (Darros de Matos, 1999; Castro et al., 2007, 2012; Marques et al., 2014). They
192 contain dominantly continental sediments of fluvial and lacustrine origin,
193 which attain thicknesses of 2000 m, deposited between 145-125 Ma. The
194 Potiguar basin is the best developed among these basins; its offshore domain
195 is the only one involved in the final opening of the Atlantic. The Ceará-Mirim
196 dyke swarm is a >350-km-long tholeiitic dyke swarm with a dominantly E-W
197 trend that rotates to NE-SW at the southern termination of the Potiguar basin
198 (Fig. 1). The dykes have ages between 135 Ma and 120 Ma (Hollanda et al.,
199 2018 and references therein). Their emplacement is therefore synchronous to

200 the formation of the intracontinental basins. Geochemical data points to
 201 primary magmas formed by melting of an enriched mantle (EMI \pm FOZO
 202 isotopic signatures) at shallow depths (<90 km; Ngonge et al., 2016).
 203



204
 205 Fig. 1. (a) Simplified geological map (after Vauchez et al., 1995 and Oliveira and Medeiros, 2018) displaying xenolith
 206 sampling locations. (b) Time line of the major tectono-magmatic events in the Borborema Province. (c) Aeromagnetic
 207 anomalies map (1st vertical derivative; Costa et al., 2016) imaging middle crust structures in and around the Seridó belt.
 208 (d) P-wave tomography model for the Borborema Province at 100 km depth (Simões Neto et al., 2018) characterized
 209 by a marked low velocity anomaly east of the Macau-Queimadas volcanic alignment (MQA) and north of the Patos
 210 shear zone. SKS splitting data in (a) and (c) from Assumpção et al. (2011) and Bastow et al. (2015). RP: Rio do Peixe

212 The Cenozoic is characterized by a long-lived, but small-volume alkaline
213 volcanism (both onshore and offshore), tectonic inversion of the Mesozoic
214 intracontinental basins, and uplift of the Borborema Plateau. The entire
215 Borborema Province sits nowadays on average at ~800m above the sea level
216 and the Araripe basin is topographically more than 500 m above the
217 Precambrian basement. Analysis of river profiles implies that most uplift
218 occurred within the last 30 Ma (Tribaldos et al., 2017), but earlier events may
219 have occurred (Marques et al., 2014; Nogueira et al., 2015). The present day
220 crustal thickness of 30-35 km implies some degree of sub-lithospheric support
221 for the uplift. The offshore Cenozoic volcanism forms an E-W volcanic
222 alignment that extends from the Ceará coast in northeast Brazil (the Mecejana
223 volcanic field) to the Fernando de Noronha archipelago. The Mecejana
224 volcanics yield K-Ar ages between 26 and 44 Ma (Mizusaki et al., 2002),
225 whereas volcanism in Fernando de Noronha has Ar-Ar ages between 12.5 and
226 6 Ma (Perlingeiro et al., 2013). The onshore Cenozoic volcanism is distributed
227 along a N-S trend, forming the Macau–Queimadas Alignment (MQA, Fig. 1).
228 Emplacement ages range between 52 Ma and 7 Ma, with two peaks around
229 26-29 Ma and 7-9 Ma and no clear age progression along the MQA (Souza et
230 al., 2003; Silveira 2006; Knesel et al., 2011).

231 The Macau volcanics that entrained the studied mantle xenoliths (Fig. 1)
232 erupted onto metasediments from the Neoproterozoic Seridó belt or
233 Mesoproterozoic gneisses that form their basement. In the study area, the
234 Brasiliano event is recorded by multiple NE-SW shear zones, which branch off
235 from the E-W Patos shear zone in the south, by transpressional structures in
236 the Seridó belt, and by intrusion of granitoids. The Cretaceous Ceará-Mirim
237 dyke swarm crosscuts the study area with an E-W trend and the Potiguar basin
238 outcrops to the north and west of it (Fig. 1).

239

240 *2.2 Geophysical data*

241 P-wave receiver functions, deep seismic refraction experiments, and
242 surface-wave dispersion data indicate that the crust is 30 ~ 32km thick in the
243 northern Borborema Province, but 36 ~ 38 km thick in the southern part of the
244 province (Oliveira and Medeiros, 2012; Almeida et al., 2015; Lima et al., 2015;
245 Luz et al., 2015). This variation in crustal thickness is consistent with gravity
246 data (Oliveira and Medeiros, 2018) and correlated with topography: elevated
247 regions show thicker crust. Receiver function data also imply an intra-crustal
248 discontinuity at 9-18 depth, most often observed in domains with thinner crust,
249 such as the area sampled by the studied xenoliths (Almeida et al., 2015).

250 Gravity and magnetic data illuminate the structuration of the deep crust in
251 the Borborema province (Oliveira and Medeiros, 2018). These data establish
252 the Patos, Pernambuco, and Jaguaribe shear zones as major structures,
253 which splay off from the Transbrasiliano Lineament, separating four crustal
254 blocks with different geophysical characteristics: the Southern, the Transversal,

255 the Ceará, and the Rio Grande do Norte domains. The studied xenoliths
256 sample the mantle beneath the Rio Grande do Norte domain, which is
257 delimited by the Jaguaribe and Patos shear zones to the west and south, and
258 by the Atlantic Ocean to the north and east. This domain is characterized by
259 strong magnetic contrasts with E-W or NNE trends, correlated with the
260 outcropping Brasiliano structures (Fig. 1c). Magnetic anomalies also clearly
261 define the Ceará-Mirim dykes (Fig. 1c). In addition, analysis of the Bouguer
262 gravity anomaly shows that most of the Rio Grande do Norte domain is
263 characterized by weak long wavelength positive anomalies, which follow the
264 trend of the coastline (cf. Fig. 7 of Oliveira and Medeiros, 2018), suggesting
265 that crustal thinning related to the Atlantic opening affected most of the
266 domain.

267 The first kilometers of the lithospheric mantle structure may be probed by
268 the analysis of Pn velocities, which decrease from 8.1 km/s to 7.9 km/s in a
269 NW-SE refraction profile across the Borborema Province to the south of the
270 study area (Lima et al., 2015). In the vicinity of the study area, Pn velocities are
271 ~8.0 km/s, which for an isotropic mantle with a spinel-lherzolite composition
272 would correspond to sub-Moho temperatures of 700-750°C (Schutt et al.,
273 2018).

274 No seismic velocity data is available for the deeper sections of the
275 lithospheric mantle, but a recent P-wave regional travel-time tomography
276 model (Simões Neto et al., 2019) imaged slower than average velocities at
277 depths <150 km in the northeastern Borborema province, just east of the study
278 area (Fig. 1d). This low velocity anomaly coincides with a local geoid anomaly
279 of +10 m (Ussami et al. 1999). Slower than average P-wave velocities in the
280 shallow mantle beneath the study area are coherent with a rather shallow
281 Lithosphere Asthenosphere boundary (LAB), estimated at 80 km depth based
282 on S receiver function data from the station RCBR (Heit et al., 2007). The
283 results of the regional P-wave tomography for the northeastern Borborema
284 Province are also consistent with global finite-frequency tomography models,
285 which image lower than average S-wave velocities in the mantle at 250-km
286 depth in this region (French et al. 2013).

287 SKS splitting data in the Borborema Province (Bastow et al., 2015) show a
288 complex pattern, with highly variable delay times and fast polarization
289 directions, which in some places closely follow the Neoproterozoic crustal
290 fabric, but are oblique to it in others (Fig. 1a,c). The station closest to the
291 xenolith sampling sites, RCBR (Fig. 1a,c), shows a NNE-oriented fast S-wave
292 polarization subparallel to the Brasiliano transpressive structures, which
293 structure the northern Seridó belt and its gneissic basement, and a high delay
294 time (1.9 ± 0.2 s) based on 9 individual measurements (Assumpção et al.,
295 2011).

296

297 *2.3 Previous studies of Macau peridotite xenoliths*

298 The petrology and the geochemistry of Macau mantle xenoliths were

299 studied by Comin-Chiaramonti et al. (1986), Princivalle et al. (1994), Fodor et
300 al. (2002), Rivalenti et al. (2000; 2007). These studies identified two
301 microstructural groups: protogranular (the dominant type) and (rare)
302 porphyroclastic (partially recrystallized) peridotites. Protogranular
303 microstructures are mainly observed in Iherzolites, whereas porphyroclastic
304 microstructures are usually observed in harzburgites. Protogranular peridotites
305 also yield lower two-pyroxene equilibrium temperatures ($825\pm 116^{\circ}\text{C}$) than
306 porphyroclastic ones ($>1150^{\circ}\text{C}$). Trace-element patterns and isotopic
307 compositions of the peridotites indicate variable degrees of metasomatism by
308 alkali-basaltic melts with EMI and minor EMII isotopic signatures, probably
309 during multiple metasomatic events (Rivalenti et al., 2000, 2007).

310

311 **3. Methods**

312 *3.1. Sampling*

313 We have studied 22 xenoliths from 5 different Macau volcanic sites, which
314 extend in a rough N-S trend, from the limit of the Potiguar basin to ~70 km
315 south of it (Fig. 1). Most xenoliths analyzed in the present study come from the
316 Pico do Cabugi basaltic neck ($5^{\circ}42.3'S$, $36^{\circ}19.4'W$). The Pico do Cabugi
317 basalts yield Ar-Ar ages 24.6 ± 0.8 Ma (Souza et al., 2003 and references
318 therein). Among the 16 xenoliths of the Pico do Cabugi in this study, 14
319 (marked as CA) have been collected by the authors in a field campaign in 2016
320 and two (marked as PC) were previously studied by Rivalenti et al. (2000).

321 We also analyzed 6 xenoliths previously studied by Rivalenti et al. (2007):
322 two from the Serra Aguda neck (AG, $5^{\circ}31'S$, $36^{\circ}17'W$), ~30 km north of the
323 Pico do Cabugi, at the border of the Potiguar basin (Fig. 1), one from the Serra
324 Preta do Bodó dykes (BO, $5^{\circ}58'S$, $36^{\circ}22'W$), ~20 km south of Pico do Cabugi,
325 and four from Fazenda Geroncio (GR, $5^{\circ}58'S$, $36^{\circ}14'W$) and Serra Verde
326 localities (SV, $6^{\circ}06'S$, $36^{\circ}12'W$), located ~30 and ~50 km south of Pico do
327 Cabugi (Fig. 1). There are no ages for the Fazenda Geroncio and Serra Verde
328 volcanics, but the Serra Aguda basalts were dated at 26 Ma (Silveira, 2006)
329 and those at Serra Preta do Bodó, at 7.1 Ma (Knesel et al, 2010).

330 Most xenoliths are small (on average 3 cm of diameter), but they are very
331 fresh. The samples display no macroscopic evidence for major interaction with
332 the host lava. However, small pockets of fine-grained clinopyroxene, spinel,
333 and olivine, as well as melt films along grain boundaries were observed under
334 the microscope, in particular close to the borders of many xenoliths. In addition,
335 some xenoliths have pyroxenes with spongy borders, indicative of limited
336 partial melting during extraction. These domains were carefully avoided during
337 microprobe analyses. They were sometimes included in the EBSD analyses,
338 but affect weakly the results because of their small area.

339

340 *3.1 Electron-backscattered diffraction (EBSD) data acquisition and treatment*

341 Petrostructural analyses were performed on all samples. Crystallographic
342 preferred orientations (CPOs) of olivine, pyroxenes, and spinel were measured

343 by indexing of electron-backscattered diffraction patterns in the SEM-EBSD
344 facility at Geosciences Montpellier, France. Data acquisition was performed
345 using a JEOL JSM 5600 scanning electron microscope with 17kV acceleration
346 voltage and 24mm working distance. We performed EBSD mapping over the
347 entire thin section (areas $\geq 2 \times 3 \text{ cm}^2$) with step sizes between 14 μm and 35 μm ,
348 depending on grain size. Indexation rates varied between 84% and 97%.
349 Non-indexed points correspond mainly to fractures. During post-acquisition
350 data treatment, we eliminated inaccurate indexing points (MAD $> 1.3^\circ$ and wild
351 spikes), corrected for rare olivine pseudo-symmetry misindexing, and filled up
352 non-indexed pixels with at least 6 coherent neighboring measurements with
353 the average of the neighbors' orientations.

354 The CPO data analysis, that is, the calculation of the orientation
355 distribution functions (ODF) and of the misorientations, the computation of the
356 strength and symmetry of the olivine CPO, the plotting of pole figures, and the
357 quantification of the microstructure (grains' size, shape, and orientation) was
358 performed using the MTEX toolbox in Matlab (<http://mtex-toolbox.github.io/>;
359 Hielscher and Schaeben, 2008; Bachmann et al., 2010; Bachmann et al.,
360 2011). The ODFs were calculated using a "de la Vallée Poussin" kernel
361 function with a half-width of 10° . The CPO is presented as one crystallographic
362 orientation per pixel in pole figures (lower hemispheric stereographic
363 projections). Thin sections were cut in random orientations, because of the
364 small size of the xenoliths. To facilitate comparison between samples, we
365 rotated the CPO of all samples into a common reference frame, in which the
366 maximum concentration of [100] of olivine is parallel to the E-W direction and
367 the maximum concentration of [010] axis of olivine is parallel to the N-S
368 direction of the pole figure. The choice of this reference frame can be justified
369 a-posteriori by the analysis of the CPO, which indicates that this reference
370 frame probably corresponds to the lineation and normal to the foliation
371 directions (cf. discussion).

372 The strength of the CPO was quantified using the dimensionless J-index
373 (Bunge, 1982). The J-index for a random CPO is 1; it can reach up to 20 in
374 natural peridotites, with a peak at 3-5 (Tommasi and Vauchez, 2015). We use
375 the dimensionless BA-index, based on the eigenvalues of the [100] and [010]
376 axes orientation distributions (Mainprice et al., 2014), to characterize the
377 olivine CPO symmetry. This index allows classifying the olivine CPO into three
378 types: (i) fiber-[010], characterized by a [010] point maxima and girdle
379 distributions of [100] and [001] (BA-index < 0.35), (ii) orthorhombic,
380 characterized by point maxima of [100], [010], and [001]
381 ($0.35 < \text{BA-index} < 0.65$) and (iii) fiber-[100], characterized by a point maxima
382 of [010] and girdle distributions of [010] and [001] (BA-index > 0.65). J- and
383 BA-indexes do not depend on the reference frame.

384 To analyze quantitatively the microstructure, we used the grain detection
385 method in MTEX (Bachmann et al., 2011) defining grains boundaries by
386 misorientation angles between adjacent pixels higher than 15° . Grains

387 composed by less than 10 pixels were not considered in the microstructural
388 analysis. The misorientation of each pixel relative to the mean orientation of
389 the grain (M2M) and the grain orientation spread (GOS) were calculated to
390 quantify the intracrystalline orientation gradients, which are a proxy of the
391 dislocation density. In addition, we characterized the sinuosity of the grain
392 boundaries by the shape factor, which is the ratio of the perimeter of the grain
393 over the perimeter of a circle with the same area. Both the intra-granular
394 misorientation (M2M and GOS) and the sinuosity of grain boundaries (shape
395 factor) should decrease in response to recrystallization.

396

397 *3.2 Mineral compositions and equilibrium temperatures*

398 Mineral compositions of olivine, orthopyroxene, clinopyroxene, and spinel
399 were analyzed in ten samples, selected based on their microstructure, so that
400 all microstructural types were represented. Measurements were performed in
401 a Cameca SX100 electron microprobe at the Microsonde Sud facility at the
402 University of Montpellier, France at a 20kV accelerating voltage and a 10nA
403 current. For each sample, we analyzed both rim and core composition in 3 to 4
404 grains for olivine, orthopyroxene, clinopyroxene, and spinel.

405 We calculated equilibrium temperatures based on the
406 clinopyroxene-orthopyroxene geothermometer (Taylor, 1998) and on the
407 Ca-in-opx geothermometer (Brey and Kohler, 1990; revised by Nimis and
408 Grutter, 2010). Average rim and core equilibrium temperatures were calculated
409 by averaging the temperatures calculated using 3-4 rim or core compositions
410 of individual opx grains or opx-cpx pairs for each sample. Since no reliable
411 barometers are available for spinel-facies peridotites, we chose arbitrarily 1.5
412 GPa (~46 km depth) as the pressure to calculate the two thermometers.
413 Changes in the assumed pressure produce a variation of ~10 °C per 0.5 GPa
414 (~11 km).

415

416 *3.3 Seismic properties*

417 Seismic properties of each sample were computed using the MTEX
418 toolbox (Mainprice et al., 2011), using Voigt-Reuss-Hill averaging based on the
419 CPOs and modal content of olivine, orthopyroxene, and clinopyroxene derived
420 from the EBSD maps and on the single crystal elastic constant tensors of the
421 three minerals and their temperature and pressure derivatives (Abramson et
422 al., 1997; Anderson et al., 1992; Chai et al., 1997; Isaak et al., 2006; Jackson
423 et al., 2007; Sang and Bass, 2014). Average seismic properties for the
424 lithospheric mantle beneath the study area were estimated by averaging the
425 elastic constant tensors of all samples with all CPO data rotated into a
426 common reference frame. The assumption is that the orientation of the
427 lineation and foliation is constant both laterally and vertically. The average
428 seismic properties provide therefore an estimate of the maximum seismic
429 anisotropy that could be produced, if the xenolith sampling is representative of
430 the variability of compositions and CPO in the lithospheric mantle beneath the

431 studied area.

432

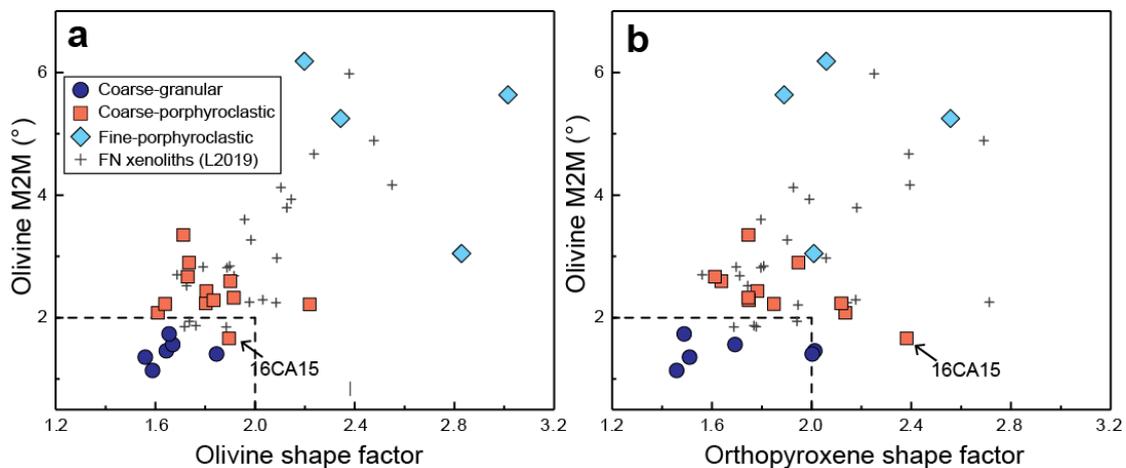
433 4. Data

434

435 4.1 Microstructures

436 Previous studies (Fodor et al., 2002; Rivalenti et al., 2000; Rivalenti et al.,
437 2007) have correlated optical observations of the microstructures with
438 chemical compositions and equilibrium temperatures to divide the Macau
439 peridotite xenoliths into two groups: protogranular and porphyroclastic. In the
440 present study, we associated to the optical observations quantitative analysis
441 of the microstructural data derived from EBSD mapping. Based on these data,
442 in particular the intragranular misorientation of olivine and the olivine and
443 orthopyroxene grain shapes (Fig. 2), we classify the Macau peridotites into
444 three microstructural groups (Table 1): coarse-granular (6),
445 coarse-porphyroclastic (12), and fine-porphyroclastic (4). The coarse-granular
446 and coarse-porphyroclastic groups correspond to a continuous variation in
447 microstructure (Figs. 2 and 3). They were described as protogranular in the
448 previous studies. The fine-porphyroclastic group is clearly different from the
449 other two groups (Figs. 2 and 3); it was described as porphyroclastic in the
450 previous studies.

451



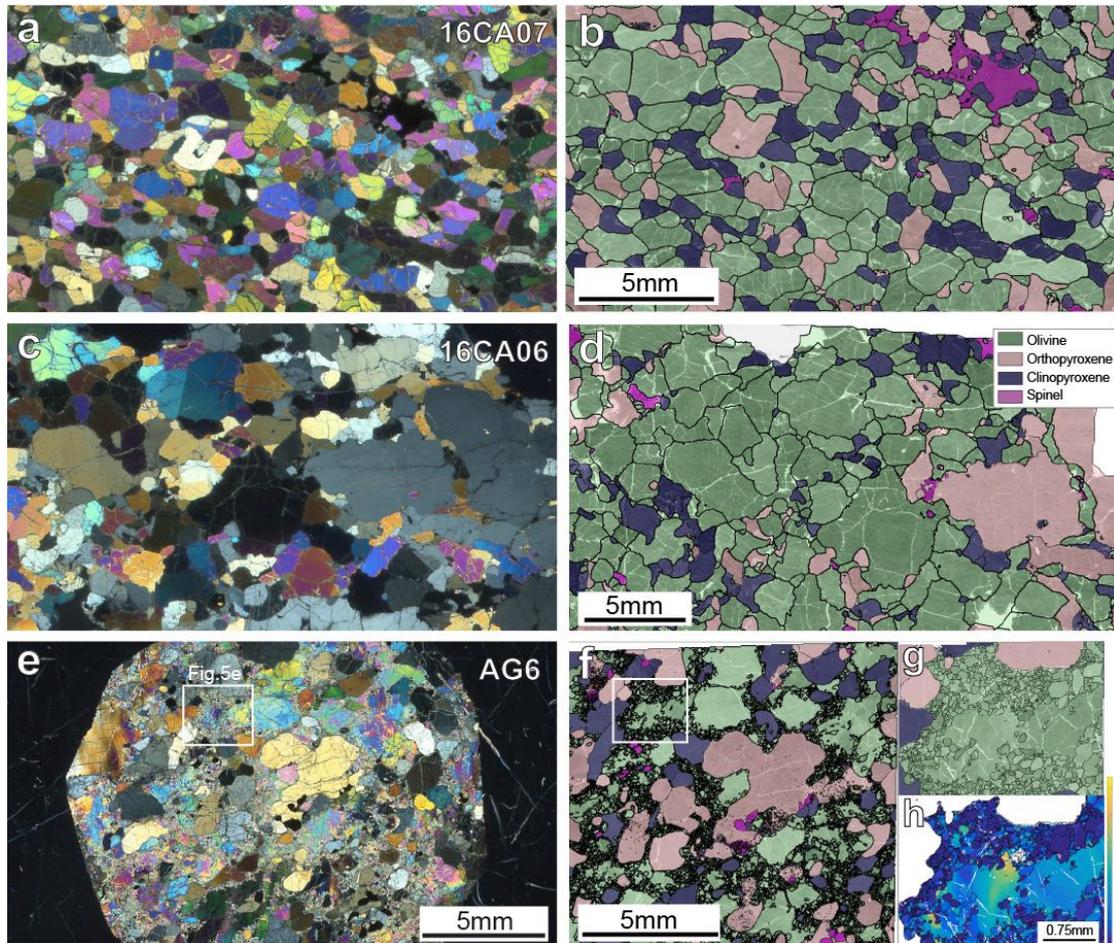
452

453 Fig. 2. Microstructural parameters calculated from EBSD data: (a) Olivine shape factor vs. intragranular misorientation
454 relative to the mean orientation of the grain (M2M); (b) Orthopyroxene shape factor vs. olivine M2M. All quantities are
455 average values at the sample scale weighted by the grains' area (cf. Table 2). Similar data for Fernando de Noronha
456 (FN) xenoliths, which sample an old domain of the equatorial Atlantic (Liu et al., 2019), are presented for comparison.

457

458 There is no relation between microstructure and sampling site. All three
459 microstructures are represented among the Pico do Cabugi peridotites, which
460 is the best-sampled site in the present study, and in at least one of the other
461 sampling localities (Table 1). This observation is corroborated by the previous
462 studies, which analyzed a larger number of peridotite xenoliths from other
463 Macau volcanic centers (Fodor et al. 2002; Rivalenti et al. 2007). Coarse

464 granular and coarse porphyroclastic microstructures predominate in all sites.
 465 Fine-porphyroclastic peridotites are always minor. They were only recovered in
 466 three sites, which are nevertheless >60 km apart (Fig. 1): Serra Aguda, Cabugi,
 467 and Serrote Preto (a small plug ca. 30 km to the SSE of Pico do Cabugi, Fodor
 468 et al. 2002).
 469



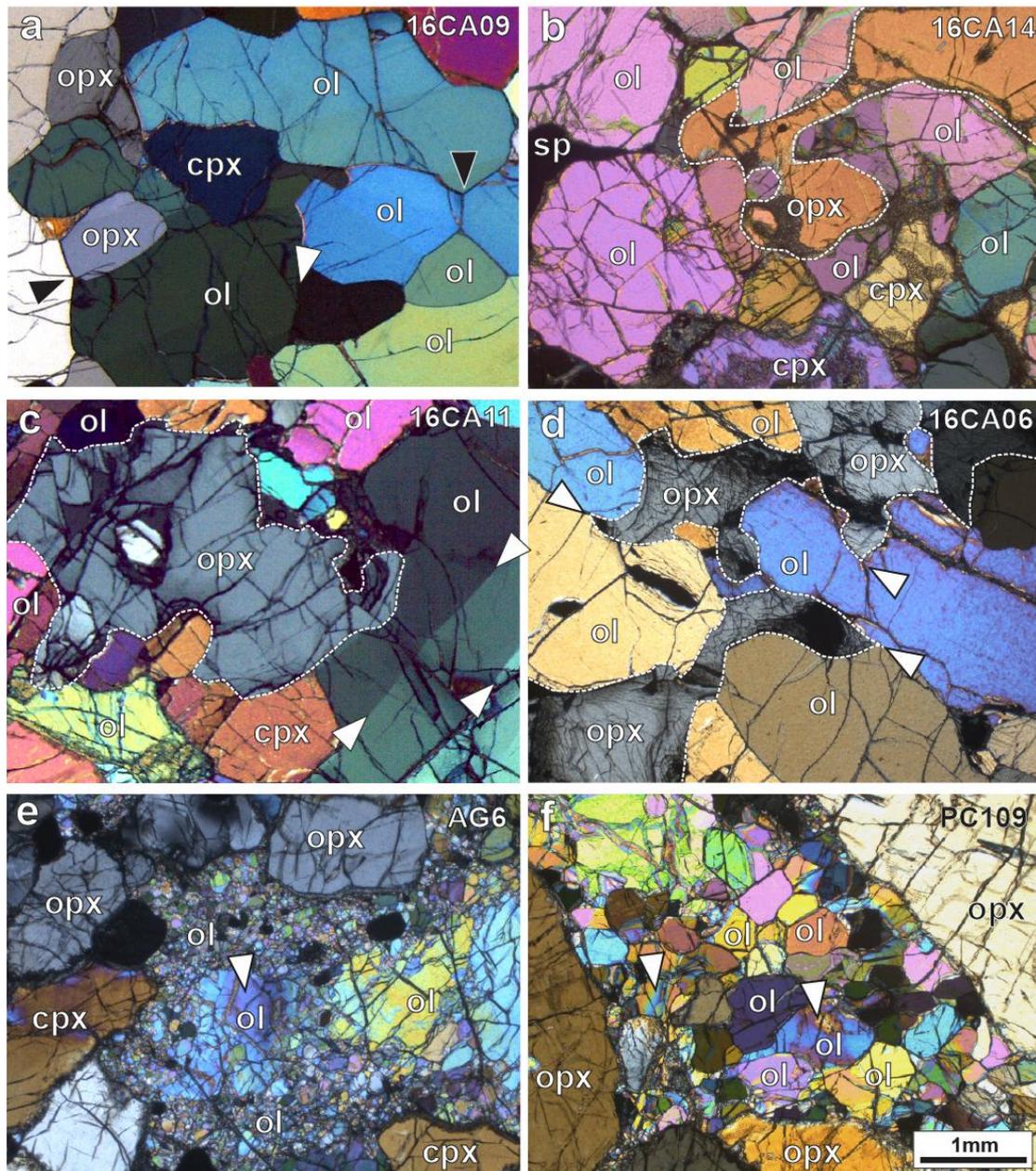
470
 471
 472 Fig. 3. Typical microstructures of Macau peridotite xenoliths: (a-b) coarse-granular, (c-d) coarse-porphyroclastic, and
 473 (e-f) fine-porphyroclastic. (a,c,e) Photomicrographs in cross-polarized light; scale bar is the same in all
 474 photomicrographs. (b,d,f,g) EBSD phase maps, which better illustrate the variation in shape of olivine and pyroxenes
 475 between the three microstructures. (g) Detail of a partially recrystallized domain in fine-porphyroclastic lherzolite AG6.
 476 (h) Misorientation relative to the mean orientation of the grain (M2M) map illustrating the difference in intragranular
 477 misorientation in olivine between porphyroclasts (high M2M) and neoblasts (low M2M) in the same domain.

478
 479 Coarse-granular peridotites are characterized by roughly equigranular
 480 microstructures (Fig. 3a-b). Olivine and orthopyroxene pyroxene grains are on
 481 average 1-3 mm, clinopyroxene slightly smaller: 0.8-1 mm (Table 2). Both
 482 olivine and pyroxenes have curvilinear to straight grain boundaries, evolving
 483 locally into polygonal shapes with 120° triple junctions (Fig. 4a), and very low
 484 densities of intracrystalline deformation features, such as undulose extinction,
 485 subgrains, or kinks (Fig. 3a and 4a). Quantitatively, these observations

486 translate into low shape factors ≤ 1.85 for olivine and ≤ 2 for orthopyroxene
487 and into low M2M values $\leq 2^\circ$ for olivine (Fig. 2). Lherzolite 16CA14 has an
488 intermediate microstructure characterized by polygonal olivine grains, but
489 pyroxenes with irregular shapes (Figs. 2 and 4b). Most coarse-granular
490 peridotites display a weak olivine shape-preferred orientation (SPO) (Fig. 3a),
491 but BO09 has higher olivine aspect ratios (Table 2) and a clear olivine SPO,
492 which results in a tabular microstructure. Pyroxene grains usually show no
493 exsolutions (Fig. 3a and 4a-b). Spinel grains have holy-leaf or interstitial
494 shapes. Coarse holy-leaf spinels may contain rounded inclusions of olivine or
495 pyroxenes (Fig. 3b).

496 Coarse-porphyroclastic microstructures are characterized by variable
497 grain sizes. In most cases, the grain size variation is continuous and it is
498 difficult to discriminate between porphyroclasts and neoblasts (Fig. 3c-d).
499 Olivine has more irregular grain shapes and clear evidence for intracrystalline
500 plastic deformation (Fig. 3c-d and 4c). Olivine grains have shape factors
501 between 1.6 and 2.2, consistently with the more sinuous grain boundaries, and
502 higher M2M values ranging from 2° to 3.4° (except 16CA15), consistent with
503 the higher frequency of undulose extinction and subgrain boundaries (Fig. 2).
504 Orthopyroxene grains also have irregular shapes (Fig. 3c-d and 4c-d), which
505 are associated with shape factors from 1.6 to 2.4 (Fig. 2b). Orthopyroxene
506 grain boundaries often show embayments or cusped shapes at the contact
507 with olivine grains (Fig. 4c-d). Clinopyroxene grains also show irregular or
508 interstitial shapes (Fig. 3c-d). In some samples, orthopyroxene grains display
509 exsolutions (Fig. 3c). Undulose extinction and kinks are observed locally.
510 Spinel usually has irregular shapes and occurs in association with ortho- or
511 clinopyroxene (Fig. 3d).

512 Fine-porphyroclastic peridotites show a well-developed bimodal olivine
513 grain size distribution, characterized by coexistence of coarse porphyroclasts
514 with serrated grain boundaries, undulose extinction, and closely-spaced
515 subgrain boundaries with a recrystallized matrix composed by polygonal
516 olivine neoblasts free of intracrystalline deformation (Figs. 3e-h and 4e-f). The
517 average size of the recrystallized grains varies between samples (Table 2): it is
518 $\sim 100 \mu\text{m}$ in AG6 (Fig. 4e), but $>300 \mu\text{m}$ in the other fine-porphyroclastic
519 peridotites (Fig. 4f, Table 2). It is noteworthy that the recrystallized olivine
520 grains are not organized in a planar mode, marking a foliation, but either form
521 irregular pockets or vein-like structures (Fig. 3e-f). Orthopyroxene is usually
522 coarser than clinopyroxene (Fig. 3f). Exsolutions were not observed in this
523 group. Shape factors of orthopyroxene vary between 1.89 and 2.56 (Fig. 2). In
524 AG6 and AG7, pyroxenes have rounded shapes (Fig. 3f), whereas in PC105
525 and PC109 fine-porphyroclastic samples, pyroxenes have irregular shapes,
526 with cusp-like terminations when in contact with olivine, similarly to pyroxenes
527 in coarse-porphyroclastic peridotites. Spinel occurs as fine rounded grains (Fig.
528 3f).



529

530

531

532

533

534

535

536

537

538

539

540

541

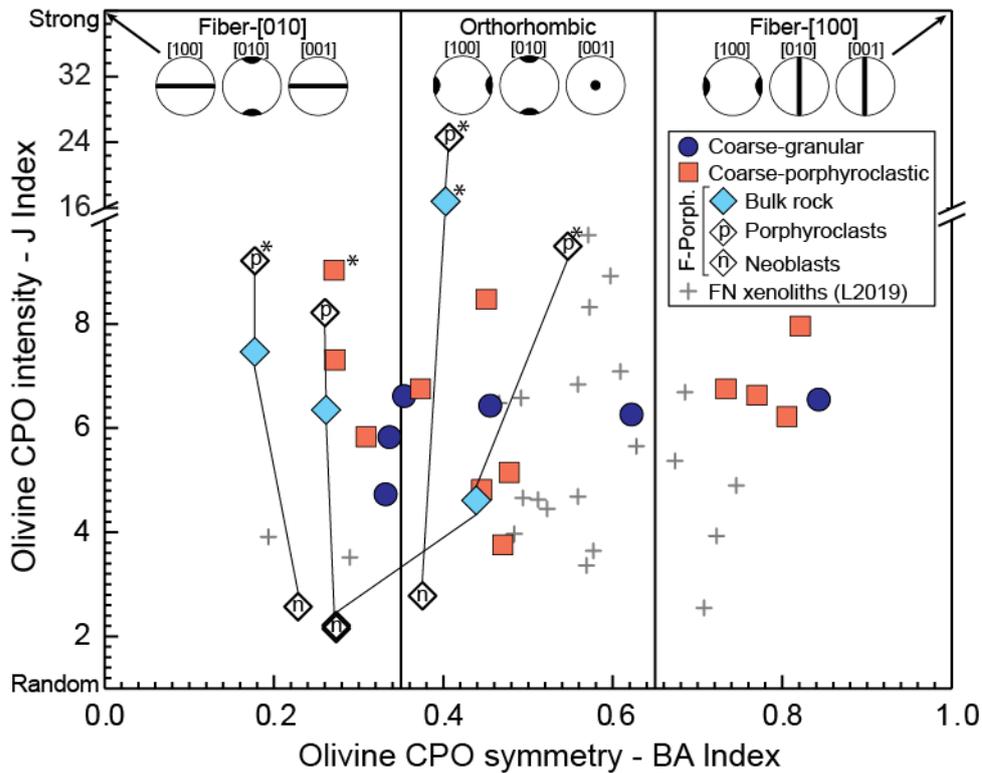
542

Fig. 4. Photomicrographs in cross-polarized light displaying typical features for the three microstructural groups. (a) Coarse granular lherzolite displaying polygonal grains (black arrow marks a 120° triple junction) almost free of intracrystalline deformation features, white arrow indicates a relict subgrain boundary in olivine (ol); (b) Coarse granular lherzolite 16CA14 displaying polygonal strain-free olivine grains, but irregularly-shaped orthopyroxene (opx) grains (highlighted by dashed line). (c-d) Coarse porphyroclastic lherzolites 16CA11 and 16CA06 showing irregularly shaped olivine, orthopyroxene (highlighted by dashed line), and clinopyroxene (cpx) grains with interpenetrating interphase boundaries, white arrows indicate subgrain boundaries in ol in (c) and mark cusp shaped opx in contact with ol in (d); (e) Fine-porphyroclastic lherzolite AG6 showing a matrix of very fine recrystallized olivine neoblasts surrounding a coarse olivine porphyroclast with undulose extinction and subgrains (white arrow); (f) Fine-porphyroclastic harzburgite PC109 showing tabular, strain-free olivine neoblasts as well as relicts of olivine porphyroclasts with undulose extinction (white arrow) within a strongly recrystallized domain between opx porphyroclasts. Scale bar is the same in all photomicrographs.

543 4.2 Crystallographic preferred orientations (CPO)

544 Quantitative data on the intensity and symmetry of the olivine CPO (J- and
 545 BA-indexes) as well as pole figures of the olivine, orthopyroxene, and
 546 clinopyroxene CPOs for all studied samples are presented in Figs. 5 and 6 and
 547 Table 2. For the samples in which a foliation was observed, the latter is
 548 displayed on the pole figures (Fig. 6). In these samples, the olivine [010]
 549 maximum is normal to the foliation.

550



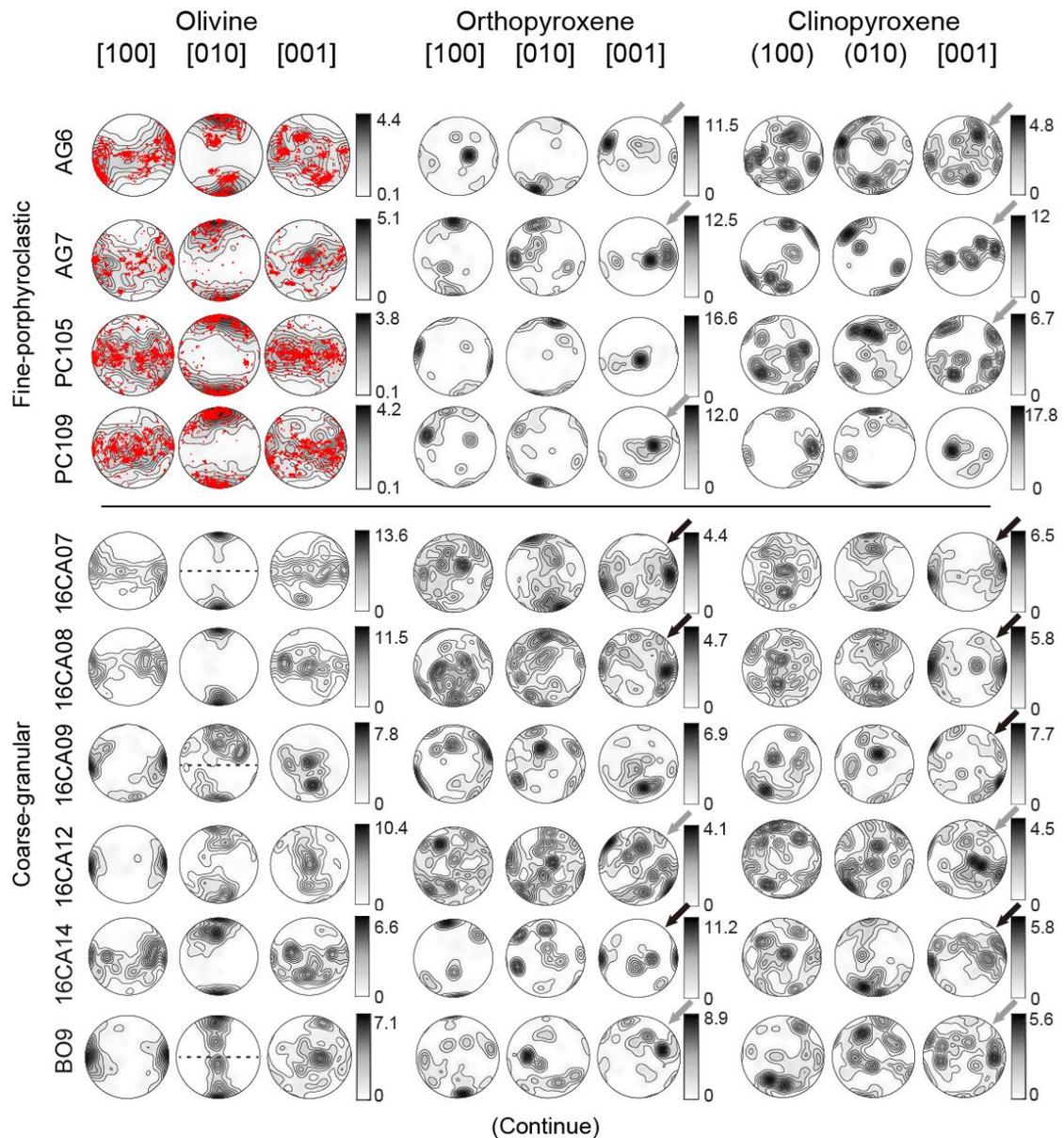
551

552 Fig. 5. Olivine CPO symmetry (BA-index) vs. CPO strength (J-index). The variation in olivine CPO pattern as a
 553 function of BA-index is illustrated at the top of the diagram. For fine-porphroclastic peridotites, we present data for the
 554 bulk rock and for porphyroclasts and neoblasts separately. Values are presented in Table 2. * indicates J-indexes that
 555 are probably overestimated, because <100 grains could be analyzed in the thin section. CPO data for Fernando de
 556 Noronha (FN) xenoliths (Liu et al., 2019) are displayed for comparison.

557

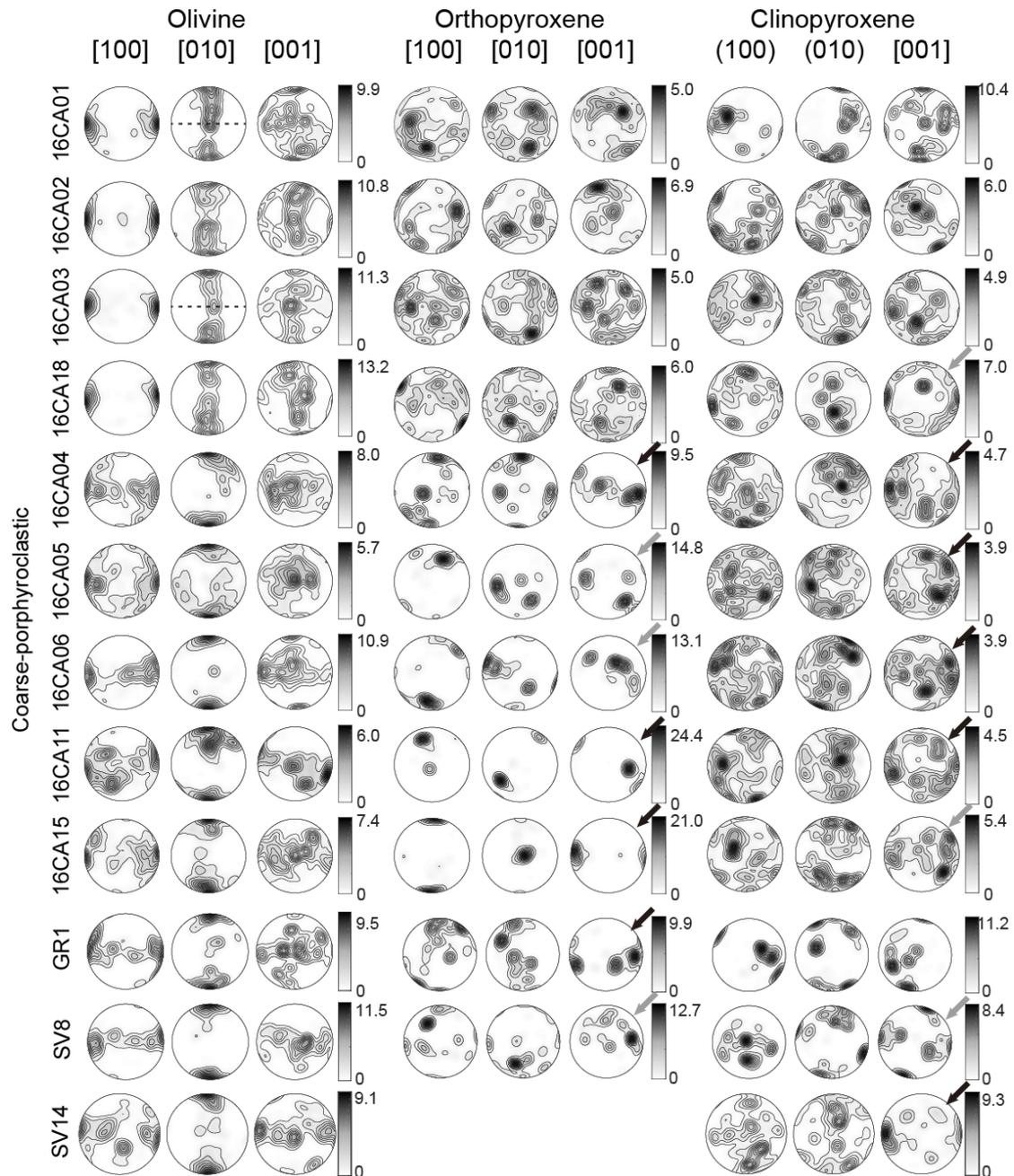
558 Olivine has moderate to strong CPO (Figs. 5 and 6). With exception of
 559 harzburgite AG7, which has a very high J-index of 16.6 due to a few coarse
 560 olivine porphyroclasts that occupy most of the thin section, the J-index varies
 561 between 3.8 and 8.5, with a mean of 6.4. The CPO symmetry varies between
 562 fiber-[010], orthorhombic, and fiber-[100] patterns. Fine-porphroclastic
 563 peridotites display olivine CPO patterns intermediate between fiber-[010] and
 564 orthorhombic and the neoblasts have systematically weaker olivine CPO (Figs.
 565 5 and 6a). There is no systematic variation of J-index or BA-index within the
 566 two other microstructural groups. Coarse-granular samples display dominantly
 567 orthorhombic olivine CPO patterns with J-indexes around 6 (Figs. 5 and 6a).
 568 Coarse-porphroclastic samples have more variable CPO patterns and

569 intensities (Fig. 5). Four coarse-porphyroclastic harzburgites (16CA01,
 570 16CA02, 16CA03, and 16CA18) have strong CPO with fiber-[100] patterns
 571 (Fig. 6b). The remainder coarse-porphyroclastic peridotites display variable
 572 CPO intensity orthorhombic patterns with a tendency towards fiber-[010],
 573 expressed as girdle of [100] with a maximum within it and a point maximum of
 574 [010] (Fig. 6b).
 575



576

577 Fig. 6. Crystal preferred orientations (CPO) of olivine, orthopyroxene, and clinopyroxene for all studied
 578 fine-porphyroclastic and coarse-granular peridotites and coarse-porphyroclastic peridotites. Lower hemisphere
 579 stereographic projections with contours at 1 multiple of a uniform distribution intervals. The three pole figures for each
 580 mineral are plotted using the same gray scale. For the fine-porphyroclastic peridotites, the CPO of olivine
 581 porphyroclasts is presented as points and the CPO of olivine neoblasts is presented as contours. Black arrows
 582 indicate pyroxene CPO that are well correlated with the olivine CPO and gray arrows indicate those partially correlated.
 583 When the sample shows a olivine shape preferred orientation marking a foliation, the latter is displayed as a dashed



585

586

587 Orthopyroxene CPO is consistent with the olivine CPO in the majority of
 588 the studied peridotites. It is characterized by $[001]_{\text{opx}}$ and $[100]_{\text{opx}}$ (or $[010]_{\text{opx}}$)
 589 maxima parallel to the $[100]_{\text{ol}}$ and $[010]_{\text{ol}}$ maxima, respectively, suggesting a
 590 common deformation history (arrows in Fig. 6). However,
 591 coarse-porphyroclastic harzburgites 16CA01, 16CA02, 16CA03, 16CA18 and
 592 coarse-granular lherzolite 16CA09 show weak orthopyroxene CPO that are not
 593 consistent with the olivine CPO, characterized by a weak $[001]_{\text{opx}}$ maximum at
 594 high angle to a well-developed $[100]_{\text{ol}}$ maximum.

595 Clinopyroxene shows more dispersed CPO, but which are, in many
 596 samples, consistent with both olivine and orthopyroxene CPO patterns,

597 characterized by $[001]_{\text{cpx}}$ and $[010]_{\text{cpx}}$ maxima subparallel to $[100]_{\text{ol}}$ and $[010]_{\text{ol}}$
598 maxima (arrows in Fig. 6). However, in most fine-porphyroclastic peridotites
599 and in coarse-granular lherzolite 16CA12, although the orthopyroxene CPO is
600 consistent with the olivine CPO, the clinopyroxene CPO is not. In
601 coarse-porphyroclastic harzburgites (16CA01, 16CA02, 16CA03, 16CA18)
602 neither the orthopyroxene nor the clinopyroxene CPO are correlated with the
603 olivine CPO.

604

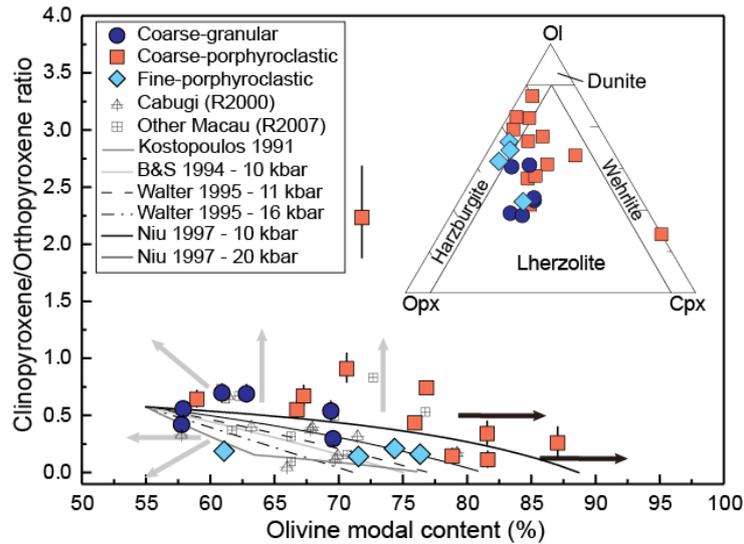
605 *4.3 Modal compositions*

606 Modal compositions were determined based on the area fraction of each
607 mineral in the EBSD maps (Fig. 7 and Table 1). All samples are spinel-facies
608 peridotites. Lherzolites predominate (14/22), but the sampling also includes 7
609 harzburgites and 1 wehrlite. Predominance of fertile peridotites with no
610 dependence on sampling site was also observed in previous sampling of
611 Macau xenoliths by Rivalenti et al. (2000, 2007; Fig. 7).

612 The compositions of the two microstructural groups overlap, but
613 coarse-granular peridotites are on average more fertile (ol contents of 57-70%)
614 than coarse-porphyroclastic peridotites (ol contents between 67-87%, with one
615 outlier – 16CA04, which has 59% of olivine and an intermediate
616 microstructure). Fine-porphyroclastic peridotites display variable olivine
617 contents overlapping with the two other groups.

618 Comparison of the modal compositions to those predicted by various
619 partial melting models highlights that roughly half of the studied Macau
620 xenoliths are enriched in clinopyroxene or olivine relatively to the models'
621 predictions (Fig. 7). Even for those samples plotting along the partial melting
622 trends, a large variability in partial melting degrees, from 1-20% melting, has to
623 be invoked to explain the full range of observed modal compositions. Such a
624 variation in partial melting is difficult to reconcile within the limited volume of
625 the mantle sampled by these volcanic eruptions. This suggests reactive melt
626 percolation leading to either crystallization of clinopyroxene and/or
627 orthopyroxene at the expenses of olivine (refertilization processes, e.g. Le
628 Roux et al., 2007; represented by gray arrows in Fig. 7) or crystallization of
629 olivine at the expense of pyroxenes (dunitization processes, e.g., Berger and
630 Vannier, 1984; Kelemen, 1990; black arrows in Fig. 7). The observed
631 deviations in modal compositions relatively to partial melting trends are
632 consistent with microstructural evidence for reactive melt percolation, such as
633 the sinuous olivine-pyroxene boundaries and the cusp-like shapes of
634 pyroxenes when in contact with two olivine grains (cf. Figs. 3 and 4).

635



636

637

Fig. 7. Modal compositions of the studied peridotites presented as clinopyroxene/orthopyroxene ratio vs. olivine modal content and as the traditional ol-opx-cpx ternary diagram (insert). Modal compositions of Cabugi and other Macau volcanics peridotite xenoliths studied by Rivalenti et al. (2000) and (2007) are plotted for comparison. Black and gray curves represent the evolution of the modal composition predicted by different partial melting models for an initial fertile modal composition of 55% ol, 28.5% opx, and 16.5% cpx (Baker and Stolper, 1994; Kostopoulos, 1991; Niu, 1997; Walter et al., 1995). Gray arrows indicate evolution trends associated with cpx and/or opx crystallization at the expense of ol (refertilization reactions). Black arrow indicates evolution trends associated with ol crystallization at the expense of pyroxenes (dunitization reactions).

645

646 4.4. Mineral compositions

647

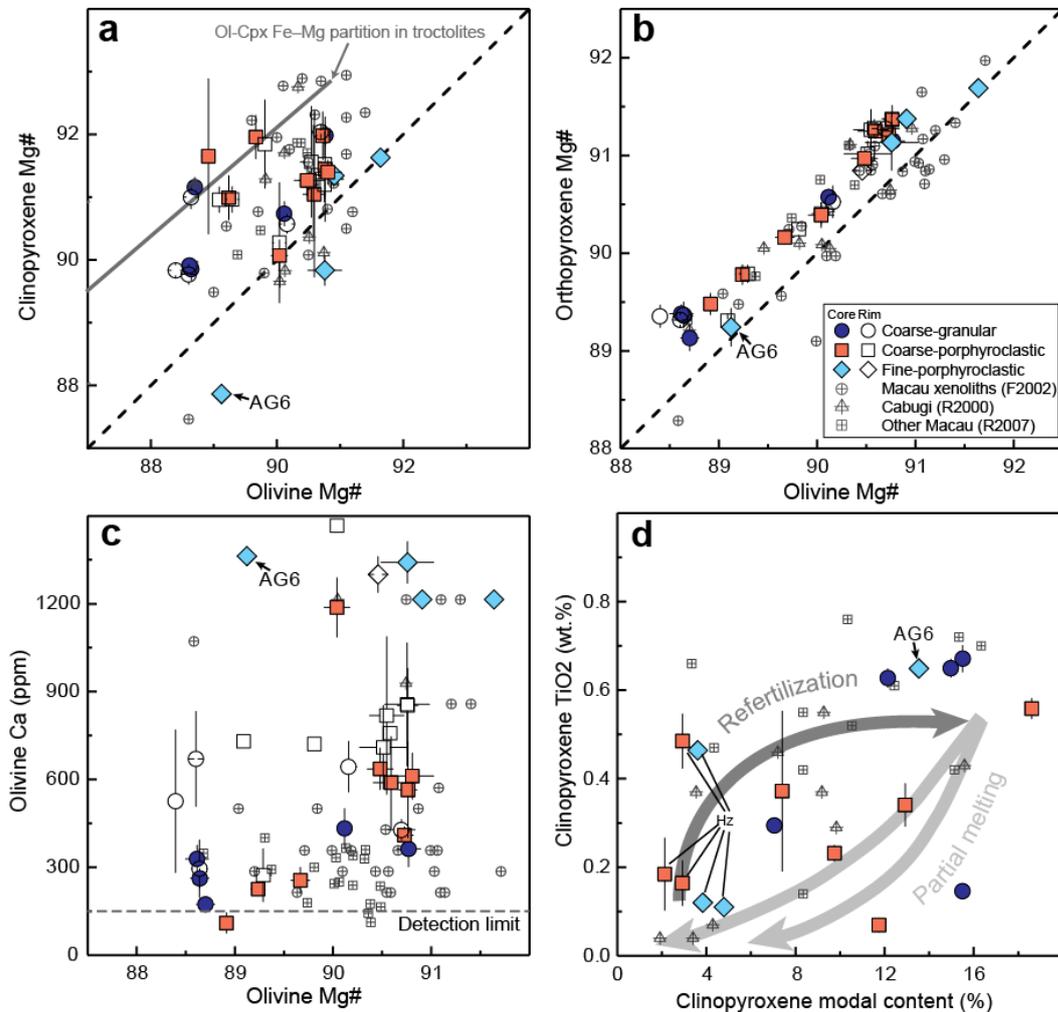
The chemical compositions of main mineral phases in the studied Macau peridotite xenoliths are listed in Supplementary Material Table S1. Representative results are displayed in Fig. 8 together with previous data for Macau peridotite xenoliths (Supplementary Material Table S2). Comparison between the present and previous datasets indicates that our sampling is representative of the variability in composition for these localities. Analysis of the present dataset indicates that there is no simple correlation between microstructural types and chemical compositions (Fig. 8). There is also no relation between mineral chemistry and sampling locality.

656

The Mg# number [$Mg\# = 100 \times Mg / (Fe + Mg)$, atomic ratio] of olivine, orthopyroxene, and clinopyroxene ranges between 89.1 – 91.6 (average 90.0), 89.1 - 91.7 (average 90.4), and 87.9 – 92 (average 90.9), respectively (Figs. 8a,b). Core-rim variations in Mg# are weak for all three minerals in most samples. Olivine and orthopyroxene Mg# within each sample show low dispersion and are positively correlated, with the orthopyroxene being slightly enriched in Mg relatively to olivine (all data points, except fine-porphyroclastic peridotites AG6 and AG7, plot slightly above the 1:1 line in Fig. 8b). Clinopyroxene Mg# shows much higher variability at the sample scale, in particular within coarse-porphyroclastic harzburgites, indicating chemical disequilibrium at the mm-scale. Moreover, average clinopyroxene Mg# in

666

667 neither cores nor rims does not display a simple correlation with the olivine
 668 Mg# (Fig. 8a). Most measured Mg# spread between the 1:1 and the Fe-Mg
 669 partition trend between olivine, clinopyroxene, and melt in oceanic troctolites
 670 (Lissenberg and Dick 2008), but fine-porphyroclastic peridotites AG6 and AG7
 671 have clinopyroxenes enriched in Fe relatively to olivine.
 672



673
 674 Fig. 8. Chemical compositions for major rock-forming minerals: (a-c) Clinopyroxene Mg#, orthopyroxene Mg#, and
 675 olivine Ca content (ppm) vs. olivine Mg#, (d) Clinopyroxene TiO₂ content (wt.%) vs. modal content (%). Minerals
 676 chemical compositions for Macau peridotite xenoliths previously studied by Rivalenti et al. (2000), (2007), and Fodor
 677 et al. (2002) are plotted for comparison. The value of the depleted mantle (DM) from Workman and Hart (2005) is
 678 displayed in (a) and (b). Fe-Mg partition between olivine and clinopyroxenes in troctolites (Lissenberg and Dick, 2008)
 679 is plotted in (a). Partial melting and refertilization trends proposed by Le Roux et al. (2007) are plotted in (d). Hollow
 680 symbols represent rim compositions and solid symbols represent core ones. Error bars represent the chemical
 681 variation at the sample scale. Data is presented in Supplementary Material Table S1.

682
 683 Olivine cores show low Ca contents with a weak positive correlation with
 684 Mg#, (Fig. 8c), except for the fine-porphyroclastic peridotites and the
 685 coarse-porphyroclastic Iherzolite GR1, which have high Ca contents. Olivine
 686 rims in all studied peridotites are enriched in Ca and show a high variability at

687 the thin section (mm-cm) scale.

688 Clinopyroxene TiO₂ contents (wt.%) do not show the simple positive
689 correlation with cpx modal content (%) expected if partial melting controlled the
690 chemistry of the peridotites (Fig. 8d). Among the harzburgites, four have
691 Ti-poor clinopyroxene grains (TiO₂ contents < 0.2%), while the other two have
692 clinopyroxene grains with TiO₂ contents of 0.55%. All harzburgites show
693 marked variability in the Ti-content of clinopyroxene at the sample scale,
694 indicating disequilibrium. Half of the lherzolites have clinopyroxenes with
695 almost constant TiO₂ contents between 0.55-0.65%, independently of the
696 clinopyroxene modal content. The remaining lherzolites have rather Ti-poor
697 clinopyroxene grains.

698 Spinel Cr# [$Cr\# = 100 \times Cr / (Cr + Al)$] and Mg# contents correlate with the
699 modal composition. Harzburgites have spinel with Cr# ranging between
700 40-57.5 and Mg# ranging between 61-72, whereas lherzolites have spinel with
701 Cr# between 8-20 and Mg# between 73-84 (Supplementary Material Fig. S1a).
702 Yet, the lherzolites show a clear enrichment in Mg in spinel relatively to a
703 typical depleted mantle composition. Fine-porphyroclastic lherzolite AG6 has
704 the highest spinel Mg# (83) and lowest Cr# (11). TiO₂ contents (wt. %) in spinel
705 are low ($\leq 0.18\%$) in most lherzolites (Supplementary Material Fig. S1b). In
706 contrast, spinel in coarse-porphyroclastic harzburgites and lherzolite GR1 has
707 a wide range of average TiO₂ contents, which may reach 1.03%, with strong
708 variations within each sample, indicating disequilibrium. Spinel in
709 fine-porphyroclastic lherzolite AG6 also has a fairly high TiO₂ content of 0.45%.

711 *4.5. Equilibrium temperatures and geotherm estimation*

712 Most equilibrium temperatures determined using two-pyroxenes
713 geothermometer of Taylor (1998; TTA98) and the Ca in orthopyroxene
714 (Ca-in-opx) of Brey and Kohler (1990) revised by Nimis and Grutter (2010)
715 agree within ± 70 °C (gray dashed lines in Fig. 9). This suggests that they are
716 reliable. However, peridotites with cpx-opx temperatures < 800 °C show larger
717 discrepancies (>90 °C), with systematically higher Ca-in-opx equilibrium
718 temperatures. This inconsistency between the predictions of the two
719 thermometers is observed for both core and rim temperatures and is not
720 correlated with the microstructure. It affects two coarse-porphyroclastic
721 peridotites (16CA06, 16CA11) and one coarse-granular lherzolite (16CA14).
722 The latter has the lowest temperature among all samples with strong
723 inconsistency between the predictions of the two thermometers (592°C and
724 790°C, Table 1).

725 Following Nimis and Grutter (2010), we use the temperature predictions
726 of TTA98 thermometer as the equilibrium temperatures. All
727 fine-porphyroclastic peridotites and the coarse-porphyroclastic peridotite GR1
728 have high equilibrium temperatures over 1200 °C. Coarse-porphyroclastic
729 peridotites have two ranges of equilibrium temperatures: harzburgites
730 (16CA01, 16CA03, 16CA18) as well as clinopyroxene-poor lherzolite SV8

731 have equilibrium temperatures between 900 and 1100 °C, whereas the
 732 clinopyroxene-rich coarse-porphyroclastic lherzolites (16CA06, 16CA11,
 733 16CA15) have lower equilibrium temperatures between 600 and 700 °C.
 734 Coarse-porphyroclastic peridotites 16CA03, 16CA15, and 16CA18 display
 735 core-rim variations in equilibrium temperature. In most cases, rim Ca-in-opx
 736 temperatures are higher than core ones, but there is no systematic trend. All
 737 samples that display core-rim variations also show significant dispersion in
 738 both core and rim temperatures at the sample scale (bars in Fig. 10).
 739 Coarse-granular peridotites 16CA07, 16CA08, 16CA09, and BO9 have
 740 intermediate equilibrium temperatures between 800 and 900 °C and less
 741 variations at sample scale as well as between cores and rims. Comparison
 742 with previous thermometry data on Macau peridotite xenoliths indicates that
 743 the present sampling is representative of the variability in this suite (Fig.9).
 744 Analysis of the full dataset also highlights that there is no relation between
 745 equilibrium temperatures and sampling site. The full range of equilibrium
 746 temperatures is observed both among Pico do Cabugi xenoliths, which is the
 747 best sampled site, and among xenoliths from other Macau volcanics.
 748

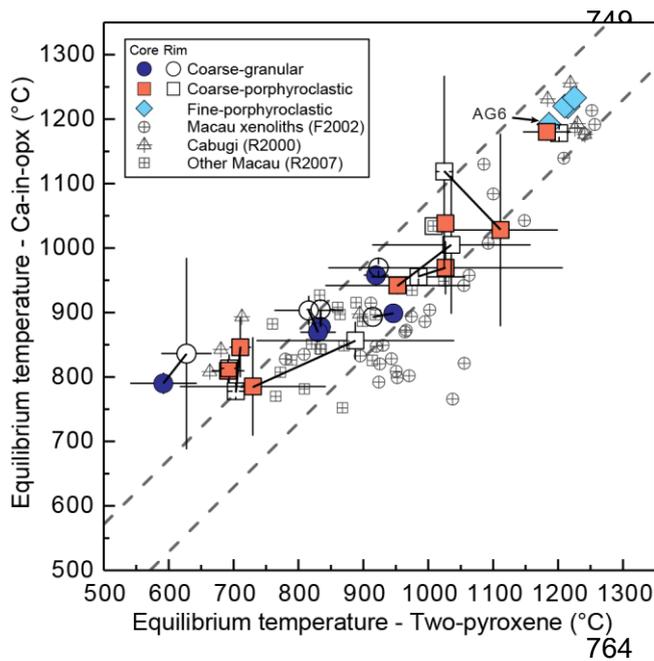


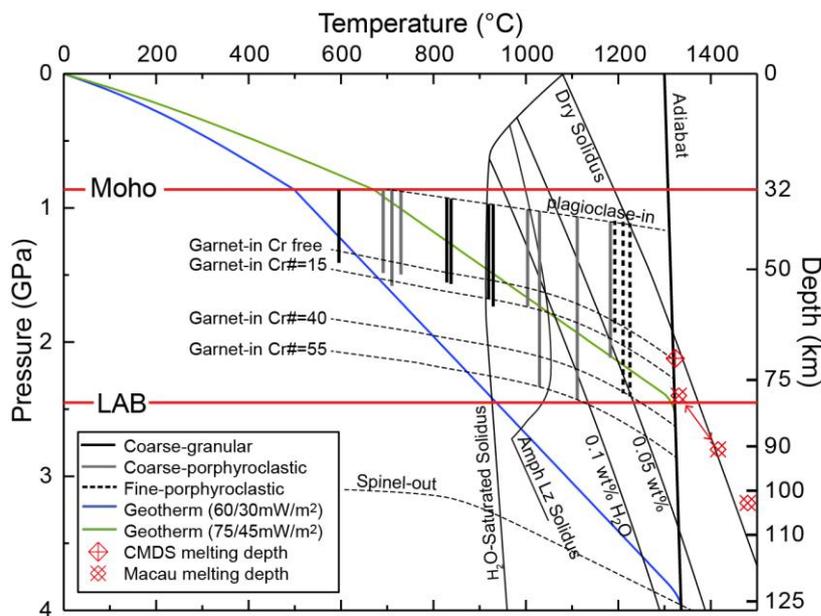
Fig. 9. Average core and rim equilibrium temperatures calculated using the cpx-opx thermometer (TTA98, Taylor, 1998) and the Ca-in-opx thermometer (Ca-in-opx, Brey and Kohler, 1990, revised by Nimis and Grutter, 2010). Chemical compositions from Rivalenti et al. (2000) and (2007) and Fodor et al. (2002) were used to recalculate the equilibrium temperatures for their samples using the same thermometers (data is presented in Supplementary Material Tables S2 and S3). Error bars represent data spread within each sample. Gray dashed lines outline a $\pm 70^\circ\text{C}$ variation between the two thermometers.

765 To estimate the Cenozoic geotherm in the northern Borborema Province,
 766 we plotted the TTA98 equilibrium temperatures of the studied xenoliths against
 767 the peridotite phase diagram in the upper mantle (spinel-out, plagioclase-out,
 768 and garnet-in curves for different Cr contents from O'Neill, 1981 and Klemme,
 769 2004), as well as equilibrium geotherms for different surface and reduced heat
 770 flows (Fig. 11). The geotherms are calculated assuming a surface heat flow of
 771 60 or 70 mW/m² (Hamza et al., 2018), a Moho depth of 32 km (Almeida et al.,
 772 2015), an exponential decrease of the radiogenic heat production with depth,
 773 with a characteristic depth of 10 km, and a reduced heat flow of either 30 or 45
 774 mW/m². For comparison, we also indicate the depth of the

775 lithosphere-asthenosphere boundary determined by P-S receiver functions for
 776 seismic station RCBR (Heit et al., 2007) and the partial melting conditions
 777 estimated based on the compositions of the most primitive melts of the
 778 Cenozoic Macau and Mesozoic Ceará-Mirim suites (Ngonge et al, 2015a,b).

779 P-T equilibrium conditions of the peridotites are displayed as a fixed
 780 temperature corresponding to the TTA98 thermometry prediction, but as a
 781 pressure range, because there are no reliable barometers for spinel-facies
 782 peridotites. The range of equilibrium pressure conditions is defined by the
 783 occurrence of spinel in absence of plagioclase or garnet for the actual Cr
 784 content of each peridotite (O'Neill, 1981; Klemme, 2004).

785 The wide range of equilibrium temperatures suggests that the studied
 786 peridotites represent a complete sampling of the subcontinental lithospheric
 787 mantle. The absence of garnet in the entire suite, which is largely composed of
 788 peridotites with Cr# < 15, constrains a maximum equilibrium depth of the
 789 studied peridotites shallower than 75 km (Fig. 10). By consequence, the
 790 equilibrium conditions of the Macau peridotite xenoliths, except 16CA14 that
 791 has the lowest and likely unreliable equilibrium temperature, are better fitted by
 792 the hotter geotherm (green line in Figure 10).



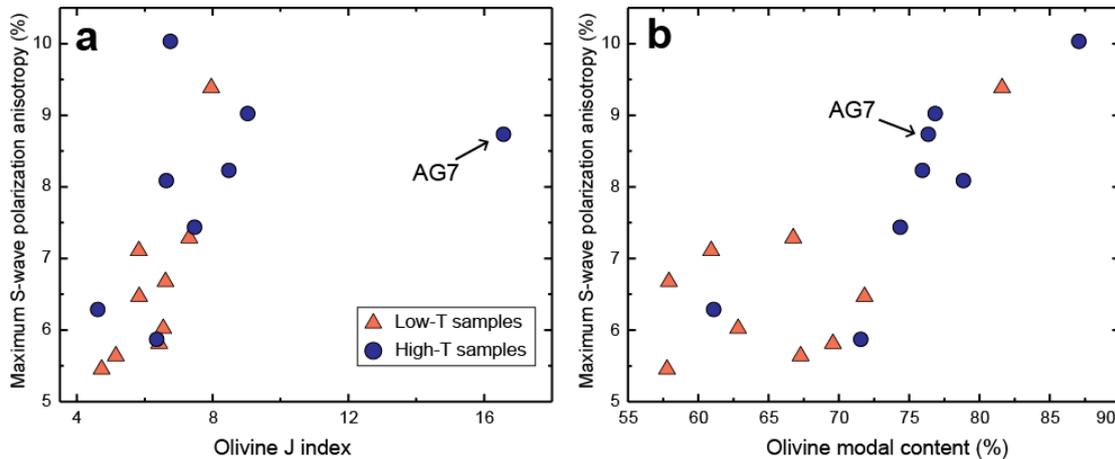
793
 794 Fig. 10. P-T diagram comparing the equilibration conditions estimated for the studied Macau xenoliths to two possible
 795 steady-state geotherms for the north Borborema province (surface / reduced heat flows indicated in parenthesis).
 796 Xenolith data is better fitted by the hotter geotherm (surface and reduced heat flows of 75 and 45mW/m², respectively).
 797 Since no barometers are available for spinel peridotites, equilibrium pressure ranges are presented as bars, based on
 798 the absence of plagioclase and of garnet in the studied peridotites, considering their Cr# (Supplementary Information
 799 Table S1) and the effect of Cr on the spinel-garnet phase transition (Klemme, 2004; O'Neill, 1981). Dry and wet solidus
 800 with variable H₂O contents after Ringwood (1975). Moho and Lithosphere-Asthenosphere Boundary (LAB) depths
 801 derived from S-P and P-S receiver function analyses (Almeida et al., 2015, Heit et al., 2007) as well as partial melting
 802 conditions derived from the primitive melt compositions of the Cretaceous Ceará-Mirim dyke system (CMDS) and
 803 Cenozoic Macau volcanics (Ngonge et al., 2016a,b) are displayed for comparison.

804 **4.6. Seismic properties**

805 For calculating seismic properties, we divided the xenoliths into two
 806 groups as a function of their TTA98 equilibrium temperature: low ($T < 1000\text{ }^{\circ}\text{C}$)
 807 and high-temperature ($T > 1000\text{ }^{\circ}\text{C}$). Then, we estimated the elastic tensors of
 808 the individual samples (Supplementary Material Table S3) for the pressure and
 809 temperature conditions predicted at 50 km (1.46 GPa and $916\text{ }^{\circ}\text{C}$) and 70 km
 810 depth (2.12 GPa and $1189\text{ }^{\circ}\text{C}$) for the geotherm that best fits the xenolith data
 811 ($Q_s=70\text{ mW/m}^2$ and $Q_m=45\text{ mW/m}^2$, green line in Figure 10).

812 Seismic anisotropy patterns of the individual samples vary slightly as a
 813 function of the olivine CPO symmetry. As usual, the seismic anisotropy
 814 intensity correlates positively with both the olivine CPO strength and olivine
 815 modal content (Figure 11). On average, the peridotites equilibrated at
 816 temperatures $<1000\text{ }^{\circ}\text{C}$ tend to display lower anisotropy, due to both more
 817 dispersed olivine CPO and lower olivine contents. However, seismic
 818 anisotropy intensity does not increase linearly with increasing J-index. By
 819 consequence, overestimation of the olivine CPO intensity for
 820 fine-porphroclastic harzburgite AG7 does not result in similar overestimation
 821 of its seismic anisotropy. The maximum S-wave polarization anisotropy varies
 822 from 5.5 to 7.3% in the low-temperature peridotites and from 5.9 to 10% in the
 823 high-temperature ones (Fig. 11 and Table 3). Similarly, the maximum P-wave
 824 propagation anisotropy varies from 7.4 to 10% and from 7.7 to 14.2% (Table 3).
 825 The maximum fast (S1) wave propagation anisotropy varies from 3.0 to 5.9%
 826 in low-temperature samples and from 4.1 to 7.2% in high-temperature ones,
 827 whereas the maximum slow (S2) wave propagation anisotropy varies from 3.4
 828 to 5.2% and from 1.9 to 7.3%, respectively (Table 3).

829



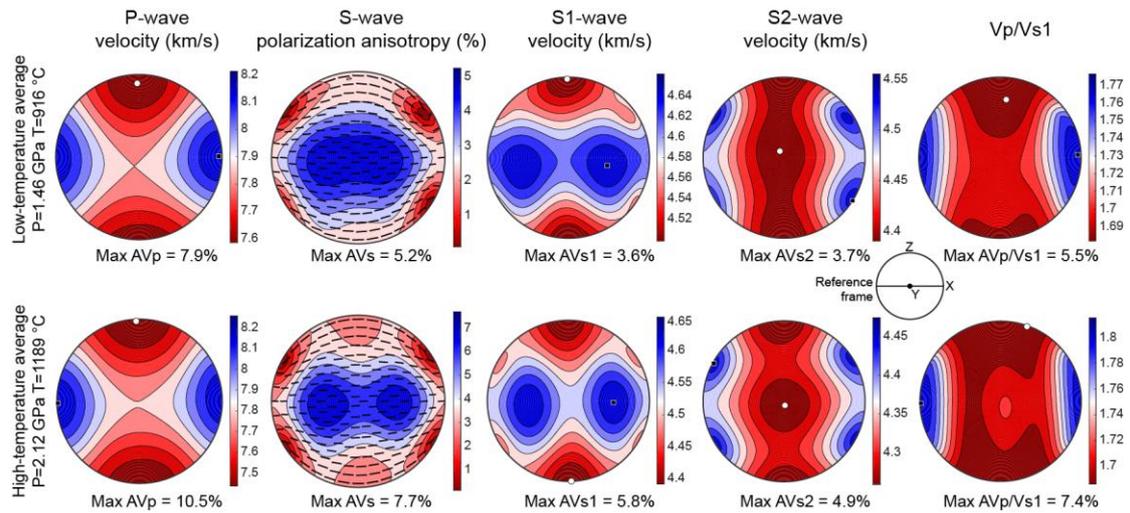
830

831 Fig. 11. Maximum S-wave polarization anisotropy (%) vs. (a) olivine CPO intensity (J index) and (b) olivine modal
 832 content (%) for all studied xenoliths. Seismic anisotropy data for individual samples are listed in Table 3.

833

834 Since seismic waves average elastic properties over large volumes, we
 835 calculated average elastic tensors for the low-temperature and
 836 high-temperature groups by assuming a coherent orientation of the past flow
 837 directions and planes (assumed as parallel to the [100] and [010] maxima of
 838 the olivine CPO, respectively – this assumption is justified by the analysis of

839 the CPO as discussed in the next section) over the entire lithosphere using a
 840 constant structural reference frame. These average seismic anisotropy
 841 patterns (Fig. 12) represent the maximum possible seismic anisotropy in the
 842 shallow (low-temperature group) and deep (high-temperature samples)
 843 sections of the lithospheric mantle, if the present sampling is representative of
 844 the volumes of the different microstructures and compositions in the
 845 lithospheric mantle in this region.
 846



847

848 Fig. 12. Average seismic properties for shallow and deep sections of the lithospheric mantle beneath the study area
 849 calculated by averaging the elastic constant tensors of the eight low-temperature samples calculated for 916°C,
 850 1.46GPa and of the nine high-temperature samples calculated for 1189°C, 2.12GPa. Lower hemisphere stereographic
 851 projections presenting the variation of the property as a function of the propagation direction relatively to the structural
 852 reference frame (flow direction (X) and normal to the flow plane (Z) shown by the insert). Black squares mark the
 853 highest values and white circles mark the lowest ones. Seismic properties data are listed in Table 3 and the elastic
 854 tensors, in Supplementary Material Table S4.

855

856 Seismic anisotropy patterns for the shallow and deep lithospheric mantle
 857 are similar, but as already discussed, the lower part of the lithosphere is more
 858 anisotropic (Fig. 12). The average maximum P-wave propagation anisotropy is
 859 10.5% with velocities ranging between 7.4 and 8.3 km/s, compared to 7.9% in
 860 the shallow lithospheric mantle, where P-wave velocities vary between 7.6
 861 km/s normal to the flow plane and 8.2 km/s parallel to the flow direction frozen
 862 in the lithospheric mantle (Table 3). The average maximum S-wave
 863 polarization anisotropy in the lower lithospheric mantle is 7.7% compared to
 864 5.2% in the shallower levels; in both cases it is observed for S-waves
 865 propagating within the flow plane but at $<45^\circ$ of the flow direction. Low S-wave
 866 polarization anisotropy is observed for all waves propagating at low angle to
 867 the XZ plane, that is, the plane that contains both the flow direction and the
 868 normal to the flow plane. For all propagation directions in which a significant
 869 polarization anisotropy is observed, the fast S-wave is polarized in the plane
 870 containing the flow direction. The average maximum S1-wave propagation

871 anisotropy increases from 3.6% to 5.8% from the shallow to the deep mantle
872 lithosphere. S1 velocity is maximum within the flow plane and minimum normal
873 to it. S1-waves propagating within this plane show a 90° periodicity in the
874 velocity variation. S2-waves average maximum propagation anisotropy
875 increases from 3.7% to 4.9% with depth. High S2 velocities are associated
876 with propagation within the XZ plane at ca. 45° to the flow direction (X) and low
877 velocities are observed for all propagation directions at high angle to X. The
878 maximum Vp/Vs1 ratio anisotropy increases from 5.5% to 7.4% with depth.
879 Highest Vp/Vs1 ratios (1.78-1.8) are observed for waves propagating parallel
880 to the flow direction and the lowest Vp/Vs1 ratios for waves propagating
881 normal to the flow plane.

882

883 **5. Discussion**

884

885 *5.1. Deformation, annealing, and reactive melt percolation*

886 Coarse-granular peridotites show polygonal grain shapes for both olivine
887 and pyroxenes and low intragranular misorientations (olivine M2M values < 2°,
888 Fig. 2). However, they have well-developed olivine CPO (Figs. 5 and 6). This
889 association supports that the deformation that produced the CPO was followed
890 by annealing (static recrystallization), which effectively reduced dislocation
891 densities and re-equilibrated grain shapes. The analysis of the relative
892 intensity of the orientation of [100], [010], and [001] of olivine, of the relations
893 between olivine and pyroxene CPOs, and of the relations between olivine CPO
894 and SPO (when a SPO was observed) points to deformation by dislocation
895 creep with dominant activation of the [100](010) slip system in olivine,
896 [001](100) in orthopyroxene, and [001](0kl) in clinopyroxene (Tommasi et al.
897 2000, Bascou et al. 2002). Coarse-granular peridotites have roughly constant
898 olivine CPO intensities, but variable symmetry, covering the entire range from
899 fiber-[100] to fiber-[010]. Fiber-[100] and orthorhombic olivine CPO are the
900 usual patterns produced by simple shear under dry, high to moderate
901 temperatures, and low pressure in the upper mantle (Tommasi et al., 1999;
902 2000; Hansen et al. 2014). Fiber-[010] patterns, on the other hand, require
903 particular conditions, which may be: transpression (Tommasi et al., 1999),
904 presence of melts during the deformation (Higgie and Tommasi, 2012, 2014;
905 Qi et al. 2018), changes in the olivine CPO by oriented growth during
906 recrystallization (e.g., Tommasi et al., 2008), or deformation under high stress
907 or high pressure conditions, which would lead to increased activation of [001]
908 glide (Durham and Goetze, 1977; Mainprice et al., 2005; Demouchy et al.
909 2013). The mineralogy and microstructures and CPO of the coarse-grained
910 peridotites falsify the last hypothesis. However, the present data does not allow
911 for discriminating among the other hypotheses. The different olivine CPO
912 symmetry may therefore result from variations in deformation regime, with
913 fiber-[010] CPO recording transpression (which is the dominant deformation
914 regime in the Seridó belt during the Brasiliano event), presence or not of melts

915 during the deformation (all deformation events were accompanied by
916 magmatism), or different CPO evolution during recrystallization.

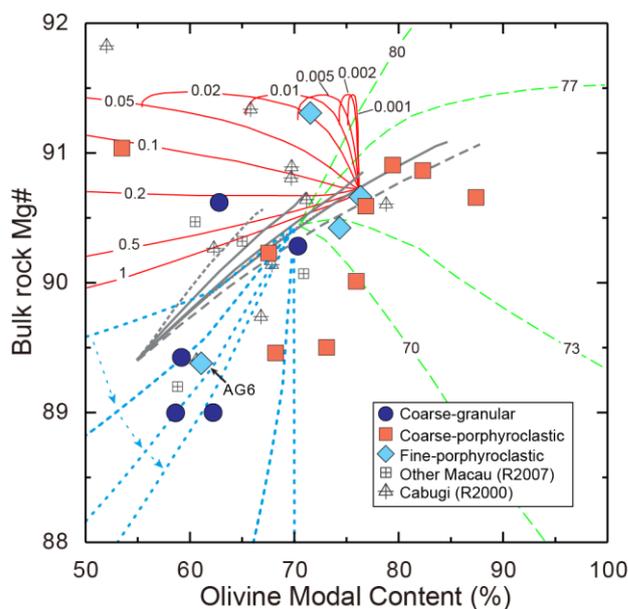
917 Analysis of the bulk rock Mg# vs. olivine modal content relation (Fig. 13)
918 shows that the compositions of most coarse-granular peridotites deviate from
919 partial melting trends, suggesting the occurrence of refertilization processes.
920 Yet, most coarse-granular peridotites have coherent olivine and pyroxene
921 CPOs (Fig. 6), which indicate co-deformation of the two phases. Thus, if
922 melt-rock reactions leading to refertilization occurred, they predated or were
923 concomitant to the deformation. Refertilization reactions produce sutured
924 pyroxene-olivine boundaries and irregular shapes for both minerals. The
925 equilibrated pyroxenes grain shapes in most coarse-granular peridotites
926 indicate therefore that reactive melt percolation producing refertilization also
927 predated annealing. It also implies that interphase grain boundary
928 rearrangements during annealing were effective. These rearrangements
929 depend on transport of ions along grain boundaries in a similar way to the
930 growth of porphyroblasts in a metamorphic rock, but with weaker driving forces
931 (grain boundary energy reduction opposed to chemical gradients). Most
932 coarse-granular peridotites equilibrated around 900°C (Fig. 9). At this
933 temperature, given the diffusivity of Si along olivine and enstatite grain
934 boundaries ($\leq 10^{-27} \text{ m}^3/\text{s}$; Fisler et al. 1997; Fei et al. 2016), grain boundary
935 equilibration at the 250 μm scale, which is the average amplitude of the
936 sinuosity of the olivine-pyroxene grain boundaries in the
937 coarse-porphyroclastic peridotites (Fig. 3c,d), will occur on time scale of
938 several hundreds of Ma. In contrast, irregular pyroxene grain shapes (high
939 orthopyroxene shape factor $\sim 2^\circ$, Fig. 2) in 16CA14 and BO9 point to a later
940 stage of melt percolation. Indeed, in both samples, part of the pyroxenes
941 shows crystal orientations not coherent with the olivine CPO (Fig. 6).

942 Coarse-porphyroclastic peridotites show less equilibrated
943 microstructures. All major phases show sinuous grain boundaries (higher
944 shape factors) as well as undulose extinction and subgrains, which translate
945 into higher M2M values (Fig. 2). This implies a less effective annealing due to
946 either a more recent deformation or lower post-deformation temperatures. The
947 former hypothesis may apply for the coarse-porphyroclastic peridotites with
948 high equilibrium temperatures ($\geq 1000^\circ\text{C}$, Fig. 9) and the latter, for those with
949 low equilibrium temperatures ($< 800^\circ\text{C}$, Fig. 9). Similarly to the coarse-granular
950 peridotites, olivine CPO in coarse-porphyroclastic peridotites is consistent with
951 deformation by dislocation creep with dominant activation of the [100](010)
952 system. This interpretation is corroborated by the high frequency of (100)
953 subgrain boundaries. However, the CPO patterns and intensities are more
954 varied than those of coarse-granular peridotites (Figs. 5 and 6).
955 Coarse-porphyroclastic harzburgites have strong fiber-[100] olivine CPOs,
956 typical of simple shear deformation (Tommasi et al., 1999; 2000; Bystricky et al,
957 2000; Hansen et al., 2014). Lherzolites have orthorhombic to fiber-[010] olivine
958 CPOs with variable strength. As for the coarse-granular peridotites, the

959 fiber-[010] olivine in these coarse-porphyroclastic peridotites may record either
960 a component of transpression or the presence of melts during the deformation.

961 Although some coarse-porphyroclastic peridotites plot along partial
962 melting trends in Fig. 13, evidence for reactive melt percolation leading to
963 crystallization of pyroxenes or olivine is widespread. It encompasses: (i) the
964 interpenetrating olivine-pyroxene grain boundaries and the locally interstitial
965 shapes of pyroxenes (Figs. 3 and 4), which imply lack of microstructural
966 equilibrium, (ii) the high variability of the olivine and pyroxenes chemical
967 compositions both within grains and between grains in a sample (Fig. 8), which
968 indicates absence of chemical equilibration, and (iii) lack or weak consistency
969 between the olivine and the pyroxenes CPOs (Fig. 6). The latter feature is
970 specific to the coarse-porphyroclastic harzburgites, which also show the
971 highest equilibrium temperatures and least equilibrated mineral compositions
972 among the coarse-porphyroclastic peridotites, implying that reactive melt
973 percolation in these rocks, which sample the lower lithospheric mantle section
974 in the province, postdates the deformation and is rather recent.

975



976

977 Fig. 13. Olivine modal content (%) vs. bulk rock Mg# in the studied xenoliths compared to evolutions predicted for
978 partial melting and reactive melt percolation in the mantle. Gray lines represent the composition evolution predicted by
979 various partial melting models using a source composition with 89.3 Mg# and 55% of olivine up to complete
980 consumption of clinopyroxene (Bodinier & Godard, 2014). Colored lines represent different melt-rock reactions after
981 Bodinier & Godard (2014). Red solid lines correspond to precipitation of clino- and orthopyroxene at the expense of
982 olivine and melt with different mass ratio of crystallized minerals versus infiltrated melt (numbers of the top of the
983 curves). Green dashed lines show olivine-forming reactions with melts with different Mg# (numbers of the top of the
984 curves). Blue dotted lines represent multiple episodes of refertilization, starting with low Mg# melts ('primitive' melt
985 Mg#=74.5), in which the peridotites successively reacted with the evolved melt resulting from previous infiltration
986 stage (Bodinier et al., 2008). Compositions of Macau peridotite xenoliths previously studied by Rivalenti et al. (2000)
987 and (2007) are plotted for comparison.

988

989 Fine-porphyroclastic peridotites show a bimodal olivine grain size
990 distribution and strong intragranular misorientations in the olivine
991 porphyroclasts characterized by dynamic recrystallization (Figs. 3 and 4). At
992 the high equilibrium temperatures recorded by these peridotites ($\geq 1200^\circ\text{C}$, Fig.
993 9), diffusion is fast. The low annealing level of the microstructures in these
994 peridotites, indicated by the high intragranular misorientations in the
995 porphyroclasts (Figs. 2 and 3), implies therefore that the deformation episode
996 that produced the recrystallization shortly predated their extraction from the
997 mantle by the Macau volcanism.

998 The variation in recrystallized grain sizes between the various
999 fine-porphyroclastic peridotites might record variations in stress (from ~ 75
1000 MPa in AG6 to ~ 10 MPa using the recrystallized grain size paleopiezometer of
1001 Van der Wal et al., 1993), but the coarser recrystallized grain sizes result more
1002 probably from partial annealing. The stresses estimated for lherzolite AG6 are
1003 high and, for the equilibrium temperature of 1200°C of these peridotites, imply
1004 extremely high strain rates of 10^{-6} to 10^{-9} s^{-1} based on usual olivine flow laws
1005 (Chopra and Paterson, 1981; Hirth and Kohlstedt, 2003; Gouriet et al. 2019).
1006 The microstructure of these peridotites is indeed very similar to the mosaic
1007 microstructure of deep sheared kimberlite-borne peridotites, which has been
1008 traditionally attributed to the initial stages of kimberlite dyke formation, due to
1009 the high stresses and high rates inferred based on the recrystallized grain
1010 sizes and equilibrium temperatures (e.g., Green and Gueguen, 1974; Boullier,
1011 1977; Skemer and Karato, 2008; Baptiste et al. 2012). Similar microstructures
1012 have been observed in the deepest mantle xenoliths from the Labait alkaline
1013 lavas in the Tanzanian craton (Vauchez et al., 2005) and from Malaita alnoites
1014 in the Ontong Java plateau (Tommasi and Ishikawa, 2014) with similar
1015 interpretations proposed.

1016 Fine-porphyroclastic peridotites with high equilibration temperatures have
1017 also been described in Cenozoic peridotite xenoliths, which sample the mantle
1018 beneath major Neoproterozoic shear zones at the border of the Hoggar swell,
1019 N Africa (Kourim et al., 2015). However, in the Hoggar peridotites, equilibration
1020 temperatures are lower ($1000\text{-}1100^\circ\text{C}$) and olivine recrystallization was
1021 associated with crystallization of elongated aggregates of clinopyroxene and
1022 amphibole. The fine-porphyroclastic microstructures were therefore interpreted
1023 as resulting from ductile reactivation and melt channeling in Neoproterozoic
1024 shear zones in response to the development of the Hoggar swell in the
1025 Cenozoic. However, in the fine-porphyroclastic peridotites from the Borborema
1026 province, there is no evidence for neocrystallization of pyroxenes or
1027 amphiboles within the recrystallized domains. Moreover, the fact that the
1028 recrystallized domains in the fine-porphyroclastic peridotites from the
1029 Borborema province do not align marking a foliation (Fig. 3e,f) and the lack of
1030 elongation of the pyroxenes suggests that the recrystallization was associated
1031 with high stresses, but low finite strains. This association, together with the
1032 equilibrium temperatures, which imply that these peridotites are derived from

1033 the base of the lithospheric mantle, is consistent with localized deformation
1034 associated with the formation of the dykes that fed the Cenozoic magmatism.

1035 Analysis of the olivine modal composition relative to the bulk rock Mg#
1036 implies that Iherzolites AG7 and AG6 were affected by refertilization processes
1037 (Fig. 13). The Fe-rich compositions of olivine and pyroxenes in AG6 (Fig. 8a,b)
1038 further point to high cumulated melt-rock ratios. Ortho- and clinopyroxenes in
1039 these two Iherzolites have unusual irregular, but rounded shapes, which clearly
1040 differ from those in coarse-porphyroclastic peridotites (cf. EBSD phase maps
1041 in Fig. 3). Yet determining when this refertilization occurred is difficult. At the
1042 high temperatures at which these peridotites equilibrated, chemical diffusion is
1043 fast. Disequilibrium in mineral chemistry at the sample scale, which would
1044 point to melt-rock interaction shortly before extraction is only observed in AG7.
1045 Harzburgites PC105 and PC109 plot along the partial melting trends in Fig. 13
1046 and have higher Mg# in olivine and pyroxenes (Fig. 8a,b), but they also display
1047 chemical evidence for some melt-rock interaction, like enrichment in Ca in
1048 olivine (Fig. 8c).

1049

1050 *5.2. Cenozoic geotherm and thermal evolution of the NBP lithospheric mantle*

1051 We do have evidence in this study supporting a rather hot Cenozoic
1052 geotherm beneath the North Borborema Province. As illustrated in Fig. 10,
1053 equilibrium conditions of the xenoliths are consistent with the surface heat flow
1054 of 60-70 mW/m² measured in the Borborema Province (Hamza et al., 2018) if
1055 the heat flow from the convective mantle is rather high (45mW/m²). This
1056 implies a slightly hotter than average sublithospheric mantle beneath the
1057 Borborema Province, consistently with the low P-wave velocity anomaly
1058 imaged at 100 km depth beneath the Northern Borborema Province east of the
1059 Macau-Queimadas volcanic alignment (Simões Neto et al., 2019; Fig. 1) and
1060 with the weak low S-wave velocity anomaly imaged beneath this region in a
1061 recent global full-waveform tomography model (ca. -2%; French et al., 2013).
1062 The equilibrium geotherm that best fits the equilibrium temperature and
1063 pressure conditions of the Borborema Province is also consistent with the
1064 seismological constraints for the lithosphere-asthenosphere boundary (LAB)
1065 depth of 80km (Heit et al., 2007) and with the partial melting conditions
1066 calculated for the Macau most primitive basalts (1330-1415°C at 80-93km;
1067 Ngonge et al. 2015b; cf. Fig. 10). A hotter than average sublithospheric mantle
1068 may also account for the Cenozoic uplift of the Borborema Plateau (Almeida et
1069 al., 2015; Luz et al., 2015; Klöcking et al., 2018). The equilibrium temperatures
1070 of the studied xenoliths may therefore represent an equilibrium geotherm
1071 established in the Cenozoic in response to a slightly hotter than normal
1072 convective mantle temperature. Geophysical data imply that these conditions
1073 are still active today. Data on the xenoliths does not bring any constraints on
1074 the causes of the higher than average sublithospheric temperatures, which
1075 may result from a diffuse mantle upwelling, perturbation of the convective
1076 pattern by the São Francisco craton, or, as suggested by Simoes Neto et al.

1077 (2018), lateral channeling of hot material from a mantle plume upwelling to the
1078 SW of the Province.

1079 The present data also do not constrain the evolution through time of the
1080 upper mantle temperatures beneath the North Borborema Province. The
1081 estimated melting conditions for the most primitive basalts of the Cretaceous
1082 CMDS (ca. 1320 °C at 70 km depth; Ngonge et al., 2015a) are shallower than
1083 those for the Cenozoic Macau volcanics (Ngonge et al., 2015b, suggesting an
1084 even shallower LAB beneath the North Borborema Province in the Mesozoic.
1085 This suggests that the lithosphere beneath the NBP has probably cooled and
1086 thickened after the Mesozoic extension. The deepest xenoliths might therefore
1087 represent material accreted to the base of the lithosphere after the Mesozoic.
1088

1089 *5.3. Seismic anisotropy in the lithospheric mantle: comparison with SKS* 1090 *splitting data*

1091 Shear wave splitting data in the Borborema Province is highly
1092 heterogeneous and does not relate in a simple way to neither the outcropping
1093 geological structures nor the absolute plate motion of the South American
1094 plate (Bastow et al., 2015). However, the station RCBR, which is the closest to
1095 sampling sites, being located ca. 50 km east of the Pico do Cabugi, displays a
1096 NNE-oriented fast S-wave polarization parallel to the trend of the Seridó belt
1097 and of the main Brasiliano shear zones in the region, and a high delay time
1098 (1.9 s) based on 9 individual measurements (Assumpção et al., 2011).

1099 If we consider that at least part of the SKS splitting measured at RCBR is
1100 produced in the lithospheric mantle, the orientation of the fast polarization
1101 direction constrains the projection of the lineation on the horizontal plane to be
1102 oriented in the NNE direction ($9\pm 11^\circ$). However, there are no constraints on its
1103 plunge or on the dip of the foliation. To draw constraints on the orientation of
1104 the foliation and lineation in the lithospheric mantle, which would allow to
1105 discuss possible coupling between crustal and mantle structures, we estimate
1106 the contribution of the lithospheric mantle to the SKS splitting delay time (Δt)
1107 for three end-member orientations of the foliation and lineation in the
1108 lithospheric mantle, illustrated in Fig. 14, and compare these predictions to the
1109 observations at RCBR.

1110 Common conversion point (CCP) receiver function stacks support that the
1111 Moho is at ~32km (Almeida et al., 2015) and the LAB is at ~80km based on the
1112 S receiver function data by Heit et al. (2007) in the Northern Borborema
1113 Province. The thickness of the lithospheric mantle in this region is therefore 48
1114 km. Based on the thermobarometric data (Fig. 10), we divided the lithosphere
1115 into two layers and calculated the lithospheric mantle contribution to the
1116 measured delay time using the average seismic anisotropy of the low
1117 temperature peridotites (32 - 56km, < 1000 °C) and of the high-temperature
1118 ones (56 - 80 km, > 1000 °C).

1119

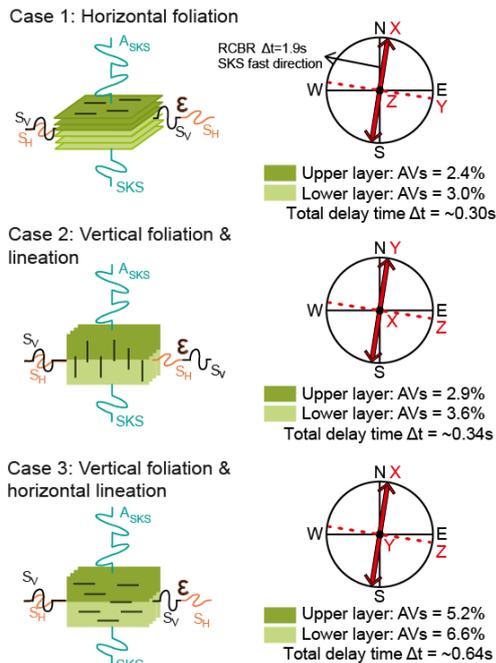


Fig. 14. Estimation of the maximum SKS splitting that may be produced in the lithospheric mantle for three end-member orientations of the flow directions and planes. Stereographic projections show the relation between the geographic (in black) and the structural (in red) reference frames in the three cases. The vertical is in all cases at the center of the diagram. The orientation of the fast SKS polarization (thick red arrow) is based on SKS splitting data for station RCBR.

1136

1137 If the foliation is horizontal and the lineation NE-SW (case 1), the S-wave
 1138 polarization anisotropy is 2.4% for upper lithospheric mantle and 3.0% for
 1139 lower lithospheric mantle. The cumulated SKS delay time is only $\sim 0.30\text{s}$. If
 1140 both the foliation and lineation are vertical (case 2), the fast SKS polarization
 1141 constrains the direction of the foliation, and SKS waves propagate parallel to
 1142 the lineation. The S-wave polarization anisotropy in this direction is low, like in
 1143 case 1. It is 2.9% for upper layer and 3.6% for lower one and the delay time
 1144 that may be cumulated in the lithospheric mantle is $\sim 0.34\text{s}$. If the foliation is
 1145 vertical but the lineation is horizontal and parallel to the fast SKS polarization
 1146 direction (case 3), the S-wave polarization anisotropy is higher. It is 5.2% in the
 1147 upper lithospheric mantle and 6.6% for lower lithospheric mantle, leading to a
 1148 cumulate delay time in the lithospheric mantle of 0.64s .

1149 In case 3, the lithospheric mantle has a fabric consistent with the crustal
 1150 deformation around station RCBR, which is dominated by dextral strike-slip in
 1151 transpressional structures in the Seridó belt and in the shear zones that border
 1152 it. This case would therefore imply a structuration in the lithospheric mantle
 1153 coherent with the crustal deformation in the Brasiliano event. Seismic
 1154 anisotropy in the lithospheric mantle, with anisotropy directions parallel or
 1155 subparallel to the main nearby Neoproterozoic shear zones is also required by
 1156 receiver function analysis on most stations in the Borborema Province
 1157 (Lamarque and Julià, 2019). At station RCBR, this study proposes horizontal
 1158 anisotropy axis trending NNE in the lithospheric mantle, consistent with case 3.

1159 It is important to note that even for case 3, the lithospheric mantle can
 1160 contribute to $< 1/3$ of the SKS delay time measured at the RCBR station ($1.9 \pm$
 1161 0.2 s , Assumpção et al., 2011). Thus a large part of the SKS signal at RCBR
 1162 has to be produced in the asthenosphere and, for the contributions of the
 1163 lithosphere and asthenosphere to add up, asthenospheric flow directions

1164 should not deviate much from NNE. Present-day absolute plate motion (APM)
1165 directions for NE Brazil do not follow this direction. Hotspot reference frame
1166 models, such as HS3-Nuvel-1A, predict an ENE direction, whereas no-net
1167 rotation models predict a NNW direction (Gripp and Gordon, 2002). However,
1168 a NNE flow direction in the sublithospheric mantle beneath the Borborema
1169 province is predicted by models in which the present-day mantle flow is
1170 calculated based on a density anomaly distribution derived from global seismic
1171 tomography models (cf. Fig. 8 in Assumpção et al., 2011).

1172

1173 *5.4. Relations between the mantle structure and the geodynamical evolution of* 1174 *the Borborema Province*

1175 When did the deformation, annealing, and reactive melt percolation
1176 processes recorded by the xenoliths happen? How do they relate to the
1177 geodynamical evolution of NE Brazil? There are no ways of dating deformation
1178 processes in the mantle. However, we may try to use the data discussed in the
1179 previous sections to constrain the imprint of the different tectonic events that
1180 affected the Borborema Province in the lithospheric mantle.

1181 The coarse-granular microstructures, with their well-equilibrated
1182 pyroxene-olivine grain boundaries, imply effective annealing, which, at the
1183 equilibrium temperatures of these xenoliths, require very long time delays, on
1184 the order of several hundreds of Ma. Although the temperature may have
1185 varied since the deformation of the xenoliths, to re-equilibrate the
1186 pyroxene-olivine grain boundaries in <100 Ma, temperatures $\geq 1100^{\circ}\text{C}$, that
1187 is $>200^{\circ}\text{C}$ above those recorded by the xenoliths at the time of extraction, are
1188 needed. Coarse-granular microstructures are well expressed among the
1189 low-equilibrium temperature xenoliths, suggesting that the shallow part of the
1190 lithospheric mantle beneath the Macau volcanics records essentially old
1191 tectonic events. The seismic anisotropy data in the region imply that this
1192 lithospheric structure results from Neoproterozoic dextral strike-slip and
1193 transpressional intraplate deformation. A pre-Mesozoic origin of this mantle
1194 fabric is also suggested by the higher annealing degree of the microstructures
1195 of the Borborema xenoliths relative to the Fernando de Noronha ones (Fig. 2).
1196 Indeed, Fernando de Noronha mantle xenoliths record deformation frozen up
1197 in the oceanic lithosphere by plate cooling, which post-dates the opening of the
1198 Atlantic (Liu et al., 2019).

1199 Some coarse-porphroclastic Iherzolites have lower equilibrium
1200 temperatures ($\leq 800^{\circ}\text{C}$) than most coarse-granular peridotites. This is not a
1201 sampling location effect, since both types of peridotites were sampled in Pico
1202 do Cabugi. The higher olivine intragranular misorientations and less
1203 equilibrated olivine and pyroxene grain shapes indicate less effective
1204 annealing, pointing to either slower annealing due to cooler temperatures or a
1205 more recent deformation event. In the first case, the deformation recorded by
1206 these xenoliths may also be associated with the Brasiliano event. In the
1207 second, these peridotite may record Cretaceous or even younger deformation

1208 preserved in the shallow lithospheric mantle. In any case, the studied xenolith
1209 suite exhibits no evidence of strong deformation under low temperature
1210 conditions. There are no xenoliths displaying mylonitic microstructures, with
1211 marked grain size reduction of olivine and strong elongation of olivine and
1212 orthopyroxene porphyroclasts, which are usually observed in extensional
1213 shear zones developed in peridotite massifs under low temperature conditions
1214 ($\leq 1000^{\circ}\text{C}$; e.g., Drury et al., 1991; Frets et al., 2014; Kaczmarek and Tommasi,
1215 2011). Although annealing could have partially obliterated the olivine
1216 deformation microstructure, it cannot change the aspect ratios of the
1217 orthopyroxenes in a time scale of 100 Ma. Thus if the low temperature
1218 coarse-granular peridotites correspond to sections of the lithospheric mantle
1219 deformed during the Mesozoic, these are low strain zones and their olivine
1220 CPO and seismic anisotropy may still preserve orientations produced by the
1221 previous deformation episodes.

1222 Based on the microstructures of the low equilibration temperature
1223 peridotites and on the seismic anisotropy data, we conclude therefore that the
1224 shallow sections of the lithospheric mantle of the Borborema province records
1225 coupled crust and mantle deformation during the formation of the Borborema
1226 shear zone system in the Neoproterozoic. The present dataset has no
1227 evidence for extensive reworking of the shallow lithospheric mantle by the
1228 extensional deformation in the Cretaceous or during the Cenozoic uplift of the
1229 province. Yet, the absence of (annealed or not) low-temperature mylonites in
1230 the xenolith sampling does not imply that shear zones accommodating a
1231 Mesozoic extension did not form in the shallow lithospheric mantle of the
1232 Borborema Province, since our sampling is punctual and such a deformation
1233 would be by nature heterogeneous.

1234 The coarse-porphyroclastic microstructures that characterize the lower
1235 part of the lithospheric mantle (equilibrium temperatures $> 1000^{\circ}\text{C}$) are more
1236 difficult to relate to a given tectonic episode. Partial melting conditions
1237 estimated for the Ceará-Mirim basalts suggest that the
1238 lithosphere-asthenosphere boundary was shallower in the Cretaceous than in
1239 the Cenozoic (cf. Fig. 10 and discussion section 5.2). This would imply that the
1240 deep lithospheric mantle beneath Borborema might be composed by material
1241 accreted by cooling since the Cretaceous. Comparison between the
1242 microstructures of these coarse-porphyroclastic peridotites with those of
1243 peridotite xenoliths from the nearby Fernando de Noronha (FN) archipelago,
1244 formed in response to Cenozoic volcanism onto 100-105 Ma old crust in the
1245 equatorial Atlantic (Liu et al., 2019) also favors a Mesozoic age for the
1246 deformation of these peridotites. Indeed, the olivine M2M and olivine and
1247 pyroxene shape factors of the Borborema peridotites overlap with the lower
1248 range of olivine M2M and shape factors, that is, with the most annealed
1249 microstructures of the FN peridotites (Fig. 2). The FN peridotites record an
1250 asthenospheric deformation, which has been frozen in the oceanic lithosphere
1251 by cooling and evolved by annealing since then (Liu et al., 2019). Their

1252 deformation is therefore younger than the opening of the Equatorial Atlantic.
1253 The similar to slightly stronger annealing degree of the coarse-porphyroclastic
1254 Macau peridotites suggests that the deformation that produced these
1255 microstructures in the deep section of the lithospheric mantle beneath the
1256 Borborema province might be related to the Cretaceous extension. Thus,
1257 although this event did not result in widespread deformation of the shallow
1258 levels of the lithospheric mantle, it might have reworked the base of the
1259 lithosphere (bottom-up lithospheric thinning?). Yet the high SKS delay times in
1260 station RCBR do not favor strong deviations in flow direction from NNE across
1261 the entire lithosphere-asthenosphere section, implying that either strains were
1262 small and did not change the Neoproterozoic CPO orientations or that flow
1263 directions during extension were at high angle to the extensional structures in
1264 the Cretaceous basins and to the Equatorial Atlantic spreading directions.

1265 Finally, the very low annealing degree and high equilibration temperatures
1266 of the fine-grained porphyroclastic microstructures necessarily imply
1267 deformation close in time to the Cenozoic Macau volcanism. Based on the
1268 microstructural evidence for high stresses, but low finite strains for this
1269 deformation, we propose that it is related to the formation of the dykes that
1270 brought the Macau volcanics to the surface. The equilibration temperatures of
1271 the xenoliths imply a rather hot geotherm, leading to an 80-km thick
1272 lithosphere in the Cenozoic. This is consistent with melting depths inferred for
1273 the Macau basalts (Ngonge et al., 2015b). It is also coherent with the
1274 present-day thermal state of the lithospheric and sublithospheric mantle as
1275 imaged by geophysical data .

1276 Dating the reactive melt percolation events is even more difficult than the
1277 deformation ones. The equilibrated microstructures and the coherent olivine
1278 and pyroxene CPOs in the coarse-granular peridotites undoubtedly point to a
1279 Brasiliano or older reactive melt percolation event. Coarse-porphyroclastic
1280 peridotites record younger melt percolation events, which in some cases
1281 predated or were synchronous to the main deformation recorded by the
1282 samples. However, in other cases, like in the coarse-porphyroclastic
1283 harzburgites, melt percolation post-dated the main deformation. Finally, the
1284 non-equilibrated chemical compositions at the sample scale observed in many
1285 samples, in particular the high-temperature coarse-porphyroclastic
1286 harzburgites, point to a last event of melt percolation that shortly predated the
1287 extraction of the peridotites. In summary, this peridotite suite recorded multiple
1288 reactive melt percolation events, probably well separated in time and related to
1289 the different magmatic episodes recorded in the crust.

1290

1291 **6. Conclusion**

1292 Integrated analysis of microstructures, crystal preferred orientations,
1293 mineral chemical compositions, and equilibrium temperatures in a suite of 22
1294 peridotite xenoliths reveals that the lithospheric mantle beneath the Northern
1295 Borborema Province preserves microstructures related to different deformation

1296 episodes since at least the Neoproterozoic. In all cases, olivine CPO points to
1297 deformation by dislocation creep with dominant activation of [100](010) slip
1298 system. However, the deformation microstructures were modified by variable
1299 degrees of annealing. The analysis of the extent of the annealing considering
1300 the equilibration temperatures allows rough 'dating' of the deformation
1301 episodes and relating them to the major deformation events recorded in the
1302 crust. The well-annealed olivine microstructures and pyroxene shapes in
1303 coarse-granular peridotites equilibrated at ca. 900°C indicate that the last
1304 deformation event that affected these peridotites is several hundreds of Ma old.
1305 In contrast, the fine-porphroclastic peridotites, which have equilibrium
1306 temperatures $\geq 1200^\circ\text{C}$, have suffered a high stress deformation, which shortly
1307 predated their extraction, probably related to the dykes that fed the Cenozoic
1308 volcanism. The coarse-porphroclastic microstructures, which are observed
1309 both in the shallow and deep lithospheric mantle are more difficult to relate to a
1310 given tectonic episode. Yet comparison between the microstructures of these
1311 peridotites and those of peridotite xenoliths from nearby Fernando de Noronha
1312 island, which sample the oceanic mantle lithosphere of an old domain of the
1313 Equatorial Atlantic, suggest that the high-temperature coarse-porphroclastic
1314 peridotites may record deformation related to the Cretaceous extension.
1315 Multiple reactive melt percolation events, probably well spaced in time, may
1316 also be inferred based on the microstructures, modal, and mineral
1317 compositions of the xenoliths.

1318 Comparison of the computed seismic anisotropy of the lithospheric mantle
1319 based on the xenolith data to SKS splitting in nearby RCBR station supports
1320 that the strongest contribution of the lithospheric mantle to the measured
1321 anisotropy would correspond to a frozen strike-slip fabric parallel to the major
1322 NNE-NE Neoproterozoic shear zones in the region. A shallow lithospheric
1323 mantle fabric parallel to the Neoproterozoic shear zones is also suggested by
1324 anisotropic receiver functions (Lamarque and Julia, 2019). These observations
1325 corroborate the conclusion that the shallow lithospheric mantle in the Northern
1326 Borborema province still preserves a structure acquired by coupled
1327 crust-mantle deformation during the formation of the Borborema shear zone
1328 system in the Neoproterozoic. It also suggests that Cretaceous extension,
1329 which seems to be recorded in the deeper sections of the lithosphere, did not
1330 produce pervasive reworking of the shallow lithospheric mantle, pointing to
1331 'partial' or total crust-mantle decoupling during this event. However, even if the
1332 entire lithospheric mantle has a frozen strike-slip fabric parallel to the major
1333 NNE-NE Neoproterozoic shear zones in the region, it can produce $<1/3$ of the
1334 measured delay time of 1.9s in station RCBR. Thus most of the measured SKS
1335 splitting in RCBR should record flow in the sublithospheric mantle, which also
1336 has a NNE orientation, which is not parallel to the APM, but is consistent with
1337 predictions of mantle circulation models for this region.

1338 Finally, equilibrium temperatures and petrological compositions of the
1339 xenoliths indicate a rather hot Cenozoic geotherm, implying a ca. 80 km thick

1340 lithosphere. This estimate is consistent with the melting conditions estimated
1341 for the formation of the Macau basalts (Ngonge et al. 2015). It is also coherent
1342 with geophysical data that point to a present-day 80-km thick lithosphere (Heit
1343 et al. 2007) and hotter than average sublithospheric mantle beneath this region
1344 (French et al., 2013; Simões Neto et al., 2019).

1345

1346 **7. Acknowledgments**

1347 S. Liu was funded during his 14-month stay in Geosciences Montpellier
1348 by the China Scholarship Council. M.M. was supported by projects MIUR-PRIN
1349 2005-2005049821 and 2017-20178LPCPW. C. Archanjo and M.H.B.M.
1350 Hollanda are thanked for assistance during field work. D. Delmas and C.
1351 Nevado prepared high-quality polished thin sections for EBSD measurements,
1352 which were performed with the assistance of F. Barou at the EBSD-SEM INSU
1353 national facility at Géosciences Montpellier (CNRS & Université de
1354 Montpellier). Electron microprobe analyses were carried out with the
1355 assistance of B. Boyer at the Service Microsonde Sud, Université de
1356 Montpellier. S. Piazzolo and T. Morishita are warmly thanked for their
1357 constructive reviews.

1358 The data used in this article are presented in the figures, tables, and
1359 supporting material. The raw EBSD data are available from the corresponding
1360 author upon request.

1361

1362

1363 **8. References**

- 1364 Abramson, E.H., Brown, J.M., Slutsky, L.J., Zaug, J.M., 1997. The elastic
1365 constants of San Carlos olivine to 17 GPa. *J Geophys Res-Sol Ea.* 102,
1366 12253-12263.
- 1367 Almeida, Y.B., Julià, J., Frassetto, A., 2015. Crustal architecture of the
1368 Borborema Province, NE Brazil, from receiver function CCP stacks:
1369 Implications for Mesozoic stretching and Cenozoic uplift. *Tectonophys.* 649,
1370 68-80.
- 1371 Ancelmi, M.F., Santos, T.J.S.d., Amaral, W.d.S., Fuck, R.A., Dantas, E.L.,
1372 Zincone, S.A., 2015. Provenance of metasedimentary rocks from the Ceará
1373 Central Domain of Borborema Province, NE Brazil: implications for the
1374 significance of associated retrograded eclogites. *J. S. Am. Earth Sci.* 58,
1375 82-99.
- 1376 Anderson, O.L., Isaak, D., Oda, H., 1992. High-Temperature Elastic-Constant
1377 Data on Minerals Relevant to Geophysics. *Rev. Geophys.* 30, 57-90.
- 1378 Archanjo, C. J., Hollanda, M. H. B., Rodrigues, S. W., Neves, B. B., &
1379 Armstrong, R., 2008. Fabrics of pre-and syntectonic granite plutons and
1380 chronology of shear zones in the Eastern Borborema Province, NE Brazil. *J.*
1381 *Struct. Geol.* 30, 310-326
- 1382 Archanjo, C.J., Viegas, L.G.F., Hollanda, M.H.B.M., Souza, L.C., Liu, D., 2013.
1383 Timing of the HT/LP transpression in the Neoproterozoic Seridó Belt

1384 (Borborema Province, Brazil): Constraints from UPb (SHRIMP)
1385 geochronology and implications for the connections between NE Brazil and
1386 West Africa. *Gondwana Res.* 23, 701-714.

1387 Assumpção, M., Guarido, M., van der Lee, S., Dourado, J.C., 2011.
1388 Upper-mantle seismic anisotropy from SKS splitting in the South American
1389 stable platform: A test of asthenospheric flow models beneath the
1390 lithosphere. *Lithosphere* 3, 173-180.

1391 Bachmann, F., Hielscher, R., Schaeben, H., 2010. Texture Analysis with MTEX
1392 – Free and Open Source Software Toolbox. *Solid State Phenomena* 160,
1393 63-68.

1394 Bachmann, F., Hielscher, R., Schaeben, H., 2011. Grain detection from 2d and
1395 3d EBSD data - Specification of the MTEX algorithm. *Ultramicroscopy* 111,
1396 1720-1733.

1397 Baker, M.B., Stolper, E.M., 1994. Determining the Composition of
1398 High-Pressure Mantle Melts Using Diamond Aggregates. *Geochim.*
1399 *Cosmochim. Acta* 58, 2811-2827.

1400 Baptiste V., Tommasi, A., Demouchy S., 2012. Deformation and hydration of
1401 the lithospheric mantle beneath the Kaapval craton. *Lithos* 149, 31-50.

1402 Barruol, G., Bonnin, M., Pedersen, H., Bokelmann, G.H.R. & Tiberi, C., 2011.
1403 Belt-parallel mantle flow beneath a halted continental collision: the Western
1404 Alps. *Earth Planet. Sci. Lett.* 302, 429-438 .

1405 Bascou, J., Tommasi, A., Mainprice, D. 2002. Plastic deformation and
1406 development of clinopyroxene lattice preferred orientations in eclogites. *J.*
1407 *Struct. Geol.*, **24**: 1357-1368

1408 Bastow, I.D., Julia, J., do Nascimento, A.F., Fuck, R.A., Buckthorp, T.L.,
1409 McClellan, J.J., 2015. Upper mantle anisotropy of the Borborema Province,
1410 NE Brazil: Implications for intra-plate deformation and sub-cratonic
1411 asthenospheric flow. *Tectonophys.* 657, 81-93.

1412 Berger, E.T., Vannier, M., 1984. Dunites in Alkali Basalts from Oceanic
1413 Volcanic Islands - Petrological Approach. *Bull. Mineral.* 107, 649-663.

1414 Bodinier, J.L., Garrido, C.J., Chanefo, I., Bruguier, O., Gervilla, F., 2008. Origin
1415 of Pyroxenite-Peridotite Veined Mantle by Refertilization Reactions:
1416 Evidence from the Ronda Peridotite (Southern Spain). *J. Petrol.* 49,
1417 999-1025.

1418 Bodinier, J.L., Godard, M., 2014. Orogenic, Ophiolitic, and Abyssal Peridotites,
1419 in: Holland, H.D., Turekian, K.K. (Eds.), *Treatise on Geochemistry* (Second
1420 Edition). Elsevier, Oxford, pp. 103-167.

1421 Boullier, A.M., 1977. Structure des péridotites en enclaves dans les kimberlites
1422 d'Afrique du Sud. Conséquences sur la constitution du manteau supérieur.
1423 *Bull. Soc. Française Minéral. Cristall.* 100, 214–229.

1424 Bonnin, M., Tommasi, A., Hassani, R., Chévrot, S., Wookey, J., Barruol, G.,
1425 2011. Numerical modeling of upper mantle anisotropy beneath a migrating
1426 strike-slip plate boundary : the San Andreas fault system. *Geophys. J.*
1427 *Intern.* 191, 436-458

1428 Brey, G.P., Kohler, T., 1990. Geothermobarometry in 4-Phase Lherzolites .2.
1429 New Thermobarometers, and Practical Assessment of Existing
1430 Thermobarometers. *J. Petrol.* 31, 1353-1378.

1431 Bystricky, M., Kunze, K., Burlini, L., Burg, J. P.; 2000. High shear strain of
1432 olivine aggregates: Rheological and seismic consequences. *Science* 290,
1433 1564-1567

1434 Caby, R., Sial, A., Arthaud, M., Vauchez, A., 1991. Crustal evolution and the
1435 Brasiliano orogeny in Northeast Brazil. In: *The West African orogens and*
1436 *Circum Atlantic correlatives.* Springer- Verlag, pp. 373-397.

1437 Castro, D.L., Bezerra, F.H.R., Sousa, M.O.L., Fuck, R.A., 2012. Influence of
1438 Neoproterozoic tectonic fabric on the origin of the Potiguar Basin,
1439 northeastern Brazil and its links with West Africa based on gravity and
1440 magnetic data. *J. Geodyn.* 54, 29-42

1441 Castro, D.L., Oliveira, D.C., Gomes Castelo Branco, R.M., 2007. On the
1442 tectonics of the Neocomian Rio do Peixe Rift Basin, NE Brazil: Lessons
1443 from gravity, magnetics, and radiometric data. *J. S. Am. Earth Sci.* 24,
1444 184-202.

1445 Chai, M., Brown, J.M., Slutsky, L.J., 1997. The elastic constants of an
1446 aluminous orthopyroxene to 12.5 GPa. *J Geophys Res-Sol Ea.* 102,
1447 14779-14785.

1448 Chopra, P.N., Paterson, M.S., 1981. The Experimental Deformation of Dunite.
1449 *Tectonophys.* 78, 453-473.

1450 Comin-Chiaramonti, P., Demarchi, G., Girardi, V.A.V., Princivalle, F., Sinigoi, S.,
1451 1986. Evidence of mantle metasomatism and heterogeneity from peridotite
1452 inclusions of northeastern Brazil and Paraguay. *Earth Planet. Sci. Lett.* 77,
1453 203-217.

1454 Costa, A.P., Cunha, A.L.C., Cavalcante, R., Medeiros, V.C., Spicily, A.L.,
1455 Dantas, A.R., 2016. Carta Geologica Integrada de Areas de Relevante
1456 Interesse Mineral: Area Serido, Projeto Metalogenia das Provincias
1457 Mineraias do Brasil, Escala 1:350000, Companhia de Pesquisas de
1458 Recursos Mineraias, Brazil.

1459 Darros de Matos, R.M., 1999. History of the northeast Brazilian rift system:
1460 kinematic implications for the break-up between Brazil and West Africa.
1461 *Geological Society, London, Special Publications.* 153, 55-73.

1462 Drury, M.R., Vissers, R.L.M., Vanderwal, D., Strating, E.H.H., 1991. Shear
1463 Localization in Upper Mantle Peridotites. *Pure Appl Geophys.* 137, 439-460.

1464 Durham, W.B. and Goetze, C., 1977. Plastic flow of oriented single crystals of
1465 olivine. 1.Mechanical data. *J. Geophys. Res.* 82: doi:
1466 10.1029/JB082i036p05737

1467 Fei, H., Koizumi, S., Sakamoto, N., Hashiguchi, M., Yurimoto, H., Marquardt,
1468 K., Katsura, T., 2016. New constraints on upper mantle creep mechanism
1469 inferred from silicon grain-boundary diffusion rates. *Earth Planet. Sci. Lett.*
1470 433, 350-359.

1471 Fisler, D. K., Mackwell, S. J., Petsch, S., 1997. Grain boundary diffusion in

1472 enstatite. *Physics and Chemistry of Minerals*. 24, 264-273.

1473 Fodor, R.V., Sial, A.N., Gandhok, G., 2002. Petrology of spinel peridotite
1474 xenoliths from northeastern Brazil: lithosphere with a high geothermal
1475 gradient imparted by Fernando de Noronha plume. *J. S. Am. Earth Sci.* 15,
1476 199-214.

1477 French, S., Lekic, V., Romanowicz, B., 2013. Waveform Tomography Reveals
1478 Channeled Flow at the Base of the Oceanic Asthenosphere. *Science*. 342,
1479 227-230.

1480 Frets, E.C., Tommasi, A., Garrido, C.J., Vauchez, A., Mainprice, D., Targuisti,
1481 K., Amri, I., 2014. The Beni Bousera Peridotite (Rif Belt, Morocco): an
1482 Oblique-slip Low-angle Shear Zone Thinning the Subcontinental Mantle
1483 Lithosphere. *J. Petrol.* 55, 283-313.

1484 Ganade de Araujo, C.E., Weinberg, R.F. and Cordani, U.G., 2014. Extruding
1485 the Borborema Province (NE-Brazil): a two-stage Neoproterozoic collision
1486 process. *Terra Nova* 26, 157-168.

1487 Ganade de Araujo, C.E., Cordani, U.G., Agbossoumounde, Y., Caby, R., Basei,
1488 M.A.S., Weinberg, R.F. and Sato, K., 2016. Tightening-up NE Brazil and NW
1489 Africa connections: New U–Pb/Lu–Hf zircon data of a complete plate
1490 tectonic cycle in the Dahomey belt of the West Gondwana Orogen in Togo
1491 and Benin. *Precambrian Res.* 276, 24-42.

1492 Gourié, K., Cordier, P., Garel, F., Thoraval, C., Demouchy, S., Tommasi, A.,
1493 Carrez, P., 2019. Dislocation dynamics modelling of the power-law
1494 breakdown in olivine single crystals: Toward a unified creep law for the
1495 upper mantle. *Earth Planet. Sci. Lett.* 506, 282-291.

1496 Green, H.W., Gueguen, Y., 1974. Origin of Kimberlite Pipes by Diapiric
1497 Upwelling in Upper Mantle. *Nature*. 249, 617-620.

1498 Gripp, A., Gordon, R., 2002. Young tracks of hotspots and current plate
1499 velocities. *Geophys. J. Int.* 150, 321–361.

1500 Hammond, J.O.S., Kendall, J.M., Wookey, J., Stuart, G.W., Keir, D., Ayele, A.,
1501 2014. Differentiating flow, melt, or fossil seismic anisotropy beneath Ethiopia,
1502 *Geochem. Geophys. Geosyst.* (15) DOI:10.1002/2013GC005185

1503 Hamza, V.M., Vieira, F.P., Silva, R.T.A., 2018. Anomalous heat flow belt along
1504 the continental margin of Brazil. *Int J Earth Sci.* 107, 19-33.

1505 Hansen, L. N., Zhao, Y. H., Zimmerman, M. E., Kohlstedt, D. L., 2014.
1506 Protracted fabric evolution in olivine: Implications for the relationship among
1507 strain, crystallographic fabric, and seismic anisotropy. *Earth Planet. Sci. Lett.*
1508 387, 157-168

1509 Heine, C., Zoethout, J., Muller, R.D., 2013. Kinematics of the South Atlantic rift.
1510 *Solid Earth* 4, 215-253.

1511 Heit, B., Sodoudi, F., Yuan, X., Bianchi, M., Kind, R., 2007. An S receiver
1512 function analysis of the lithospheric structure in South America. *Geophys.*
1513 *Res. Lett.* 34.

1514 Hielscher, R., Schaeben, H., 2008. A novel pole figure inversion method:
1515 specification of the MTEX algorithm. *J. Appl. Crystallogr.* 41, 1024-1037.

1516 Higgle, K., Tommasi, A., 2012. Feedbacks between deformation and melt
1517 distribution in the crust–mantle transition zone of the Oman ophiolite. *Earth*
1518 *Planet. Sci. Lett.* 359, 61-72.

1519 Higgle, K., Tommasi, A., 2014. Deformation in a partially molten mantle:
1520 Constraints from plagioclase Iherzolites from Lanzo, western Alps.
1521 *Tectonophys.* 615, 167-181.

1522 Hirth, G., Kohlstedt, D., 2003. Rheology of the upper mantle and the mantle
1523 wedge: A view from the experimentalists. *Inside the subduction Factory*, 138,
1524 83-105.

1525 Hollanda, M.H.B.M., Archanjo, C.J., Bautista, J.R., Souza, L.C., 2015. Detrital
1526 zircon ages and Nd isotope compositions of the Seridó and Lavras da
1527 Mangabeira basins (Borborema Province, NE Brazil): Evidence for
1528 exhumation and recycling associated with a major shift in sedimentary
1529 provenance. *Precambrian Res.* 258, 186-207.

1530 Hollanda, M.H.B.M., Archanjo, C.J., Souza, L.C., Liu, D.Y., Armstrong, R.,
1531 2011. Long-lived Paleoproterozoic granitic magmatism in the
1532 Serido-Jaguaribe domain, Borborema Province-NE Brazil. *J. S. Am. Earth*
1533 *Sci.* 32, 287-300.

1534 Hollanda, M.H.B.M., Archanjo, C.J., Macedo Filho, A.A., Fossen, H., Ernst,
1535 R.E., Castro, D.L.d., Melo, A.C. and Oliveira, A.L., 2018. The Mesozoic
1536 Equatorial Atlantic Magmatic Province (EQUAMP), in: *Dyke Swarms of the*
1537 *World: A Modern Perspective*. Springer Singapore, pp. 87-110.

1538 Isaak, D.G., Ohno, I., Lee, P.C., 2006. The elastic constants of monoclinic
1539 single-crystal chrome-diopside to 1,300 K. *Phys Chem Miner.* 32, 691-699.

1540 Jackson, J.M., Sinogeikin, S.V., Bass, J.D., 2007. Sound velocities and
1541 single-crystal elasticity of orthoenstatite to 1073 K at ambient pressure.
1542 *Phys. Earth Planet. Inter.* 161, 1-12.

1543 Kaczmarek, M.A., Tommasi, A., 2011. Anatomy of an extensional shear zone
1544 in the mantle, Lanzo massif, Italy. *Geochem. Geophys. Geosyst.* 12,
1545 Q0AG06, doi:10.1029/2011GC003627.

1546 Kelemen, P.B., 1990. Reaction between Ultramafic Rock and Fractionating
1547 Basaltic Magma .1. Phase-Relations, the Origin of Calc-Alkaline Magma
1548 Series, and the Formation of Discordant Dunite. *J. Petrol.* 31, 51-98.

1549 Klemme, S., 2004. The influence of Cr on the garnet-spinel transition in the
1550 Earth's mantle: experiments in the system MgO-Cr₂O₃-SiO₂ and
1551 thermodynamic modelling. *Lithos* 77, 639-646.

1552 Knesel, K.M., Souza, Z.S., Vasconcelos, P.M., Cohen, B.E., Silveira, F.V., 2011.
1553 Young volcanism in the Borborema Province, NE Brazil, shows no evidence
1554 for a trace of the Fernando de Noronha plume on the continent. *Earth*
1555 *Planet. Sci. Lett.* 302, 38-50.

1556 Kostopoulos, D.K., 1991. Melting of the Shallow Upper Mantle - a New
1557 Perspective. *J. Petrol.* 32, 671-699.

1558 Kourim, F., Vauchez, A., Bodinier, J. L., Alard, O., Bendaoud, A., 2015.
1559 Subcontinental lithosphere reactivation beneath the Hoggar swell (Algeria):

1560 Localized deformation, melt channeling and heat advection. *Tectonophys.*
1561 650, 18-33.

1562 Lamarque, G., Julià, J., 2019. Lithospheric and sub-lithospheric deformation
1563 under the Borborema Province of NE Brazil from receiver function harmonic
1564 stripping. *Solid Earth Discuss.* <https://doi.org/10.5194/se-2019-41>.

1565 Le Roux, V., Bodinier, J.L., Tommasi, A., Alard, O., Dautria, J.M., Vauchez, A.,
1566 Riches, A.J.V., 2007. The Lherz spinel Iherzolite: Refertilized rather than
1567 pristine mantle. *Earth Planet. Sci. Lett.* 259, 599-612.

1568 Lima, M.V.A.G.d., Berrocal, J., Soares, J.E.P., Fuck, R.A., 2015. Deep seismic
1569 refraction experiment in northeast Brazil: New constraints for Borborema
1570 province evolution. *J. S. Am. Earth Sci.* 58, 335-349.

1571 Lissenberg, C.J., Dick, H.J.B., 2008. Melt–rock reaction in the lower oceanic
1572 crust and its implications for the genesis of mid-ocean ridge basalt. *Earth*
1573 *Planet. Sci. Lett.* 271, 311-325.

1574 Liu, S., Tommasi, A., Vauchez, A., Mazzucchelli, M. 2019. Deformation,
1575 annealing, melt-rock interaction, and seismic properties of an old domain of
1576 the equatorial Atlantic lithospheric mantle. *Tectonics* 38, doi:
1577 1029/2018TC005373.

1578 Luz, R.M.N., Julià, J., do Nascimento, A.F., 2015. Bulk crustal properties of the
1579 Borborema Province, NE Brazil, from P-wave receiver functions:
1580 Implications for models of intraplate Cenozoic uplift. *Tectonophys.* 644-645,
1581 81-91.

1582 Mainprice, D., Bachmann, F., Hielscher, R., Schaeben, H., 2014. Descriptive
1583 tools for the analysis of texture projects with large datasets using MTEX:
1584 strength, symmetry and components. *Geol. Soc. London Spec. Publ.* 409,
1585 251-271.

1586 Mainprice, D., Hielscher, R., Schaeben, H., 2011. Calculating anisotropic
1587 physical properties from texture data using the MTEX open-source package.
1588 *Geol. Soc. London Spec. Publ.* 360, 175-192.

1589 Marques, F.O., Nogueira, F.C.C., Bezerra, F.H.R., Castro, D.L., 2014. The
1590 Araripe Basin in NE Brazil: An intracontinental graben inverted to a
1591 high-standing horst. *Tectonophys.* 630, 251-264.

1592 Klöcking, M., Tribaldos, V.R., Hoggard, M., White, N., Maclennan, J., Rezende
1593 Guimarães, A., 2018. Linking uplift and volcanism of the Borborema
1594 Province, northeast Brazil. *EGU General Assembly Conference Abstracts*
1595 20, p. 9966.

1596 Mercier, J.C.C., Nicolas, A., 1975. Textures and Fabrics of Upper-Mantle
1597 Peridotites as Illustrated by Xenoliths from Basalts. *J. Petrol.* 16, 454-487.

1598 Mizusaki, A.M.P., Thomaz-Filho, A., Milani, E.J., Césero, P., 2002. Mesozoic
1599 and Cenozoic igneous activity and its tectonic control in northeastern Brazil.
1600 *J. S. Am. Earth Sci.* 15, 183-198.

1601 Moulin, M., Aslanian, D., Unternehr, P., 2010. A new starting point for the South
1602 and Equatorial Atlantic Ocean. *Earth-Sci Rev.* 98, 1-37.

1603 Neves, S.P., 2003. Proterozoic history of the Borborema province (NE Brazil):

1604 Correlations with neighboring cratons and Pan-African belts and
1605 implications for the evolution of western Gondwana. *Tectonics*. 22.

1606 Ngonge, E.D., de Hollanda, M.H.B.M., Archanjo, C.J., de Oliveira, D.C.,
1607 Vasconcelos, P.M.P., Muñoz, P.R.M., 2016a. Petrology of continental
1608 tholeiitic magmas forming a 350-km-long Mesozoic dyke swarm in NE Brazil:
1609 Constraints of geochemical and isotopic data. *Lithos*. 258-259, 228-252.

1610 Ngonge, E.D., de Hollanda, M.H.B.M., Pimentel, M.M., de Oliveira, D.C.,
1611 2016b. Petrology of the alkaline rocks of the Macau Volcanic Field, NE
1612 Brazil. *Lithos*. 266-267, 453-470.

1613 Nimis, P., Grutter, H., 2010. Internally consistent geothermometers for garnet
1614 peridotites and pyroxenites. *Contrib Mineral Petr.* 159, 411-427.

1615 Niu, Y.L., 1997. Mantle melting and melt extraction processes beneath ocean
1616 ridges: Evidence from abyssal peridotites. *J. Petrol.* 38, 1047-1074.

1617 Nogueira, F.C.C., Marques, F.O., Bezerra, F.H.R., Castro, D.L., Fuck, R.A.,
1618 2015. Cretaceous intracontinental rifting and post-rift inversion in NE Brazil:
1619 Insights from the Rio do Peixe Basin. *Tectonophysics* 644-645, 92-107.

1620 O'Neill, H.S.C., 1981. The transition between spinel lherzolite and garnet
1621 lherzolite, and its use as a geobarometer. *Contrib. Mineral. Petr.* 77,
1622 185-194.

1623 Oliveira, R.G., Medeiros, W.E., 2012. Evidences of buried loads in the base of
1624 the crust of Borborema Plateau (NE Brazil) from Bouguer admittance
1625 estimates. *J. S. Am. Earth Sci.* 37, 60-76.

1626 Perlingeiro, G., Vasconcelos, P.M., Knesel, K.M., Thiede, D.S., Cordani, U.G.,
1627 2013. Ar-40/Ar-39 geochronology of the Fernando de Noronha Archipelago
1628 and implications for the origin of alkaline volcanism in the NE Brazil. *J.*
1629 *Volcanol. Geotherm. Res.* 249, 140-154.

1630 Princivalle, F., Salviulo, G., Fabro, C., Demarchi, G., 1994. Inter- and
1631 intra-crystalline temperature and pressure estimates on pyroxenes from NE
1632 Brazil mantle xenoliths. *Contrib. Mineral. Petr.* 116, 1-6.

1633 Qi, C., Hansen, L. N., Wallis, D., Holtzman, B. K., Kohlstedt, D. L., 2018.
1634 Crystallographic Preferred Orientation of Olivine in Sheared Partially Molten
1635 Rocks: The Source of the "a-c Switch". *Geochem. Geophys. Geosyst.* 19,
1636 316-336

1637 Rivalenti, G., Mazzucchelli, M., Girardi, V.A.V., Vannucci, R., Barbieri, M.A.,
1638 Zanetti, A., Goldstein, S.L., 2000. Composition and processes of the mantle
1639 lithosphere in northeastern Brazil and Fernando de Noronha: evidence from
1640 mantle xenoliths. *Contrib. Mineral. Petr.* 138, 308-325.

1641 Rivalenti, G., Zanetti, A., Girardi, V.A.V., Mazzucchelli, M., Tassinari, C.C.G.,
1642 Bertotto, G.W., 2007. The effect of the Fernando de Noronha plume on the
1643 mantle lithosphere in north-eastern Brazil. *Lithos* 94, 111-131.

1644 Tribaldos, V. R., White, N.J., Roberts, G.G., Hoggard, M.J., 2017. Spatial and
1645 temporal uplift history of South America from calibrated drainage analysis.
1646 *Geochem. Geophys. Geosyst.* 18, 2321-2353.

1647 Sa, J.M., McReath, I., Leterrier, J., 1995. Petrology, Geochemistry and

1648 Geodynamic Setting of Proterozoic Igneous Suites of the Oros Fold Belt
1649 (Borborema Province, Northeast Brazil). *J. S. Am. Earth Sci.* 8, 299-314.
1650 Sang, L.Q., Bass, J.D., 2014. Single-crystal elasticity of diopside to 14 GPa by
1651 Brillouin scattering. *Phys. Earth Planet. Inter.* 228, 75-79.
1652 Santos, E.J., Schmus, W.R.V., Kozuch, M., Neves, B.B.d.B., 2010. The Cariris
1653 Velhos tectonic event in Northeast Brazil. *J. S. Am. Earth Sci.* 29, 61-76.
1654 Schutt, D.L., Lowry, A.R., Buehler, J.S., 2018. Moho temperature and mobility of lower
1655 crust in the western United States. *Geology* 46, 219-222.
1656 Silveira, F.V., 2006. Magmatismo cenozóico da porção central do Rio Grande
1657 do Norte, NE do Brasil. Ph.D. Thesis, Univ. Federal do Rio Grande do Norte,
1658 UFRN, Brazil.
1659 Simões Neto, F.L., Julià, J., Schimmel, M., 2019. Upper-mantle structure of the
1660 Borborema Province, NE Brazil, from *P*-wave tomography: implications for
1661 rheology and volcanism, *Geophys. J. Intern.* 216, 231-250.
1662 Skemer, P., Karato, S.I., 2008. Sheared Iherzolite xenoliths revisited. *J.*
1663 *Geophys. Res.* 113, B07205.
1664 Souza, Z.S., Vasconcelos, P., Nascimento, M., V. Silveira, F., S. Paiva, H., Dias,
1665 L.G., Thiede, D., Carmo, I., 2003. 40Ar/39Ar geochronology of Mesozoic
1666 and Cenozoic magmatism in NE Brazil. In Proceedings of IV South
1667 American Symposium on Isotope Geology, 691-694.
1668 Souza, Z.S., Kalsbeek, F., Deng, X.-D., Frei, R., Kokfelt, T.F., Dantas, E.L., Li,
1669 J.-W., Pimentel, M.M., Galindo, A.C., 2016. Generation of continental crust
1670 in the northern part of the Borborema Province, northeastern Brazil, from
1671 Archaean to Neoproterozoic. *J. S. Am. Earth Sci.* 68, 68-96.
1672 Taylor, W.R., 1998. An experimental test of some geothermometer and
1673 geobarometer formulations for upper mantle peridotites with application to
1674 the thermobarometry of fertile Iherzolite and garnet websterite. *Neues Jb.*
1675 *Miner. Abh.* 172, 381-408.
1676 Tikoff, B., Russo, R., Teyssier, C., Tommasi, A. 2004. Mantle-driven
1677 deformation of orogenic zones and clutch tectonics. In: J. Grocott, K.
1678 McCaffrey, G. Taylor & B. Tikoff (eds) Vertical and horizontal decoupling in
1679 the lithosphere. *Geol. Soc. London Spec. Publ.* 227, 41-64.
1680 Tommasi, A., Ishikawa, A., 2014. Microstructures, composition, and seismic
1681 properties of the Ontong Java Plateau mantle root. *Geochem. Geophys.*
1682 *Geosyst.*, 15, doi: 10.1002/2014GC005452
1683 Tommasi, A., Vauchez, A., 2015. Heterogeneity and anisotropy in the
1684 lithospheric mantle. *Tectonophys.* 661, 11-37.
1685 Tommasi, A., Vauchez, A., Daudré, B., 1995. Initiation and propagation of
1686 shear zones in a heterogeneous continental lithosphere. *J. Geophys. Res.*
1687 100, 22083-22101.
1688 Tommasi, A., Mainprice, D., Canova, G., Chastel, Y. 2000. Viscoplastic
1689 self-consistent and equilibrium-based modeling of olivine lattice preferred

1690 orientations. Implications for the upper mantle seismic anisotropy. *J.*
1691 *Geophys. Res.* 105, 7893-7908

1692 Tommasi, A., Tikoff, B., Vauchez, A. 1999. Upper mantle tectonics:
1693 Three-dimensional deformation, olivine crystallographic fabrics and seismic
1694 properties. *Earth Planet. Sci. Lett.* 168, 173-186

1695 Tommasi, A., Vauchez, A., Ionov, D. A., 2008. Deformation, static
1696 recrystallization, and reactive melt transport in shallow subcontinental
1697 mantle xenoliths (Tok Cenozoic volcanic field, SE Siberia). *Earth Planet. Sci.*
1698 *Lett.* 272, 65-77

1699 Ussami, N., Molina, E.C., Medeiros, W.E., 1999. Novos vinculos sobre a
1700 Evolução Térmica da Margem Continental Leste do Brasil, *In VII National*
1701 *Symposium on Tectonic Studies, Extended Abstracts*: Sociedade Brasileira
1702 de Geologia, 7, 20–23

1703 Van der Wal, D., Chopra, P., Drury, M., Gerald, J.F., 1993. Relationships
1704 between dynamically recrystallized grain size and deformation conditions in
1705 experimentally deformed olivine rocks. *Geophys. Res. Lett.* 20, 1479-1482.

1706 van Schmus, W.R., Oliveira, E.P., da Silva Filho, A.F., Toteu, S.F., Penaye, J.,
1707 Guimarães, I.P., 2008. Proterozoic links between the Borborema Province,
1708 NE Brazil, and the Central African Fold Belt. *Geol. Soc. London Spec. Publ.*
1709 294, 69-99.

1710 Vauchez, A., Nicolas, A., 1991. Mountain building: strike-parallel motion and
1711 mantle anisotropy. *Tectonophys.* 185, 183-201

1712 Vauchez, A., Dineur, F., Rudnick, R., 2005. Microstructure, texture and seismic
1713 anisotropy of the lithospheric mantle above a mantle plume: insights from
1714 the Labait volcano xenoliths (Tanzania). *Earth and Planetary Science*
1715 *Letters* 232, 295-314.

1716 Vauchez, A., Neves, S., Caby, R., Corsini, M., Egydio-Silva, M., Arthaud, M.,
1717 Amaro, V., 1995. The Borborema Shear Zone System, NE Brazil. *J. S. Am.*
1718 *Earth Sci.* 8, 247-266.

1719 Vauchez, A., Tommasi, A., Mainprice, D., 2012. Faults (shear zones) in the
1720 Earth's mantle. *Tectonophys.* 558, 1-27.

1721 Viegas, L.G.F., Archanjo, C.J., Hollanda, M.H.B.M. and Vauchez, A., 2014.
1722 Microfabrics and zircon U/Pb (SHRIMP) chronology of mylonites from the
1723 Patos shear zone (Borborema Province, NE Brazil). *Precambrian Research*
1724 243, 1-17.

1725 Walter, M.J., Sisson, T.W., Presnall, D.C., 1995. A Mass Proportion Method for
1726 Calculating Melting Reactions and Application to Melting of Model
1727 Upper-Mantle Lherzolite. *Earth Planet. Sci. Lett.* 135, 77-90.

1728 Workman, R.K., Hart, S.R., 2005. Major and trace element composition of the
1729 depleted MORB mantle (DMM). *Earth Planet. Sci. Lett.* 231, 53-72.

1730 Wu, J., Zhang, Z., Kong, F., Yang, B.B., Yu, Y., Liu, K.H., Gao, S.S., 2015.
1731 Complex seismic anisotropy beneath western Tibet and its geodynamic
1732 implications. *Earth Planet. Sci. Lett.* 413, 167-175.

Table 1

[Click here to download Table: Table 1_revised_v2.pdf](#)

Table 1 Microstructures, modal compositions, thermometry, and bulk rock Mg# of Macau volcanics peridotite xenoliths

Sample	Location	Rock type	Microstructure	Modal compositions (%)				Thermometry (°C)								Bulk rock Mg#*	
								Two-pyroxene				Ca-in-opx					
				ol	opx	cpx	sp	Rim	sd	Core	sd	Rim	sd	Core	sd		
16CA07	Pico Cabugi	Lz	Coarse-granular	61	22	15	2	816	52	834	9	903	22	878	11	89.0	
16CA08	Pico Cabugi	Lz		58	26	15	1	834	21	830	27	905	13	869	6	89.0	
16CA09	Pico Cabugi	Lz		70	23	7	1	923	17	920	18	970	5	958	7	90.3	
16CA12	Pico Cabugi	Lz		69	19	10	2										
16CA14	Pico Cabugi	Lz		58	28	12	2	627	38	592	51	836	148	790	16	89.4	
BO9^^	Bodo	Lz		63	22	15	<1	946		914		899		893		90.6	
16CA01	Pico Cabugi	Hz	Coarse-porphroclastic	79	18	3	1	985	72	1026	180	955	3	970	41	90.9	
16CA02	Pico Cabugi	Hz		82	13	5	1										
16CA03	Pico Cabugi	Hz		87	10	3	<1	1024	8	1111	87	1119	148	1028	149	90.7	
16CA04	Pico Cabugi	Lz		59	23	15	4										
16CA05	Pico Cabugi	Lz		71	15	14	<1										
16CA06	Pico Cabugi	Lz		67	21	11	1	703	5	710	16	778	9	846	46	90.2	
16CA11	Pico Cabugi	Lz		72	8	18	2	693	8	691	25	813	21	809	18	89.5	
16CA15	Pico Cabugi	Lz		67	19	13	1	887	152	730	112	856	29	785	76	89.5	
16CA18	Pico Cabugi	Hz		82	16	2	1	1036	121	952	110	1005	106	942	8	90.9	
GR1^^	Fazenda Geroncio	Lz		76	16	7	1	1202	14	1183	36	1178	6	1180	7	90.0	
SV8^^	Serra Verde	Lz		77	13	9	1	1026		1008		1038		1034		90.6	
SV14^^	Serra Verde	We		53	0	45	1									91.0	
AG6^^	Serra Aguda	Lz		Fine-porphroclastic	61	24	13	1			1186	1			1192	10	89.4
AG7^^	Serra Aguda	Hz			76	20	3	<1			1215	6			1220	3	90.7
Pc105^	Pico Cabugi	Hz	72		25	4	<1			1210				1220		91.3	
Pc109^	Pico Cabugi	Hz	74		21	4	<1			1225				1233		90.4	

Hz: Harzburgite; Lz: Lherzolite; We: Wehrlite; ^ Samples from Rivalenti et al. (2000); ^^ Samples from Rivalenti et al, (2007)

* estimated by average chemical composition and the modal contents (area.%) of each mineral except Pc105, Pc109, AG6, BO9 and SV8 for which whole rock chemical data were available

Table 2

[Click here to download Table: Table 2_revised.pdf](#)

Table 2 Quantitative texture and microstructure parameters derived from EBSD mapping for olivine, orthopyroxene, and clinopyroxene

Sample	Microstructure Rex area fraction		Olivine							Orthopyroxene			Clinopyroxene	
			J-index	BA-index	M2M (°)	GOS (°)	Shape factor [^]	Grain size [^] (μm)	Aspect ratio [^]	Shape factor [^]	Grain size [^] (μm)	Aspect ratio [^]	Grain size [^] (μm)	Aspect ratio [^]
16CA07	Coarse-granular	Bulk-rock	5.8	0.34	1.56	1.27	1.67	1381	1.81	1.69	1126	1.61	827	1.61
16CA08		Bulk-rock	6.6	0.35	1.36	0.74	1.56	1493	1.69	1.51	1322	1.57	936	1.62
16CA09		Bulk-rock	6.4	0.46	1.14	1.61	1.59	2233	1.55	1.46	1401	1.53	872	1.60
16CA12		Bulk-rock	6.3	0.62	1.73	1.33	1.65	2134	1.59	1.49	1190	1.51	926	1.67
16CA14		Bulk-rock	4.7	0.33	1.46	1.06	1.64	1899	1.55	2.01	2688	1.57	1007	1.61
BO9 ^{^^}		Bulk-rock	6.5	0.84	1.41	0.88	1.85	1577	2.04	2.00	1669	1.80	1008	1.80
16CA01	Coarse-porphyroclastic	Bulk-rock	6.6	0.77	2.60	1.27	1.90	2428	1.86	1.64	1088	1.83	799	1.61
16CA02		Bulk-rock	6.2	0.81	2.67	1.52	1.73	2001	1.66	1.61	1194	1.62	636	1.70
16CA03		Bulk-rock	6.8	0.73	2.90	1.34	1.74	2140	1.78	1.95	894	1.94	248	1.86
16CA04		Bulk-rock	4.8	0.45	2.08	1.90	1.61	1813	1.56	2.14	2678	1.54	1124	1.63
16CA05		Bulk-rock	3.8	0.47	2.22	1.39	1.64	1856	1.62	1.85	2503	1.96	1074	1.61
16CA06		Bulk-rock	7.3	0.27	2.23	1.17	1.80	2192	1.64	2.12	3376	1.82	925	1.59
16CA11		Bulk-rock	5.8	0.31	3.35	1.48	1.71	2281	1.63	1.75	3298	1.58	1131	1.70
16CA15		Bulk-rock	5.1	0.48	1.66	0.78	1.90	1837	1.66	2.38	3279	1.76	947	1.69
16CA18		Bulk-rock	8.0	0.82	2.44	1.36	1.80	2416	1.93	1.78	1197	2.02	459	1.89
GR1		Bulk-rock	8.5	0.45	2.28	1.37	1.83	2507	1.87	1.75	1378	1.62	1048	1.52
SV8		Bulk-rock	9.0*	0.27	2.33	2.26	1.91	1589	2.04	1.75	1337	1.85	775	1.56
SV14		Bulk-rock	6.8	0.37	2.22	1.38	2.22	1655	1.68				1713	1.56
AG6		Fine-porphyroclastic	Bulk-rock	4.6	0.44	5.64	1.96	3.02	1221	1.69	1.89	1646	1.45	920
	Porphyroclasts		9.5	0.55	8.07	5.66	4.04	2143	1.74	1.89	1672	1.45	965	1.73
	Neoblasts (30%)		2.2	0.27	2.51	1.50	1.77	117	1.61	1.69	157	1.59	175	1.72
AG7	Bulk-rock		16.6*	0.40	3.05	0.89	2.83	3062	1.67	2.01	1301	1.62	729	1.63
	Porphyroclasts		24.6*	0.41	3.67	4.05	3.17	3814	1.67	2.06	1511	1.56	906	1.61
	Neoblasts (22%)		2.8	0.38	0.57	0.49	1.60	365	1.65	1.82	540	1.82	418	1.66
Pc105	Bulk-rock		6.4	0.25	6.18	1.47	2.20	2133	1.68	2.06	2772	1.75	748	1.52
	Porphyroclasts		8.2	0.26	7.22	5.30	2.36	2538	1.70	2.10	2919	1.74	865	1.48
	Neoblasts (15%)		2.1	0.26	0.54	0.56	1.49	311	1.57	1.48	464	1.91	435	1.61
Pc109	Bulk-rock		7.5	0.18	5.25	1.48	2.34	2330	1.74	2.56	1881	2.04	1082	1.95
	Porphyroclasts		9.2	0.18	5.96	4.86	2.48	2682	1.78	2.63	2046	2.04	1231	2.01
	Neoblasts (14%)		2.6	0.20	0.61	0.59	1.55	358	1.52	1.90	476	2.09	358	1.65

All values are averages over the entire EBSD map, weighted by the grain area

M2M: Misorientation relative to mean orientation of the grain; GOS: Grain orientation spread; Rex: recrystallized

[^] Grain sizes, aspect ratios, and shape factors are apparent 2D values. Aspect ratios may be underestimated since many sections were not cut on the XY structural plane.

* marks J-index values that are probably overestimated due to the small number of coarse grains in the thin section (small samples)

Table 3

[Click here to download Table: Table 3_revised.pdf](#)

Table 3. Calculated seismic properties for the individual and for the average low- and high-temperature Macau peridotites

	Sample	Rock type	Microstructures	Seismic properties (velocity in km/s and anisotropy in %)												
				Vp max	Vp min	Avp	AVs	Vs1 max	Vs1 min	AVs1	Vs2 max	Vs2 min	AVs2	Vp/Vs1 max	Vp/Vs1 min	AVp/Vs1
Low-temperature samples (<1000°C) P=1.46GPa T=916°C	16CA07	Lz	Coarse-granular	8.2	7.4	10.0	7.1	4.7	4.4	5.2	4.5	4.3	4.4	1.8	1.7	6.3
	16CA08	Lz	Coarse-granular	8.2	7.5	9.2	6.7	4.7	4.5	4.2	4.5	4.3	4.1	1.8	1.7	5.7
	16CA09	Lz	Coarse-granular	8.2	7.6	7.8	5.8	4.7	4.5	3.5	4.5	4.4	4.0	1.8	1.7	7.7
	16CA14	Lz	Coarse-granular	8.1	7.5	7.8	5.5	4.7	4.5	4.2	4.5	4.4	3.4	1.8	1.7	5.4
	BO9	Lz	Coarse-granular	8.3	7.6	9.3	6.0	4.7	4.5	3.0	4.6	4.3	5.2	1.8	1.7	9.2
	16CA06	Lz	Coarse-porph	8.2	7.5	9.1	7.3	4.7	4.4	5.9	4.6	4.3	4.6	1.8	1.7	5.5
	16CA11	Lz	Coarse-porph	8.2	7.6	7.7	6.5	4.7	4.5	3.8	4.5	4.4	3.2	1.8	1.7	5.5
	16CA15	Lz	Coarse-porph	8.2	7.6	7.4	5.6	4.6	4.5	3.4	4.5	4.4	4.3	1.8	1.7	5.3
	Average	-	-	-	8.2	7.6	7.9	5.2	4.7	4.5	3.6	4.6	4.4	3.7	1.8	1.7
High-temperature samples (>1000°C) P=2.12GPa T=1189°C	16CA01	Hx	Coarse-porph	8.4	7.4	12.3	8.1	4.6	4.4	4.1	4.5	4.2	7.1	1.9	1.7	10.9
	16CA03	Hx	Coarse-porph	8.5	7.4	14.0	10.0	4.7	4.4	6.3	4.5	4.2	7.3	1.9	1.6	13.5
	16CA18	Hx	Coarse-porph	8.6	7.4	14.2	9.7	4.7	4.4	5.7	4.5	4.2	7.2	1.9	1.6	14.7
	GR1	Lz	Coarse-porph	8.2	7.4	10.4	8.2	4.6	4.4	5.9	4.5	4.2	5.4	1.8	1.7	7.3
	SV8	Lz	Coarse-porph	8.2	7.4	11.3	9.0	4.7	4.3	7.2	4.5	4.2	5.8	1.8	1.7	6.6
	AG6	Lz	Fine-porph	8.0	7.4	8.2	6.3	4.6	4.4	5.0	4.4	4.3	3.6	1.8	1.7	5.9
	PC105	Hx	Fine-porph	8.0	7.4	7.7	5.9	4.6	4.3	5.7	4.4	4.3	1.9	1.8	1.7	3.5
	PC109	Hx	Fine-porph	8.1	7.4	9.1	7.4	4.6	4.3	6.4	4.4	4.3	3.0	1.8	1.7	4.3
	AG7	Hx	Fine-porph	8.2	7.4	10.4	8.7	4.7	4.4	6.3	4.5	4.2	5.8	1.8	1.7	9.3
	Average	-	-	-	8.3	7.4	10.5	7.7	4.7	4.4	5.8	4.5	4.3	4.9	1.8	1.7

Coarse-porph: Coarse-porphroclastic; Fine-porph: Fine-porphroclastic; Hx: Harzburgite; Lz: Lherzolite

AVp: Maximum P-wave propagation anisotropy; AVs: Maximum S-wave polarization anisotropy; AVs1: Maximum fast S-wave propagation anisotropy;

AVs2: Maximum slow S-wave propagation anisotropy; AVs1/Vp: Maximum anisotropy of Vp/Vs1 ratio

Supplementary material description

[Click here to download Supplementary material for online publication only: SupplementaryMaterial.docx](#)

Supplementary Fig. S1

[Click here to download Supplementary material for online publication only: FigS1_spinel_chemical.pdf](#)

Supplementary Table S1

[Click here to download Supplementary material for online publication only: Table_S1.xlsx](#)

Supplementary Table S2

[Click here to download Supplementary material for online publication only: Table_S2.xlsx](#)

Supplementary Table S3

[Click here to download Supplementary material for online publication only: Table_S3.xlsx](#)

Supplementary Table S4

[Click here to download Supplementary material for online publication only: Table_S4.xlsx](#)