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TESE DE DOUTORADO

Joint inversion of magnetotelluric and seismological observations: Case study in NE Brazil

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*To my mother for teaching me to be
a strong, sagemess and persevering
woman, capable of overcoming any
obstacle...*

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Resumo

Os métodos geofísicos tem como finalidade obter informações sobre as rochas e fluidos em subsuperfície. Para alcançar este objetivo, diferentes metodologias são empregadas para determinar com maior precisão tanto a geometria, quanto as propriedades físicas dos estratos da Terra em profundidade. Os métodos sismológicos recuperaram informações estruturais e de velocidade em subsuperfície, já as técnicas de MT são poderosas ferramentas para caracterizar a saturação de fluidos, como hidrocarbonetos ou água salina, nas rochas em subsuperfície. Medições sismológicas ou magnetotelúricas (MT) fornecem informações incompletas quando utilizadas independentemente. Combinadas, essas técnicas têm o potencial para melhorar consideravelmente o imageamento do arcabouço estrutural em profundidade. Este trabalho avalia o desempenho da análise conjunta de dados sismológicos e magnetotelúricos na determinação de estrutura subsuperficial sob a Província Borborema e Norte do Cráton São Francisco, através de dados coletados ao longo de um perfil N-S centrado na Bacia do Araripe.

Primeiramente, modelos de resistividade e velocidade de onda S foram obtidos através de inversões independentes. Os modelos de velocidade foram obtidos através da inversão simultânea de funções do receptor e curvas de dispersão, e revelaram a presença de um estrato com alta velocidade (4,1-4,3 km/s) abaixo da Bacia do Araripe - que foi interpretado como uma camada máfica na crosta inferior - e uma diminuição da velocidade ($< 4,3$ km/s) abaixo dos 120 km de profundidade. Os modelos de resistividade foram retirados de uma inversão 3D já existente de curvas de resistividade aparente e fase, observando-se

a presença de alta condutividade ($25 \Omega\text{m}$) no manto (120 km) sob a bacia. Esta anomalia condutiva foi interpretada como um material enriquecido com pequenas quantidades de magma remanescente dos processos de rifteamento do Mesozóico. A interpretação conjunta dos resultados evidencia assim um afinamento da litosfera abaixo da Bacia do Araripe que contribui a melhorar nossa compreensão dos processos de inversão tectônica dessa bacia.

Foi investigado também o desempenho da inversão conjunta dos dados magnetotélúricos e sismológicos coletados. A inversão conjunta utilizou as mesmas funções do receptor e curvas de fase e resistividade aparente das inversões independentes; porém, uma curva de dispersão de velocidade de fase média foi desenvolvida. A inversão assume uma estrutura sísmica e de resistividade comum (1D), representada por um conjunto de camadas acima de um semi-espaço. Além disso, a inversão é realizada através de um algoritmo genético multiobjetivo (GA) que explora eficientemente e totalmente o espaço de parâmetros. Nossos resultados demonstram que, apesar de uma complexidade geológica considerável, é possível determinar estruturas de velocidade e resistividade estruturalmente consistentes sob a Província Borborema e o cráton São Francisco. Os modelos invertidos (1D) não permitem desenvolver relações empíricas entre resistividade e velocidade sísmica de onda S em profundidades crustais ou mantélicas, mas conseguem imagear uma Moho elétrica sob a maior parte dos locais na região de estudo.

Palavras-chave: Inversão conjunta; Seismologia passiva; Magnetotelúrica; Inversão tectônica; Moho elétrica.

Abstract

Geophysical methods aim at obtaining information about the subsurface rocks and fluids. To achieve this goal, different methodologies are employed to more accurately determine both the geometry and physical properties of the Earth's strata at depth. Seismological methods recover structural and velocity information in the subsurface, while MT techniques are powerful tools to characterize the saturation of fluids, such as hydrocarbons or saline water, in subsurface rocks. Seismological or magnetotelluric (MT) measurements provide incomplete information when used independently. Combined, these techniques have the potential to considerably improve the imaging of structural features at depth. This paper evaluates the performance of joint analysis of seismological and magnetotelluric data for determination of subsurface structure beneath the Borborema Province and northern São Francisco Craton, using data collected along a N-S profile centered at the Araripe Basin.

First, resistivity and S-wave velocity models were obtained through independent inversions. The velocity models were obtained through simultaneous inversion of receiver functions and dispersion curves, and revealed the presence of a layer with high velocity (4.1-4.3 km/s) below the Araripe Basin - which was interpreted as a mafic layer in the lower crust - and a decrease in velocity (< 4.3 km/s) below 120 km depth. The resistivity models were taken from an existing 3D inversion of apparent resistivity and phase curves, noting the presence of high conductivity ($> 25 \Omega\text{m}$) in the mantle (120 km) beneath the basin. This conductive anomaly was interpreted as material enriched with small amounts

of magma derived from Mesozoic rifting processes. The joint interpretation of the results thus evidences a thinning of the lithosphere beneath the Araripe Basin, contributing to improve our understanding of the tectonic inversion processes of this basin.

The performance of the joint inversion of magnetotelluric and seismological data was also investigated. The joint inversion used the same receiver functions, phase and apparent resistivity curves as the independent inversions; however, a brand-new mean phase-velocity dispersion curve had to be utilized. The inversion assumes a common seismic and resistivity structure (1D), represented by a set of layers above a half-space. Furthermore, the inversion is performed using a multi-objective genetic algorithm (GA) that efficiently exploits the parameter space. Our results demonstrate that, in spite of considerable geological complexity, it is possible to determine structurally consistent velocity and resistivity structures beneath the Borborema Province and the São Francisco Craton. The inverse (1D) models do not allowed to develop empirical relationships between resistivity and S-wave seismic velocity at crustal or mantle depths, but they were able to image an electric Moho beneath most locations within the study region.

Keywords: Joint inversion; Passive Seismology; Magnetotelluric; Tectonic inversion; Electrical Moho.

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Chapter 1

Introduction

With the advance of passive-source geophysical techniques and the increase in computational power, it has been possible to improve on approaches for the joint inversion of disparate geophysical data sets. For example, seismological methods image seismic velocity and are a powerful tool for constraining temperature and composition of the subsurface, while magnetotelluric (MT) measurements image electrical conductivity and are more sensitive to the presence of fluids. Notably, the ability of MT techniques to discriminate rocks containing hydrocarbons from those containing saline water has made them a viable tool in the exploration and monitoring of reservoirs. Thus, altogether, these techniques have the potential for improving subsurface imaging considerably, reduce the influence of noise and limit the range of acceptable models.

However, because they are sensitive to different physical parameters, their joint inversion remains challenging. For this reason, different approaches have been proposed. The joint inversion using an empirical relationship between velocity and resistivity (Abubakar et al., 2012; Moorkamp et al., 2011; Jegen et al., 2009) is perhaps the approach providing the strongest coupling between the two physical sets of parameters, as one physical parameter is calculated from the other. Another approach is the joint petrophysical inversion using empirical relationships between properties other than velocity or resistivity,

through reconstruction of petrophysical parameters such as porosity and fluid saturation. This approach yields good results provided that there is *a priori* knowledge of the petrophysical relationships. Finally, the joint inversion with structural links relies on the fact that both techniques are sensing a common subsurface structure and, consequently, the physical parameters are expected to change at the same location without requiring any *a priori* petrophysical correlation. A popular way of applying such structural coupling has been through the use of a cross-gradient function (Gallardo et al., 2005; Abubakar et al., 2012; Hu et al., 2009).

The distribution of seismic velocity or electrical properties within the Earth are both governed by temperature, pressure and composition. Their discontinuities can often be interpreted as boundaries between distinct geological units (e.g. Marquis et al., 1995; Lahti et al., 2005; Tournier and Chouteau, 2005). In fact, the two principal discontinuities in the lithosphere are the crust-mantle boundary (Moho) and the lithosphere-asthenosphere boundary (LAB), which are both likely correlated with changes in conductivity (Jones, 1999; Jones and Ferguson, 2001; Gatzemeier and M., 2005) and seismic velocity (Zhu and Kanamori, 2000; Christensen and Mooney, 1995). However, in the case of seismic velocity, these discontinuities are best observed from different methods, with observation of the LAB being more challenging than the Moho; conversely, in the case of electric resistivity, it is the observation of the Moho that has traditionally been more challenging than the LAB ((Jones and Ferguson, 2001)). A major problem is that the asthenosphere is a gradational low-velocity zone, which is difficult to observe with seismic body waves; the electric Moho, in contrast, commonly occurs at the bottom of a conductive lower crust which is difficult to image with electromagnetic methods (Kind and Li, 2007). Some authors have studied the relationship between seismic velocity and electrical resistivity, with an assessment of their compatibility (Moorkamp et al., 2007) in different geological environments, such as the lithospheric mantle (Jones et al., 2009) or sedimentary environments (Carcione et al., 2007). Moorkamp et al. (2010), however, have been the only

ones to develop a joint inversion of electromagnetic (MT) and seismic (receiver function, surface-wave dispersion) observations.

Due to the non-linear nature of geophysical inverse problems, global optimization methods - including genetic algorithms - have been commonly considered (e.g. Everett and Shultz, 1993; Jervis et al., 1993; Pérez-Flores and Schultz, 2002; Moorkamp et al., 2010). Here, we assess the performance of a JI of MT, receiver functions and phase velocities to retrieve one dimensional conductivity and velocity structure. The datasets came from a temporary network of 11 collocated seismic and magnetotelluric stations centered at the Araripe Basin and extending in the NS direction into the Borborema Province and São Francisco craton. Both data sets were collected at exactly the same locations. The algorithm was developed by Moorkamp et al. (2010), and builds on the multi-objective, genetic algorithm approach (NSGA-II) of Deb et al. (2002). Before that, however, an investigation of the constraints that each data set is able to provide separately is performed. The constraints from the MT dataset were already investigated in Garcia et al. (2019) through a 3D inversion that provided a detailed resistivity cross-section down to asthenospheric depths for the study region; the constraints from the seismological dataset, which are part of this thesis, were published in Nemocon et al. (2021) and consisted of a joint inversion of receiver functions and surface-wave dispersion. This study provided detailed S-velocity models for the same region that allowed for a joint interpretation of seismic and resistivity structure.

Genetic algorithms (GA's) are a class of numerical optimization algorithms inspired by a biological analogy based on natural selection and genetics (Goldberg, 1989). Being random in nature they are considered a type of Monte Carlo technique. GA's encode the inversion parameters and perform a global search that avoids the potential trapping in a local minimum of linearized methods. The binary coding of parameters is motivated by the analogy to DNA in an evolutionary process and makes the two modification operators, crossover and mutation, independent from model parameter storage (Moorkamp, 2007).

Thus one of the main differences to other global optimization methods is that the algorithm does not work directly with model parameters (Goldberg, 1989) but a coded form of them. To produce the final models, GAs work through many generations of a given population that are subject to genetic operators (crossover or mutation), with the most adapted individuals being most likely to reproduce. Finally, a very well-adapted individual (the global solution) will appear in the offspring. NSGA-II in particular yields a set of final models based on Pareto-optimality, the so-called Pareto-optimal front, demonstrating to what extent fitting one data set trades off against fitting the other datasets (Moorkamp et al., 2010). When jointly inverting datasets that are sensitive to different parameters, we cannot be sure that these datasets sense common structures that can be described by a joint model. We therefore use the trade-off as an indicator of the compatibility among the data sets. These advantages come at the cost of being computationally more expensive than linearized methods (Moorkamp et al., 2010).

This thesis contains six chapters. In Chapter 1 (this chapter), I have presented the motivation for this research and the joint inversion of geophysical datasets. In Chapter 2, the geology and tectonic setting of the Borborema Province and the São Francisco craton are outlined. In Chapter 3, the datasets - including receiver functions, surface-wave dispersion velocities and MT observations - are described, and the theory behind the various methodologies utilized in this research are explained. In Chapter 4, a manuscript draft entitled "Lithospheric structure of the western Borborema Province from receiver functions and surface-wave dispersion: Implications for basin inversion", already published in *Tectonophysics*, is reproduced. In Chapter 5, a second manuscript draft entitled "Joint inversion of magnetotelluric and seismological observations: Imaging the electric Moho in NE Brazil", submitted to *Geophysical Journal International* for publication, is presented. Finally, in Chapter 6, the conclusions of this thesis, along with some considerations for future work, are stated.

Chapter 2

Geology and Tectonic setting

2.1 The Borborema Province

The Borborema Province (BP) is located in the northeastern corner of the South American continent, covering an area of $\sim 450,000 \text{ km}^2$. It is limited by the São Francisco Craton to the South, the Parnaíba Basin to the West, and a number of marginal basins to the North and East related to the opening of the Atlantic Ocean (Brito Neves et al., 1984). The Borborema Province was defined by Almeida et al. (1981) as a geological and structural domain that developed during the Brasiliano orogenic cycle at the end of the Neoproterozoic, which resulted in a complex orogenic system affected by deformational, metamorphic, and magmatic processes.

Its Precambrian basement (see Figure 2.1) consists of a complex mosaic of Archean to Paleoproterozoic metamorphic rocks (Oliveira and Medeiros, 2018) pervaded by a wide system of Late Proterozoic shear zones. Those shear zones mark the pre-rift tectonic framework of the Province, and likely developed along paleo-sutures that separated formerly independent tectonic blocks (Ganade de Araujo et al., 2013). The general trend of the basement structures is that of a fan opening towards the northeast, diagonal to the coast, that suffered deviations and inflections due to faulting movements (Almeida et al.,

1981). Some of the shear-zones that stand out in this assemblage due to their size are: The dextral Transbrasiliano-Kandi strike-slip belt, which formed as a result of relative movement obliquity during deformation imposed by NNE to NE escape in the Neoproterozoic; and the E-W trending Patos and Pernambuco dextral shear zones, formed during the Brasiliano orogeny at ~ 600 My and reactivated during the Pangea breakup in the Cretaceous, which divide the Province into Northern, Central and Southern domains (Figure 2.1).

According to Ganade de Araujo et al. (2013), the structuration of the Borborema Province occurred during the Brasiliano-Pan African orogeny. It started with an early collision associated with the closure of the large Pharusian-Goias Ocean between the Parnaíba block and the old basement of the Borborema Province at 620-600 My, followed by generalized crustal thickening and the development of eclogites and high-temperature thrusting foliation that defined the West Gondwana Orogen (Ganade de Araujo et al., 2013). The suture zone was subsequently reactivated by a set of dextral shear zones that led to the formation of the Transbrasiliano Lineament, acting as a transform plate boundary and allowing the closure of the restricted Sergipano Ocean and collision with the São Francisco craton at 590-580 My. Eastward and northward pushes resulting from those interactions gave rise to an extensive network of strike-slip shear zones across the Province that forced its northeastward extrusion (Ganade de Araujo et al., 2013).

This network of shear zones was partly reactivated in the Mesozoic, helping accommodate extensional stresses related to the opening of the South Atlantic ocean. Extension developed along three principal axes of deformation: Cariri-Potiguar, Recôncavo-Tucano-Jatobá, and Gabon-Sergipe-Alagoas (De Matos, 1992, 1999; Matos, 2021). According to De Matos (1992), this deformation started progressively along the Gabon-Sergipe-Alagoas and Recôncavo-Tucano-Jatobá to then continue West along the Cariri-Potiguar trend. However, the development of the Equatorial margin could have aborted the rifting process along the Cariri-Potiguar, allowing depositions of thick sedimentary

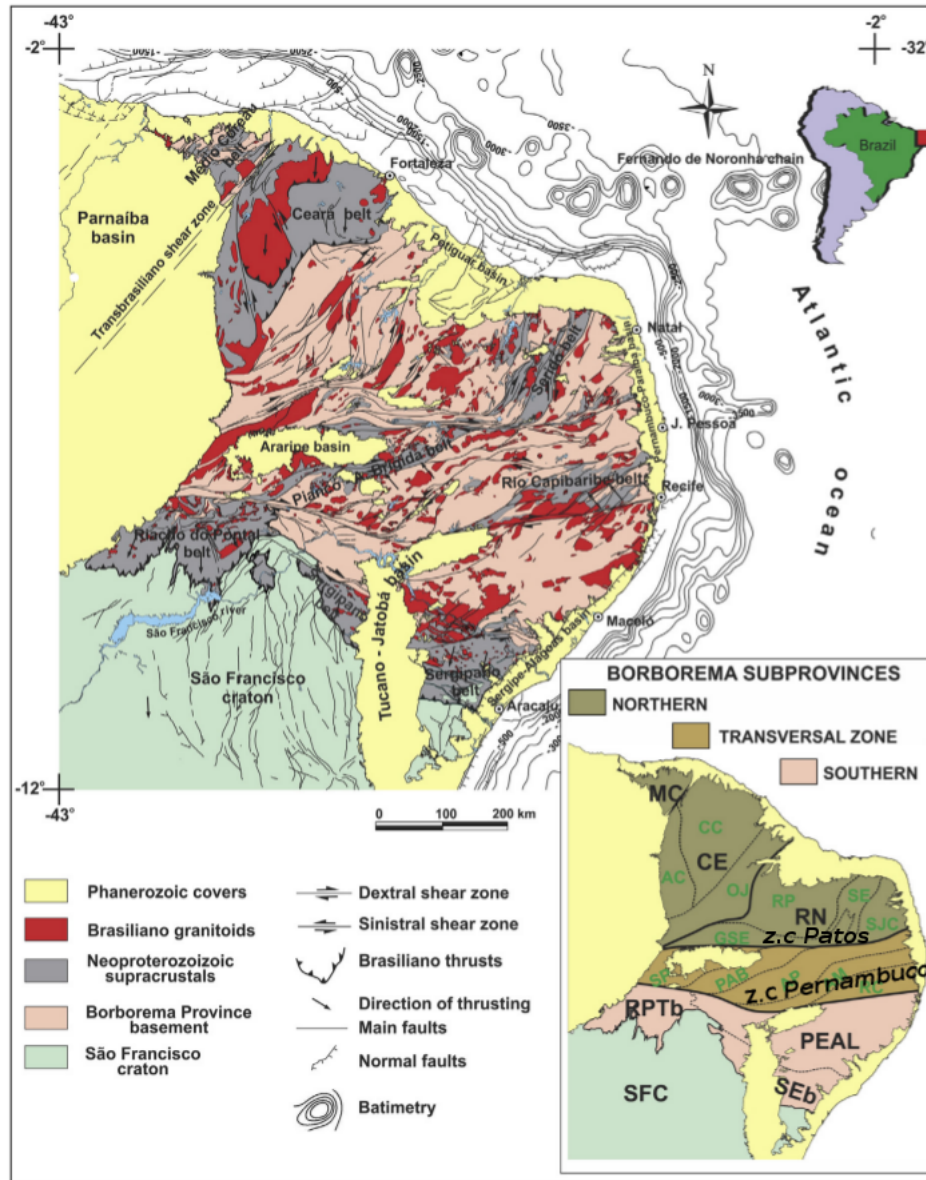


Figure 2.1: Geology and tectonic framework of the Borborema Province (BP) of Northeast Brazil. Shear zones and Brasiliano thrusts divide the Province in five domains: Médio Coreau (MC), Ceará (CE), Rio Grande do Norte (RN), Transversal Domain (TD) and Southern Domain (SD). The SD is divided into several subdomains, including Pernambuco-Alagoas (PEAL), Riacho do Pontal belt (RPTb), and Sergipano belt (SEb). Other terrains within the domains include: Ceará Central (CC), Acaraú (AC), Orós-Jaguaribe (OJ), Granjeiro-Seridó (GSE), Seridó (SE), Rio Piranhas (RP), São José de Campestre (SJC), Piancó-Alto Brígida (PAB), Alto Pajeú (AP), Alto Moxotó (AM), Rio Capibaribe (RC), and São Pedro (SP) (Oliveira and Medeiros, 2018).

sequences along the other two axis of deformation. Next, with the separation between the West Africa and São Luiz cratons, rifting along the Recôncavo-Tucano-Jatobá would have aborted also and rifting would have continued along the Gabon-Sergipe-Alagoas, leaving a number of marginal basins along the coast and intra-continental rift basins (now aborted) in the interior of the Province.

After continental breakup, episodes of intra-plate volcanism and uplift characterized the complex evolution of the Province. Volcanism occurred along two orthogonal magmatic alignments: Fernando de Noronha-Mecejana (EW, off-shore) and Macau-Queimadas Alignment (NS, on-shore). Its K-Ar ages are 26-34 My for Mecejana volcanism (Mizusaki et al., 2002) and 22-2 My for volcanic rocks at the Fernando de Noronha archipelago (e.g Knesel et al., 2011), suggesting a clear age progression from West to East along the alignment; K-Ar age ranges for Macau-Queimadas Alignment are 80 to 30 My (Mizusaki et al., 2002), expanding to 93-7 My when $^{40}\text{Ar}/^{39}\text{Ar}$ dates are considered (Knesel et al., 2011), displaying no clear age progression along the alignment. Moreover, the expanded range implies a time overlap of up to 15 Ma in magmatic activity along both alignments, as well as a temporal superposition with Cenozoic uplift in the Province (Knesel et al., 2011).

Uplift in the Borborema Province is represented by two main topographic elevations: the Borborema Plateau to the East (~ 1200 m a.m.s.l.), and the Chapada do Araripe to the West (~ 1000 m a.m.s.l.). Several studies in the Borborema Plateau have traditionally attributed its elevation to tectonic uplift resulting from either thermal doming (Usami et al., 1999; Jardim de Sá et al., 1999) or isostatic uplift (Morais Neto et al., 2009; Oliveira and de Medeiros, 2012). More recently, however, studies suggested it might be a high-standing, rheologically strong feature that resisted crustal stretching during Mesozoic extension (Almeida et al., 2015; Luz et al., 2015; Simões Neto et al., 2019). Regarding studies in the Chapada do Araripe, two main models have been put forward: (i) basin inversion originating from plate-wide horizontal compression (Marques et al., 2014), and

(ii) epeirogenic uplift resulting from regional crustal upwarp and subsequent differential erosion (Peulvast and Bétard, 2015).

2.2 The Araripe Basin

The Araripe basin is a large intra-continental rift basin that developed within the Cariri-Potiguar trend during Mesozoic extension. It is limited to the North by the EW-trending Patos Lineament and, although its general structure is made of NE-SW asymmetric grabens, it appears elongated in the E-W direction (Figure 2.2). Its internal architecture reveals the basin is structured into two sub-basins, Feira Nova to the West and Cariri to the East (de Castro and Branco, 1999), separated by a central horst (Matos 1992; 1999).

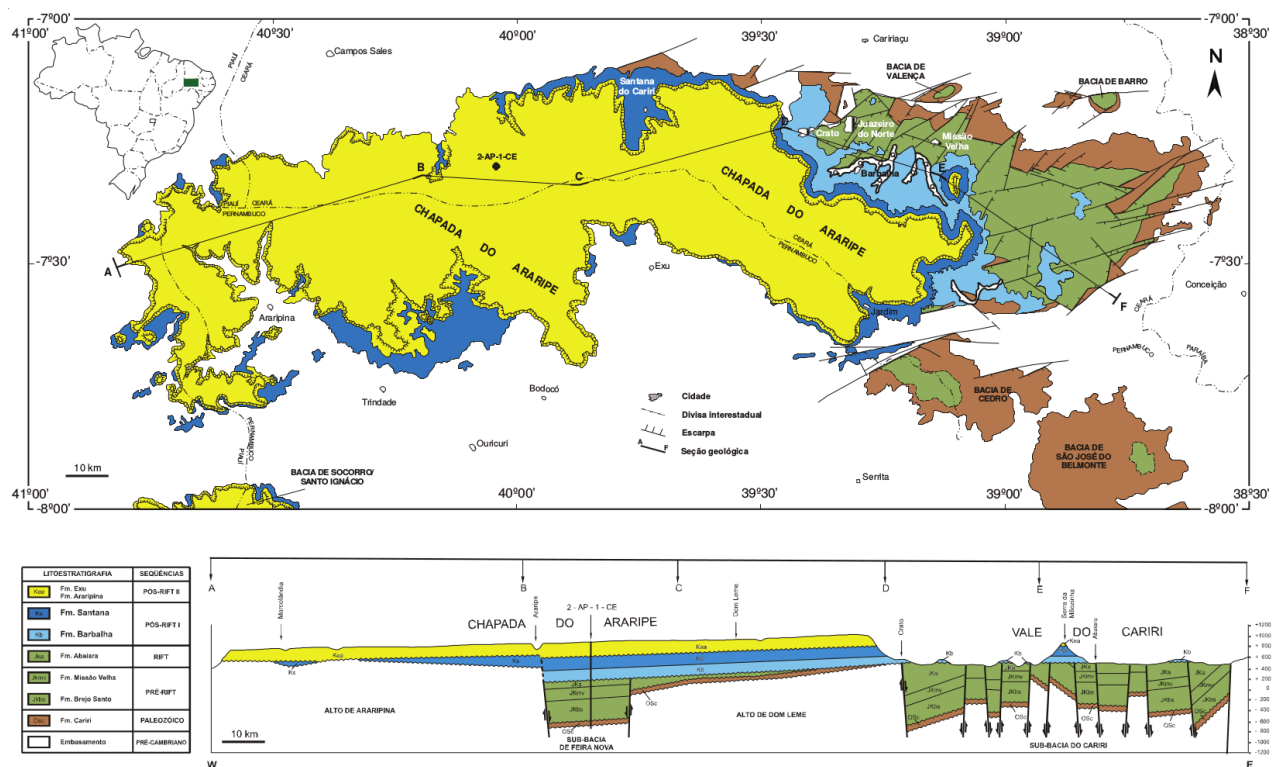


Figure 2.2: Geological map (up) and geologic cross-section (down) for the Araripe Basin. Note its subdivision into an elevated plateau (Chapada do Araripe), coinciding with the Exu formation, to the West and a depressed subbasin (Vale do Cariri) to the East. Also note its underlying architecture, consisting of a set of en-echelon NE-WNW hosts and grabens underlying the basin's sedimentary rocks (Assine, 2007).

The basin's stratigraphy is described by Hegarty et al. (2002) as: (i) a Paleozoic pre-rift sequence (Siluro-Ordovician) of fluvial and lacustrine deposits of Jurassic and Early Cretaceous age (Vale do Cariri Group), deposited on sandstones of the Paleozoic or Jurassic Mauriti or Cariri Formations (Assine, 2007); Fig. 2.3.), followed by a Late Jurassic-Early Cretaceous phase of regional subsidence and continental sedimentation of clay, siltites, sandstones, up to 600 m thick, (Coimbra et al., 2002; Valença et al., 2003) making the Brejo Santo and Missão Velha formations; (ii) a Neocomian rift sequence made of sandstones, siltstones, shales and thin carbonate clays deposited in a fluvial-lacustrine system composing the Abaiara formation; and, (iii) a Cretaceous post-rift sequence, consisting of three depositional systems: carbonaceous-fluvial-lacustrine facies deposited during Albian-Aptian time, evaporitic-lagoonal and marine facies deposited during middle Albian time, and a braided to meandering fluvial system during Albian-Cenomanian time, corresponding to the Araripe group (see Figure 2.3). This stratigraphic transgressive-regressive sequence is made of fluvial, lacustrine, lagoonal and marine sediments of Aptian to Cenomanian age. It is divided into several formations: the Rio da Batateira (or Barbalha: Assine (2007)), Santana (Crato, Ipubi, Romualdo Members), Arajara and Exu Formations. The crystalline basement is composed of gneiss-migmatitic units and granitoids, and metamorphic rocks deformed during the assembly of Gondwana in the Brasiliano (750-540 My), which comprise mainly phyllites, mica schists, quartzites, metavolcanics, banded iron formations, metacarbonates, paragneisses, metaconglomerates, and marbles (e.g. Gomes, 1981).

The most conspicuous feature associated with the inversion is a large mesa capped by alluvial sediments of the so-called Exu Formation (150 – 250 m in thickness), unconformably covering the older layers (Fig. 2.2). According to Assine (2007), it consists of two distinct lithological units separated by an erosional unconformity or syn-sedimentary deformation. The upper unit is the Exu Formation, whereas the lower unit, of middle Albian age, is the Araripina Formation. The latter is restricted to the western part of the

CHRONO-STRATIGRAPHY		LITHOLOGY		LITHOSTRATIGRAPHIC UNITS		TECTONIC PHASES		
		Thickness						
94 Myr	EARLY CRETACEOUS	Albian		150 to 200 m	sandstones	EXU Fm.	ARARIPE GROUP	POST-RIFT
100				20 m	clayey sandstones	ARAJARA Fm.		
112			 	200 m	marls	Romualdo Mb.		
		gypsum			Ipubi Mb.			
			calciferous shales	Crato Mb.				
	Late Aptian		60 to 200 m	coarse and clayey sandstones	RIO DA BATATEIRA Fm.			
118	LATE JURASSIC	Neocom.		120 m	clayey sandstones	ABAIARA Fm.	CARIRI GROUP	SYN-RIFT
138				200 to 250 m	coarse sandstones	MISSÃO VELHA Fm.		
141		Tithon.		400 m	shales	Upper Mb.	BREJO SANTO Fm.	
	Oxford./Kimeridgian		clayey sandstones		Lower Mb.			
165	SILURIAN?			100 m	sandstones	MAURITI Fm.		

Figure 2.3: Stratigraphic column for the Araripe Basin and related tectonic phases, compiled from Ponte and Ponte (1996) and Neumann (oral comm., 2008). Note the two large uncoformities separating the syn-rift from the post-rift phases, and the pre-rift from a postulated Silurian deposition.

basin and consists of rhythmites where fine-grained, reddish laminated sandstone with plurimetric lenses of coarser sandstone with load structures and cross-bedding indicate a floodplain environment. Where present, this basal layer ensures a strong contrast in mechanical resistance to erosion between the sedimentary cover and the crystalline rocks of

the basement.

Finally, a regional unconformity separates the Abaiara formation (Neocomian rift sequence) from the Rio da Batateira formation (Albian-Aptian post-rift sequence). This hiatus represents a 20 Ma gap and may be related to tectonically-controlled regional crustal uplift (Ponte, 1996). Also, due to a combination of thermal subsidence and global eustatic events, the post-rift series were deposited throughout the Araripe Basin and beyond its present limits. The post-rift basin first formed in the northeast, where the Albian to Cenomanian series are more complete, and then extended to the south and west, progressively lapping onto the basement (Figure 2.2).

2.3 The São Francisco Craton

The São Francisco Craton is located in the east-central part of Brazil, covering more than one billion square kilometers (Figure 2.1). It is limited by the Borborema Province and Parnaíba basin to the North, the Atlantic ocean to the east, and a number of Brazilian Belts that are the remnant of a cratonic bridge that once linked the São Francisco and Congo blocks. According to Alkmim et al. (2001) the basement of the São Francisco Craton consists of an Archean nucleus and two Paleoproterozoic orogens, hosting sedimentary sequences deposited between the Paleo-Archean (~ 3300 Ma) and Late Neoproterozoic (~ 580 Ma).

Specific to our study region is the northern portion of the craton, where 5 stations of the N-S profile were deployed (see Figure 3.3) between the Archean basement and the Proterozoic cover. The Proterozoic cover is contained within a tectonic domain known as the Paramirim aulacogen, which comprises two morphotectonic domains: the sedimentary rocks of the Espinhaço sequence to the north and, a little more to the south, a large plateau known as the Chapada Diamantina (Alkmim and Martins-Neto, 2012). The Espinhaço sequence provides evidence of a long subsidence history during the Protero-

zoic (Danderfer et al., 2009), while the Chapada Diamantina provides evidence of partial tectonic inversion during the Neoproterozoic Brasiliano event. Furthermore, a system of double-verging and NNW-trending faults and folds were developed in the craton (Cruz and Alkmim, 2006).

Chapter 3

Data and Methods

3.1 Data Sets

The dataset utilized in this work was acquired by the Universidade Federal do Rio Grande do Norte as part of the BORborema Deep Electromagnetic and Seismic (BODES) experiment, a temporary array of collocated seismic and magnetotelluric stations sampling the western Borborema Province and northern São Francisco craton. The acquisition was part of a broader research project funded by the Conselho de Desenvolvimento Científico e Tecnológico (CNPq) to develop a new joint inversion scheme for seismic and electromagnetic datasets. In the following, we describe each deployment separately. The seismic dataset was complemented by additional stations deployed in the Araripe Basin under a project funded by the national oil company Petrobras, which aimed at investigating the architecture of aborted syn-rift basins in NE Brazil.

3.1.1 Seismic deployment

The seismic component of the collocated deployment consisted of nine broadband stations borrowed from the *Pool de Equipamentos Geofísicos do Brasil* (PEGBR) and two permanent broadband stations belonging to the *Rede Sismográfica do Nordeste* (RSISNE).

The combined network of 11 stations formed an approximately 600 km-long linear profile trending in the NS direction, with an interstation spacing of ~ 60 km (see Figure 3.1). Two short-period stations, also belonging to the PEGBR, were additionally deployed on top of the Araripe basin between two broadband stations.

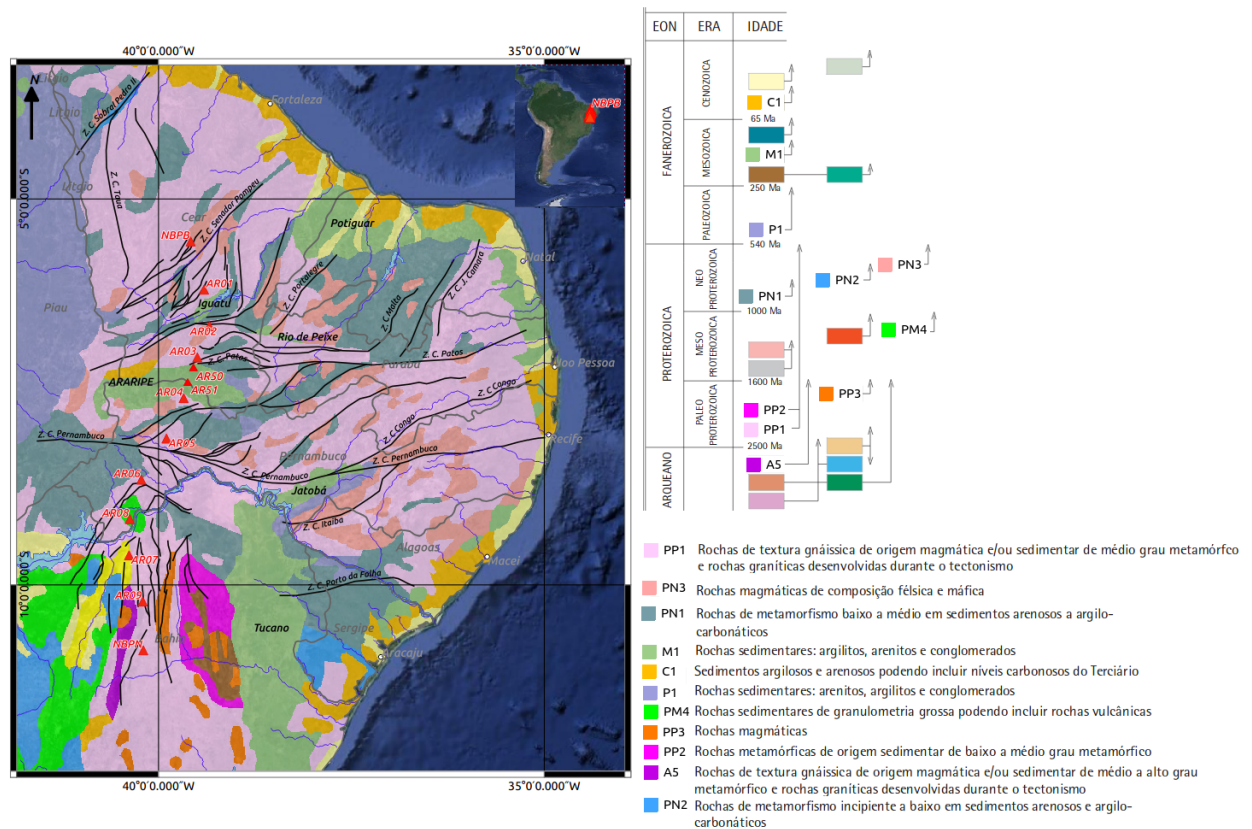


Figure 3.1: Geologic map of the Borborema Province with the seismic stations of the BODES Experiment (red triangles) and the Precambrian network of shear zones (solid lines) superimposed.

The broadband stations were equipped with RefTek 151-120A 'Observer' sensors, with flat response in velocity in the 0.008-50 Hz frequency band, feeding 24-bit RefTek RT-130 digitizers sampling at 100 Hz. The short-period stations consisted of Sercel L4A-3D sensors, with flat velocity response above 2 Hz, also feeding RT-130 digitizers and also sampling at 100 Hz. Global Positioning System antennas were used for all stations for timekeeping. The broadband network was initially deployed in January 2015 with 8 sites (AR01 - AR08), and in April 2016 the equipment in site AR01 was moved South to

site AR09 to complete the line. The deployment was dismantled in March 2017, yielding an effective time of operation of a bit over 2 years for the central stations (AR02 - AR08) and about 1 year for the two remaining stations (AR01, AR09). The permanent RSISNE stations were in operation since 2013. The two short-period stations were deployed between November 2015 and May 2017 (AR50), and between August 2016 and May 2017 (AR51), respectively. Location coordinates are listed in Table 1.

After October 2018, three additional stations were deployed on top of the Araripe Basin, as part of a broader study - funded by the national oil company Petrobras - to investigate the crustal structure under the syn-rift basins of NE Brazil. The new deployment consisted of two short-period (PEGBR) stations and one broadband station complementing an existing RSISNE station to the East. The combined deployment formed a ~ 170 km-long EW trending line that complemented the longer NS line mentioned above (see Figure 3.3). The newly installed broadband station belongs to the Laboratório Sismológico (LabSis/UFRN) and consisted of a GeoTech broadband sensor (120 s) and a high-gain SMART-24 digitizer sampling at 100 Hz. Again, GPS antennas were used for timekeeping at all stations. Station coordinates are listed in Table 1.

3.1.2 Magnetotelluric deployment

The magnetotelluric (MT) component of the experiment consisted in 11 long-period stations that were installed in May and October of 2015, at the same locations as the seismic instruments. Each of those MT stations, borrowed from the *Pool de Equipamentos Geofísicos do Brasil* (PEGBR), included a three-axis magnetometer (FluxGate), and five non-polarizable Cu-CuSO₄ electrodes. Those receivers were connected to a LEMI-417M digitizer sampling at 1 Hz, which stores the time series for the seven channels: three magnetic and four electric fields. The LEMI instruments include a GPS receiver that provides satellite synchronization to the internal clock and records the coordinates. Because the MT stations were inter-spaced at about ~ 60 km and given the decay of lateral sensitivity,

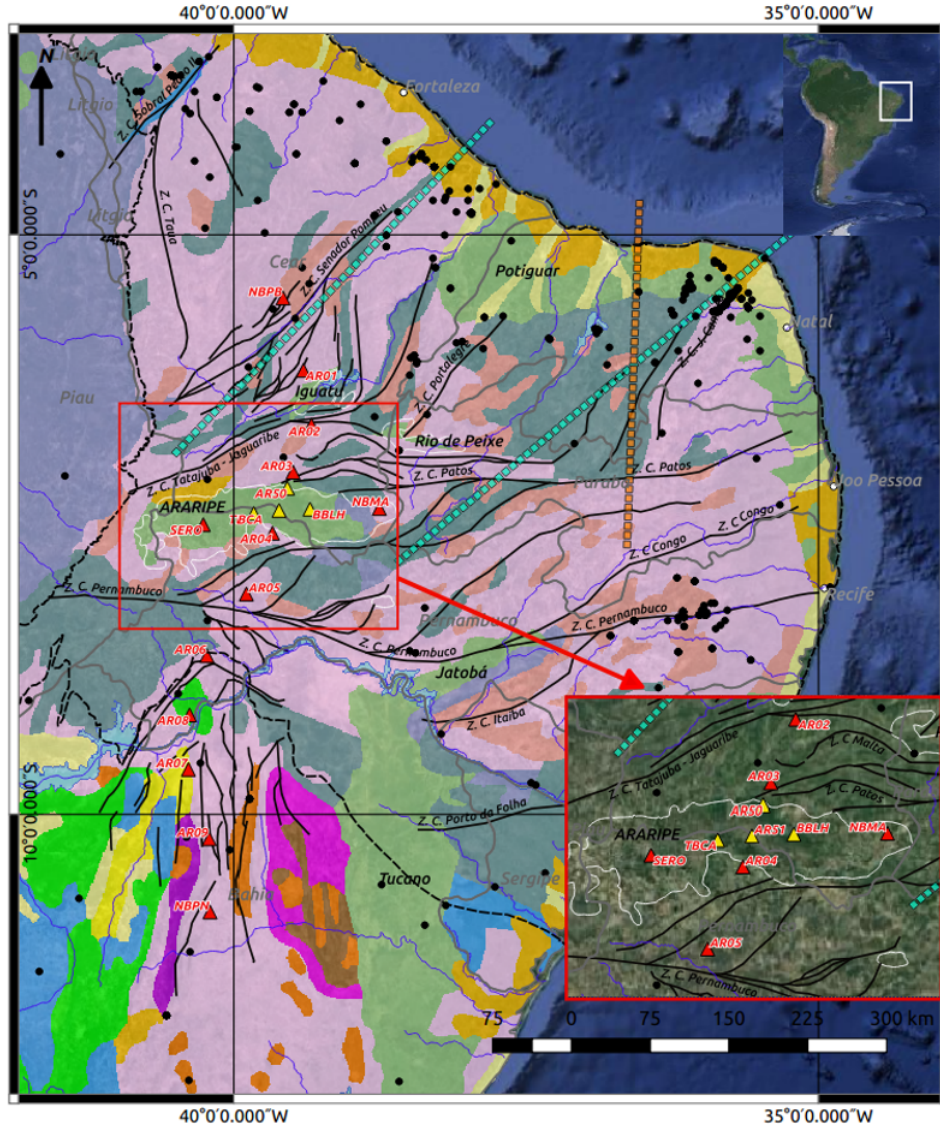


Figure 3.2: Same legend as in Figure 3.1 with the new seismic stations of the EW line. The short-period stations (yellow triangles) and broadband station (red triangles).

a poor constraint of structures in the upper crust was expected, other than the ones located directly underneath the instruments (Garcia et al., 2019). Instruments recorded between 21 and 34 days, which allow us to obtain robust estimates of the resistivity distribution up to LAB depths.

STATION	LAT	LON	m a.s.l.
NBPB	-5.5432	-39.5836	260
AR01	-6.1682	-39.4098	305
AR02	-6.6419	-39.3402	341
AR03	-7.0466	-39.4958	479
AR50	-7.1957	-39.5463	896
AR51	-7.3861	-39.6255	890
AR04	-7.5758	-39.6746	493
AR05	-8.0932	-39.8983	441
AR06	-8.6305	-40.2249	454
AR08	-9.1450	-40.3727	394
AR07	-9.6154	-40.3837	405
AR09	-10.2136	-40.1954	678
NBPN	-10.8468	-40.1988	387
SERO	-7.5000	-40.2555	749
TBCA	-7.4090	-39.8395	861
BBLH	-7.3672	-39.3447	924
NBMA	-7.3650	-38.7640	437

Table 3.1: Station coordinates.

3.1.3 Receiver functions

Receiver functions are time series developed from the coda of teleseismic P-wave recordings (Langston, 1979). Seismic sources are typically selected for epicentral distances between 30° and 90° , for any depth, and with magnitude greater than 5.5 mb. This epicentral distance range avoids interference with upper mantle transition zone triplications and the shadow zone created by the outter core. Following this criteria, a total of 3185 seismic sources were selected (see Figure 3.4) available for processing in this study. Along the NS profile, the most of the stations recorded over 150 waveforms, except stations AR01 (143 waveforms), AR09 (55 waveforms), AR50 (120 waveforms), and AR51 (69 waveforms), recorded less data. Whereas along the EW profile were exactly: SERO (142 waveforms), TBCA (166 waveforms), BBLH (168 waveforms) and NBMA (173 waveforms).

Next, the selected seismograms were rotated around the corresponding vertical component (V) and into the great-circle path to obtain the radial (R) and transverse (T) components. Following Langston (1979), the theoretical displacement response (D_i) for a plane P-wave can be expressed as a convolution of three filters: the instrument impulse

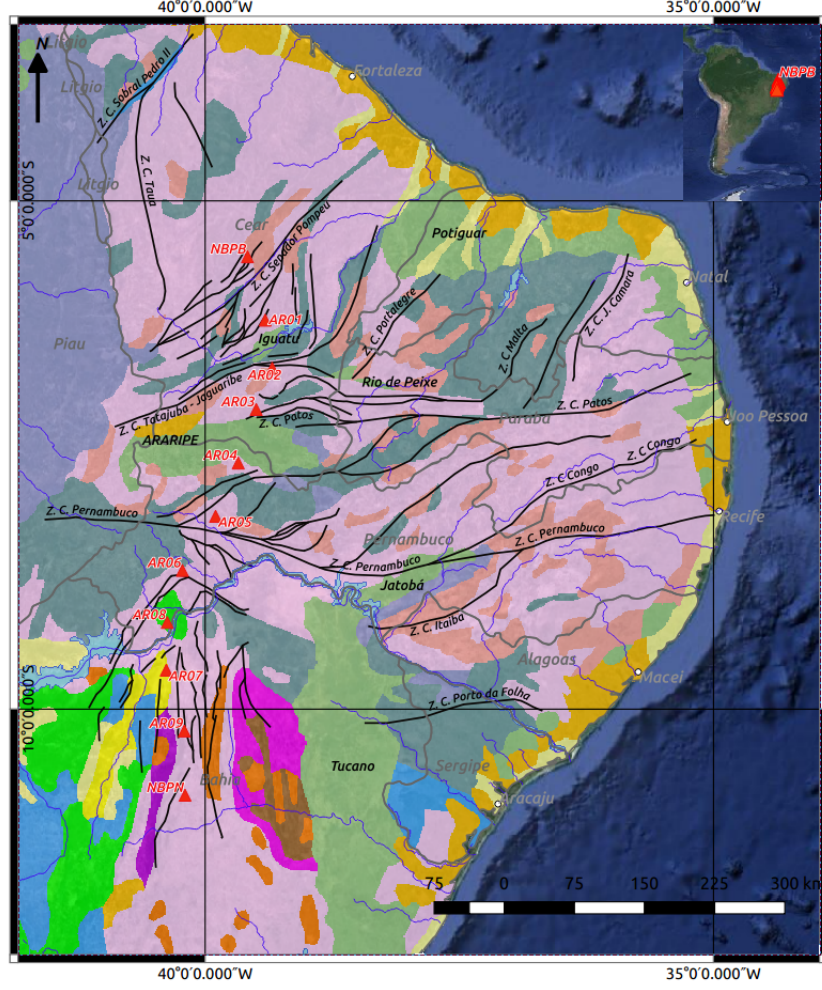


Figure 3.3: Same legend as in Figure 3.1, for magnetotelluric stations of the BODES Experiment (red triangles).

response ($I(t)$), the teleseismic source-time function ($S(t)$), and the Green's function related to propagation local to the receiver ($E(t)$). Mathematically, this is expressed as

$$\begin{aligned}
 D_V(t) &= I(t) * S(t) * E_V(t), \\
 D_R(t) &= I(t) * S(t) * E_R(t), \\
 D_T(t) &= I(t) * S(t) * E_T(t),
 \end{aligned} \tag{3.1}$$

where the $*$ denotes convolution. Receiver functions are then obtained from the deconvolution of the vertical component from the horizontal components. In the frequency

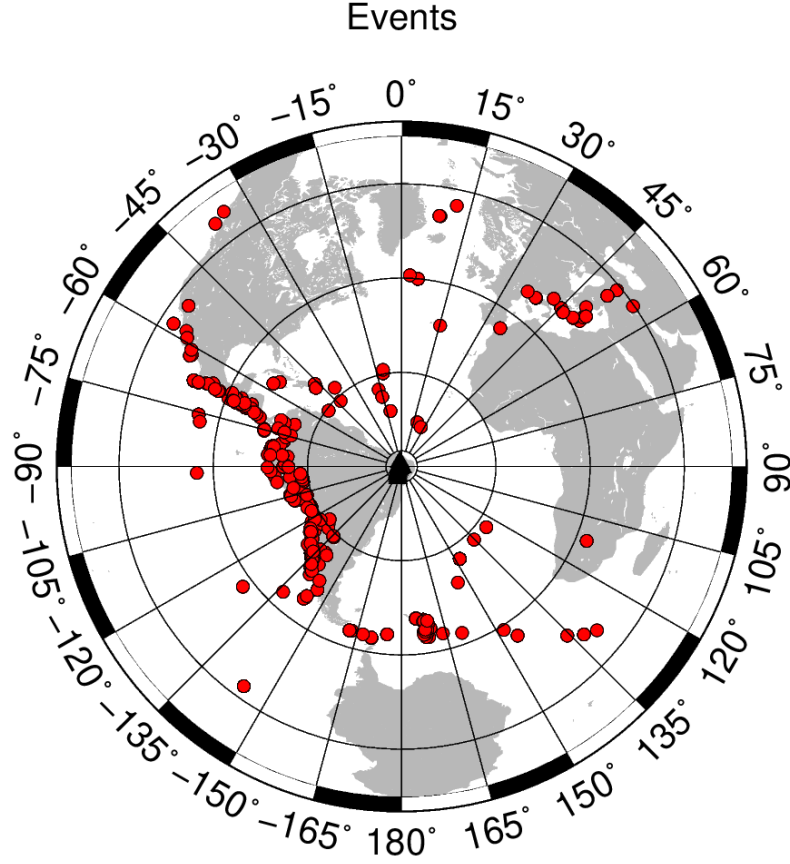


Figure 3.4: Global map seismic sources selected for this study (red circles) and stations of the BODES Experiment (black triangles).

domain, the deconvolution operation is expressed as

$$E_R(\omega) = \frac{D_R(\omega)}{I(\omega)S(\omega)} \simeq \frac{D_R(\omega)}{D_V(\omega)},$$

$$E_T(\omega) = \frac{D_T(\omega)}{I(\omega)S(\omega)} \simeq \frac{D_T(\omega)}{D_V(\omega)},$$
(3.2)

where $E_R(\omega)$ and $E_T(\omega)$ are the radial and transverse receiver functions, respectively.

The deconvolution process removes the signature of the instrument response and source time function from the resulting time series, effectively equalizing the waveforms (Langston, 1979). Radial receiver functions consist of a leading P-wave at zero lag-time followed by a number of P-to-S converted waves resulting from the interaction of

the leading P-wave with sub-surface discontinuities. On the other hand, transverse receiver functions for one-dimensional plane layered structures should not have any energy recorded. However, in practice, some energy on the transverse component is recorded due to anisotropy, dipping interfaces and/or complex 3D structures (e.g. Levin and Park, 1997; Bianchi et al., 2010). The resulting time series can then be understood as a linear combination of P-to-S converted waves that reverberate in the structure beneath the seismometer (Figure 3.5).

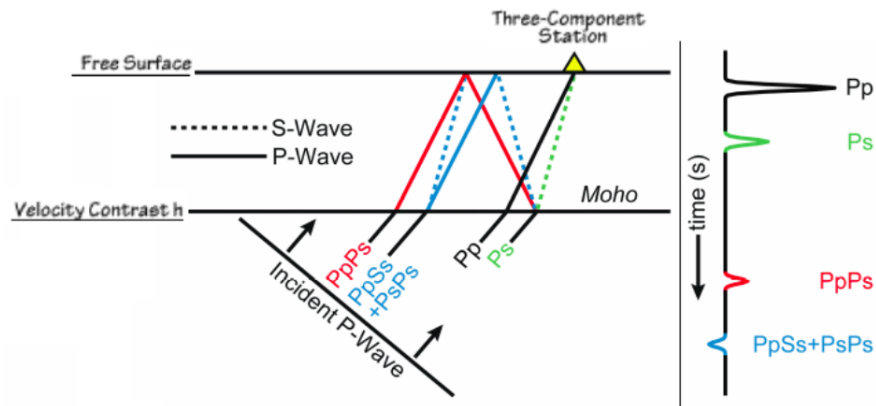


Figure 3.5: Receiver function (right) calculated for a layer-over-half space model. (left) Ray paths are indicated in the model with their conversions labelled in the receiver function. The largest amplitude (black) is observed at time 0 and is related to leading P-wave arrival (Pp). The next arrival (green) at ~ 4.5 s is related to the P-to-S conversion from the Moho interface. The first multiple reflection from the surface (red) is related to the PpPs conversion. The second multiple reflection (blue) corresponds to the PpSs + PsPs conversions; its amplitude is slightly smaller than the first multiple and its polarity is negative. (Ammon, 1991; Ammon et al., 1990)

Receiver function amplitudes depend on the incidence angle of the impinging P-wave and the velocity contrasts generating the conversions (Ps) and multiples (PsPs, PpSs, PsPs). The arrival times of the converted phase and multiples depend on the depth of the velocity contrast, the P and S velocity between the contrast and the surface, and the P-wave incidence angle, or ray parameter. Analysis of the S-P travel-times and amplitudes in receiver functions is then utilized to reveal the S-velocity structure under the recording station (e.g. Owens et al., 1984; Ammon et al., 1990).

Before the deconvolution process is carried out some additional pre-processing steps must be taken:

1. The teleseismic waveforms are windowed 10 s before and 110 s after the P-wave, demeaned, detrended and tapered.
2. A high-pass filter above 0.025 Hz is applied to remove long-period noise.
3. A low-pass filter below 8 Hz is applied to avoid aliasing.
4. Recordings are decimated to 10 s.p.s.

Then, receiver function estimation is realized through the iterative, time-domain deconvolution approach using the procedure of Ligorria and Ammon (1999), with 500 iterations. This approach is a least-squares minimization of the difference between the observed horizontal seismogram and a predicted signal, which consists of a linear combination of spikes convolved with the vertical component. In each iteration the vertical component is cross-correlated with the radial component to obtain the lag of the first and largest spike in the receiver function. After, the spike amplitude is estimated and its convolution with the vertical component is subtracted from radial component, the procedure is repeated with the next iteration to estimate other spike lags and their amplitudes. For each additional spike estimated, the misfit between the vertical component and the convolution of the receiver function with the radial component is reduced, and the iteration is stopped when the reduction in misfit is negligible.

Finally, the resulting time series are low-pass filtered with Gaussian filters of width $\Delta f = 1.0$ and 2.5 , which produce receiver function estimates at overlapping frequency bands of $f < 0.5$ and $f < 1.2$ Hz. The quality control applied to the receiver functions start by rejecting those that do not reproduce at least 85% of the original radial waveform when convolved back with the vertical component. Next, the remaining waveforms are visually inspected and those exhibiting unstable behaviour and/or exceedingly large amplitudes in

the transverse component are excluded. Besides, radial receiver functions are superimposed to identify outliers, which are also excluded. This quality control was carried out on 17 seismic stations, obtaining 513 receiver functions at high-frequency contents ($a=2.5$) and 580 receiver functions at low-frequency contents ($a=1.0$).

To improve the signal-to-noise ratio, a stacking that depends on the ray parameter with variations < 0.001 s/km in ray parameter and variations $< 8^\circ$ back-azimuth is realized. Sample radial and transverse receiver functions for station NBPB are displayed in Figure 3.6. We can see a large peak at zero lag time, corresponding to the direct P-wave, followed by the Moho conversion peak at lag time of about 5 s (Ps) and the first reverberation (PpPs) at 15 s lag time. On the other hand, the transverse receiver functions have relatively small amplitudes, which are suggestive of a medium under this station that can be approximated as isotropic and laterally homogeneous.

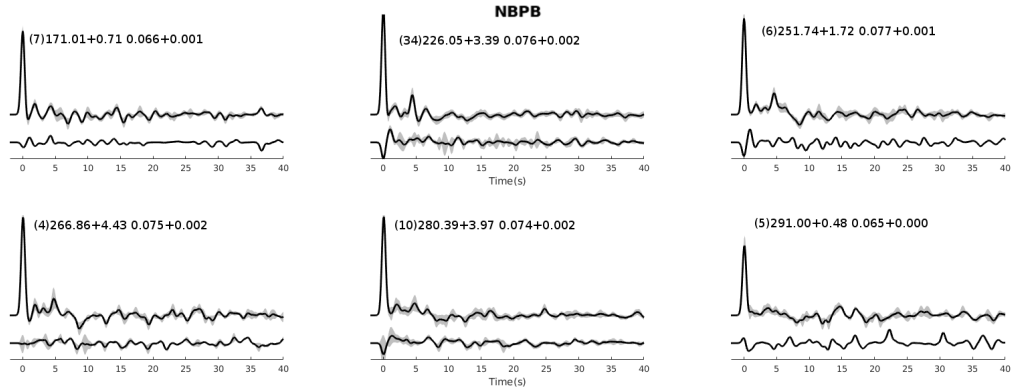


Figure 3.6: Sample receiver function stacks for station NBPB. Top and bottom traces refer to radial and transverse receiver functions, respectively. The number of waveforms included in the stack, the average back-azimuth and the average ray parameter are displayed at top each panel. The stacks are presented as a black line within a grey shade representing 1σ -confidence bounds for the stacked amplitudes.

3.1.4 Surface-wave dispersion

Dispersion measurements consist of phase and group velocities for the fundamental-mode of Rayleigh waves. Group velocities were included in the joint-inversion of receiver functions and surface-wave dispersion (Chapter 4) from two independent surface-wave tomography studies by Luz et al. (2015) and Dias et al. (2015). Phase velocities were taken from a regionalized processing approach and were included in our second joint-inversion with MT data (Chapter 5).

Long-period group velocities were taken from a continental-scale tomographic study by Luz et al. (2015). They built on the surface-wave tomography study of Feng et al. (2004), who obtained Rayleigh-wave group velocity estimates for periods between 10 and 140 s, at 10 s period intervals, using the multiple filter analysis of Dziewonski et al. (1969) after isolation of the fundamental mode with a phase-match filter (Herrin and Gofforth, 1977). The dispersion velocities were inverted tomographically with the procedure of Feng et al. (2007) using a grid of $1.0^\circ \times 1.0^\circ$ tomographic cells and covering a square area enclosing the entire continent. Short-period group velocities were obtained from the ambient noise tomography study of Dias et al. (2015). Using stacked cross correlations of 1 month of seismic ambient noise recorded in the vertical component of broadband stations in the Borborema Province and surrounding regions, they developed a total of 333 empirical Green's functions from which group velocities were measured. They built maps of group velocity variation for the fundamental-mode Rayleigh wave over a grid of $0.5^\circ \times 0.5^\circ$ tomographic cells using the Fast Marching Surface-wave Tomography (FMST) scheme of Rawlinson and Sambridge (2005). The model was parameterized with 551 velocity nodes, and 2D group-velocity maps were produced for periods of 5, 10, 15, and 20 s.

Rayleigh-wave phase velocity were obtained through the array processing approach of McMechan and Yedlin (1981), as implemented in the Computer Programs in Seismology package of Herrmann and Ammon (2002), and correspond to an average dispersion curve

for the medium under the recording network. This approach is based on the behavior of two waves that are in phase at a certain frequency, interfering constructively when they are in phase. From an array of seismic traces are realized two transformations: a first transformation from the distance-time (x, t) domain into the ray parameter-time intercept (p, τ) domain, through slant-stack integration; and a second transformation into the (p, ω) domain through 1D Fourier integration over the τ variable, which effectively yields the phase velocity dispersion relation through its equivalence in the (c, T) domain (where $p = 1/c$ and $\omega = 2\pi/T$).

Seismic sources from a total of 18 events with magnitudes between 5.0 and 5.9 Mw and epicentral distances ranging from 1050 to 3412 km, located along the South American subduction zone and Mid-Atlantic ridge, were selected. In general data preparation consisted in four steps: 1) demean, detrend and taper; 2) low-pass filter under 1.0 Hz; 3) window the waveforms between 400 and 1600 s; and 4) resample to 0.5 seconds. Rayleigh-wave phase velocities were thus obtained for periods in the 12 - 80 s period range, with velocities between 3.39 - 3.93 km/s.

3.2 Methods

3.2.1 The magnetotelluric method

The magnetotelluric method (MT) is a geophysical tool for imaging the conductivity (or resistivity) distribution in the Earth's crust and upper mantle. It is defined as a passive exploration technique that uses a broad spectrum of naturally occurring geomagnetic variations as a power source for electromagnetic induction in the Earth (see Figure 3.7). Using simultaneous measurements of the horizontal components of the electric $\mathbf{E}$ and magnetic $\mathbf{H}$ fields recorded at the surface, it estimates the complex impedance tensor $\mathbf{Z}$ in the frequency domain.

The MT method is based on electromagnetic wave propagation that can be explained

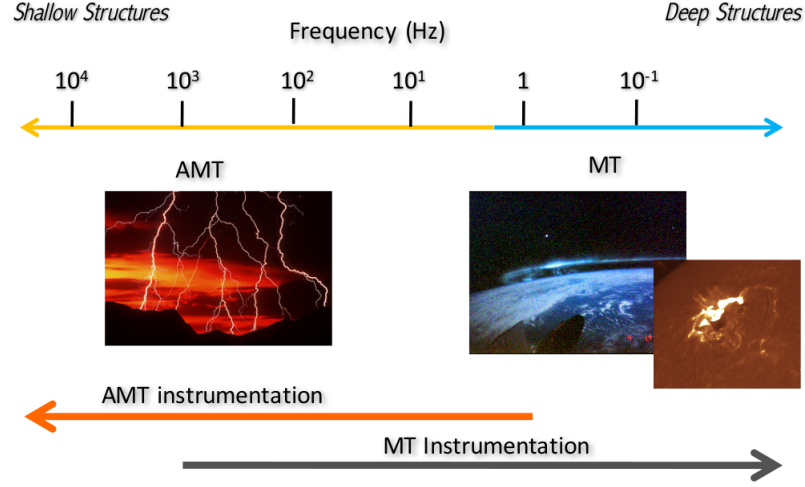


Figure 3.7: Ranges of frequencies used by AMT and MT instrumentations. Typical MT frequencies range from 0.0005 Hz to 10^{-5} Hz. Shorter-period signals (>8 Hz) have their origin in worldwide thunderstorm activity, and specifically their associated lightning discharges, radiating EM fields that propagate to great distances and are recorded at Audio magnetotelluric (AMT) stations. AMT instrumentation is a multifrequency, narrow-band, and its depth of investigation is used for upper crustal studies. At the transition at about 1 Hz the amplitude of the signal is very low and is referred to as the MT dead-band.

using Maxwell's equations:

$$\nabla \times \mathbf{E} = -\frac{\delta \mathbf{B}}{\delta t}, \quad (3.3)$$

$$\nabla \times \mathbf{H} = \mathbf{j} + \frac{\delta \mathbf{D}}{\delta t}, \quad (3.4)$$

$$\nabla \cdot \mathbf{B} = 0, \quad (3.5)$$

$$\nabla \cdot \mathbf{D} = \mathbf{q}, \quad (3.6)$$

where $\mathbf{E}$ [V/m] is the electric field, $\mathbf{B}$ [T] is the magnetic induction, $\mathbf{H}$ [A/m] is the magnetic field, $\mathbf{j}$ [A/m²] is the electric current density, $\mathbf{D}$ [C/m²] is the electric displacement, and $\mathbf{q}$ [C/m³] is the charge density. Some assumptions have to be considered next in order to obtain an expression for the electromagnetic induction in the Earth, which is the MT or quasi-static approximation:

1. Assume a planar, horizontally impinging magnetic source field. All natural electromagnetic sources are generated by large-scale ionospheric current systems, which are relatively far from the Earth's surface and can then be treated as uniform, plane-polarized electromagnetic waves impinging on the Earth's surface at near-vertical incidence.

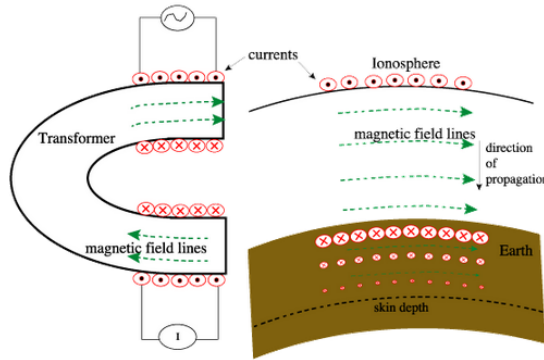


Figure 3.8: For terrestrial conductivities the source waves reaches Earth's surface at a near vertical angle and may be considered plane electromagnetic waves.

2. The charge is conserved and the Earth becomes an ohmic conductor, implying that current density $\mathbf{j}$ and electric field $\mathbf{E}$ are linearly related through the scalar electrical conductivity σ ,

$$\mathbf{j} = \sigma \mathbf{E}. \quad (3.7)$$

3. Assuming a linear relationships between $\mathbf{D}$ and $\mathbf{E}$; and $\mathbf{B}$ and $\mathbf{H}$ applies to all Earth materials, neglecting variations in electrical permittivity (ϵ) and magnetic permeability (μ). Thus:

$$\mathbf{D} = \epsilon_0 \mathbf{E}, \quad (3.8)$$

$$\mathbf{B} = \mu_0 \mathbf{H}. \quad (3.9)$$

4. Assuming that this is a diffusive process, where the Earth does not generate energy but only dissipates and absorbs it, the electric displacement field is quasi static for MT sounding periods (100.000-0.0001 Hz). Then, given typical resistivities of $10^6 - 10^{-2} \Omega\text{m}$ for the Earth, the right-hand terms in equation (3.4) verifies that,

$$\frac{\delta \mathbf{D} / \delta t}{\mathbf{j}} = \frac{i\omega\epsilon_0}{\sigma} \simeq 0.1 - 10^{-19}, \quad (3.10)$$

therefore, displacement currents can be neglected when compared to conductivity currents since ($\sigma \gg \omega\epsilon_0$).

Taking those assumptions into Maxwell's equations we obtain

$$\nabla \times \mathbf{E} = -\frac{\delta \mathbf{B}}{\delta t}, \quad (3.11)$$

$$\nabla \times \mathbf{B} = \mu_0 \sigma \mathbf{E}, \quad (3.12)$$

$$\nabla \cdot \mathbf{B} = 0, \quad (3.13)$$

$$\nabla \cdot \mathbf{E} = \frac{\mathbf{q}}{\epsilon_0}, \quad (3.14)$$

or, in the frequency domain

$$\nabla \times \mathbf{E} = -i\omega \mathbf{B}, \quad (3.15)$$

$$\nabla \times \mathbf{B} = \mu_0 \sigma \mathbf{E}, \quad (3.16)$$

$$\nabla \cdot \mathbf{B} = 0, \quad (3.17)$$

$$\nabla \cdot \mathbf{E} = -\mathbf{E} \nabla \ln \sigma. \quad (3.18)$$

Applying the curl functional to eqs (3.15) and (3.16), we obtain

$$\nabla^2 \mathbf{E} = -i\omega\mu_0\sigma\mathbf{E} - \nabla(\mathbf{E}\nabla\ln\sigma), \quad (3.19)$$

$$\nabla^2 \mathbf{B} = i\omega\mu_0\sigma\mathbf{B} + \mu_0\mathbf{E} \times \nabla\sigma. \quad (3.20)$$

Since the electric and magnetic fields are non-orthogonal, we can introduce the impedance tensor (Z), which is the transfer function between the electric and magnetic fields as

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}, \quad (3.21)$$

and the magnetic field transfer function or tipper vector, which defines the relationship between the vertical and horizontal magnetic field as

$$H_z = \begin{pmatrix} A & B \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}. \quad (3.22)$$

The impedance tensor is alternatively expressed as amplitude (apparent resistivities, ρ_{ij}) and phase (ϕ_{ij}):

$$\rho_{ij} = \frac{1}{\omega\mu_0} |Z_{ij}(\omega)|^2, \quad \phi_{ij} = \tan^{-1} \left(\frac{\Im Z_{ij}}{\Re Z_{ij}} \right), \quad (3.23)$$

where ω is angular frequency, μ is magnetic permeability, and the sub-indexes i and j indicate the impedance tensor elements. For a homogeneous half-space (Zhdanov, 2009; Cagniard, 1953), it can be shown that

$$Z = (1 + i) \sqrt{\frac{\omega\mu_0}{2\sigma}} \Rightarrow \phi = \tan^{-1} 1 = \frac{\pi}{4}. \quad (3.24)$$

Usually, apparent resistivity and phase are used to interpret magnetotelluric data, due to their intuitive physical relationship to the properties of the Earth (e.g. Cavaliere and Jones,

1984).

The MT processing of the time series is explained in (Garcia et al., 2019), using a non-stationary robust remote reference regression algorithm (Neukirch and Garcia, 2014):

1. Decompose the time series using a multivariate empirical mode decomposition (MEMD).
2. Estimate instantaneous parameters, including the instantaneous frequency defined as the time derivative of the phase.
3. Gather independent data points, ensuring linear independency of contiguous frequencies.
4. Organize data in the frequency domain. The spectra are collected in narrow bins, typically 5 to 10 bins per decade, ensuring enough estimates per decade and statistical stability of the impedance estimation for each bin by exploiting the fact that the MT transfer functions vary slowly with frequency.
5. Estimate transfer functions: (i) robust least-squares regression of the normal equation using a remote reference:

$$e = b \cdot z, \quad (3.25)$$

and (ii) estimate confidence intervals by bootstrapping the spectral estimates.

3.2.2 H- κ stacking

The H- κ stacking technique of Zhu and Kanamori (2000) estimates basic properties of the crust, such as thickness (H) and bulk V_p/V_s ratio (κ), by use of the moveout of the secondary phases generated at the Moho.

The difference between the arrival times of the direct P-wave and their conversions into S-wave (generated at the Moho) can be used to estimate crustal thickness through the equation

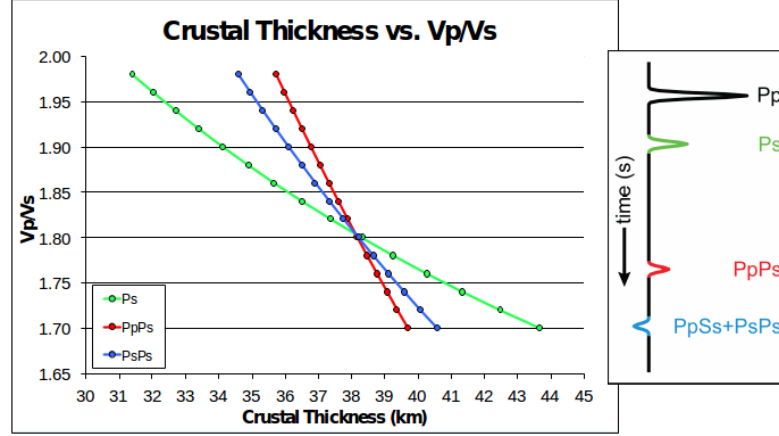


Figure 3.9: H, κ relations (left panel) for the different phases (right panel) generated at the Moho discontinuity. Note, the convergence point between the contributions of each phases.

$$H = \frac{t_{Ps}}{\sqrt{V_s^{-2} - p^2} - \sqrt{V_p^{-2} - p^2}}, \quad (3.26)$$

where the ray parameter p can be obtained from the incidence angle and V_P and V_S are given by the Earth model. The dependence of H on V_p is not as strong as on V_s , implying that small changes in κ can lead to big changes in H . Hence, Zhu and Kanamori (2000) included later phases to constrain crustal thickness,

$$H = \frac{t_{PpPs}}{\sqrt{V_s^{-2} + p^2} - \sqrt{V_p^{-2} - p^2}}, \quad (3.27)$$

$$H = \frac{t_{PpPs+PsPs}}{2\sqrt{V_s^{-2} - p^2}}. \quad (3.28)$$

In practice, background noise and scattering make the identification of the three phases (Moho Ps and crustal multiples PpPs and PpSs+PsPs) more complicated. To increase the signal/noise ratio, the algorithm includes a stacking of the receiver function amplitudes

along the predicted phase-moveout curves for the three phases, removing the need to identify phases and pick arrival times. Through a grid search over H and κ , a stacking surface is built by summing a weighted combination of Ps, PpPs and PpSs + PsPs amplitudes from individual receiver functions,

$$S(H, \kappa) = w_1 \times r(t_1) + w_2 \times r(t_2) + w_3 \times r(t_3), \quad (3.29)$$

where $r(t)$ is the radial receiver function, t_1 , t_2 and t_3 are the predicted Ps, PpPs and PsPs + PpSs arrival times for a given crustal thickness H and V_p/V_s ratio κ , and w_i are *a priori* weighting factors. The best estimates of H and κ are found when the three phases are stacked coherently along the phase-moveout curves (Zhu and Kanamori, 2000).

The approach requires the assumption of a P-wave velocity, which was taken as 6.5 km/s from the global average for continental areas (Christensen and Mooney, 1995). It also requires *a priori* weights for the three crustal phases (Ps, PpPs, PpSs+PsPs). Following the criterion of Julià and Mejia (2004), we assign values of 0.4, 0.3, and 0.3, respectively, when all three phases are clearly observed; values of 0.5, 0.5, 0.0, when the second crustal multiple is not observed; and values of 0.5, 0.0, 0.5, when the first multiple is not observed.

To estimate the confidence bounds for each parameter, we used the *bootstrap* methodology proposed by Efron and Tibshirani (1991). This methodology (see Figure 3.10.) consists of generating new sets of traces from an original receiver function set for a given station through random re-sampling. These new sets are created during each iteration and have the same size as the initial set, but the receiver functions can be replicated and/or replaced. For each re-sampled set of traces, estimates of H and κ are developed from application of the H- κ stacking procedure. The standard deviation for H and κ are then obtained as

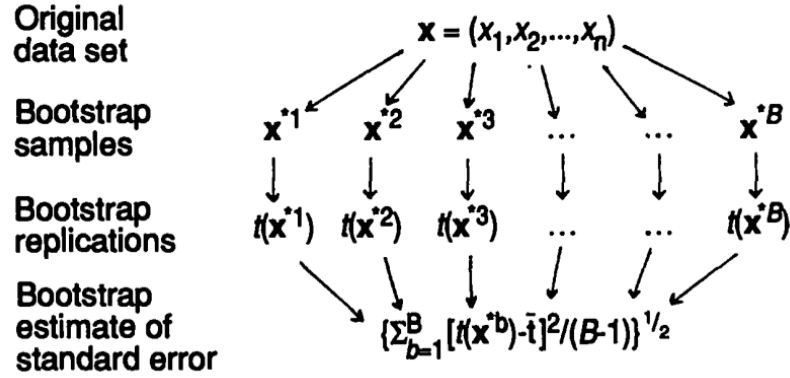


Figure 3.10: Bootstrap algorithm for estimating the standard deviation of a statistic $t(x)$; each of the B bootstrap samples is a random sample of size n ; $\bar{t}$ is the average of the B bootstrap replications $t(\mathbf{x}^{*b})$ (Efron and Tibshirani, 1991).

$$\sigma(\bar{x}) = \sqrt{\frac{\sum_{i=1}^n (x_i - \bar{x})^2}{n(n-1)}}, \quad (3.30)$$

where $\bar{x}$ is the thickness or V_P/V_S average, the x_i are the individual estimates from a re-sampled set of traces, and n is the number of replications.

An example illustrating the performance of the H- κ stacking procedure at station AR02 is given in Figure 3.11. The figure shows the H- κ stacking surface in the left panel and the receiver function waveforms, sorted by ray parameter, in the right panel. A single local maximum is observed on the H- κ stacking surface, displaying crustal thicknesses of 34.6 ± 0.4 km and a V_P/V_S ratio of 1.75 ± 0.02 . Confidence bounds were obtained after bootstrapping the receiver function dataset with 200 replications.

3.2.3 Common Conversion Point (CCP) stacking

The Common Conversion Point (CCP) migration and stacking of Frassetto et al. (2010) is a procedure that combines the CCP stacking of Gilbert and Sheehan (2004) with the phase-weighting scheme of Schimmel and Paulssen (1997). Its main purpose is to create a three-dimensional structural image of the crust and upper mantle, taking advantage of

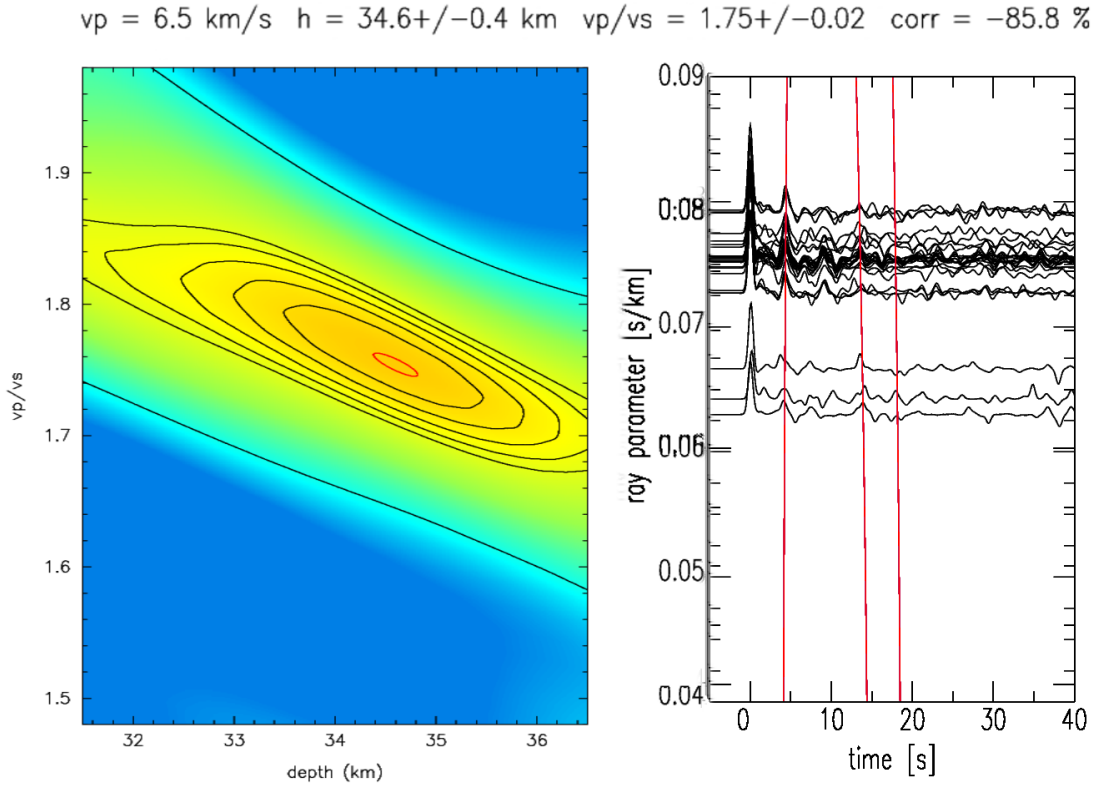


Figure 3.11: Sample H- κ stacking results for station AR02. The right panel shows the corresponding radial receiver functions used during the stacking procedure, sorted by ray parameter. The thin red lines are the phase moveout curves for the Ps, PpPs and PpSs + PsPs phases associated with the maximum in the H- κ stacking surface. The left panel shows the H- κ stacking surface. The thick red line is the 1σ confidence ellipse from bootstrapping.

the lateral coherence between receivers, and produce reflection-style images of discontinuities in the crust and mantle, rather than point estimates (Gilbert and Sheehan, 2004).

The CCP approach of Gilbert and Sheehan (2004) determines lateral variations in depth to seismic discontinuities. This procedure first sorts receiver functions into common conversion point bins. Next, receiver function amplitudes whose ray paths pierce the discontinuities at short distances from one another are binned and stacked, hoping that bins combining data from multiple stations improve the signal-to-noise ratio (Gilbert and Sheehan, 2004). To achieve this goal it is necessary to migrate the receiver functions from the time domain to the depth domain through

$$T_{Ps_i}(p_i) = dz_i \cdot \left(\sum_{i=0}^{z-1} \sqrt{V_{s_i}^{-2} + p_i^2} - \sqrt{V_{p_i}^{-2} - p_i^2} \right), \quad (3.31)$$

where T_{Ps} is the arrival time for Ps phase (Moho), i is current layer and z the number layers, p is the ray parameter, V_p and V_s are P and S velocities, respectively.

Next, a back-projection or Pre-stack Depth Migration (PSDM) is applied. This is achieved through use of the following equations

$$\sin \theta_i = p_i \cdot V_{s_i}, \quad (3.32)$$

$$\theta_i = \sin^{-1}(p_i V_{s_i}), \quad (3.33)$$

where θ_i is the incidence angle for the Ps phase at the i -th interface and for

$$\beta_i = \sin^{-1}(p_i V_{p_i}), \quad (3.34)$$

where β_i is the incidence angle for the P -wave at the i -th interface. The migration results are associated with a grid of bins. Each *bin* has 1 km thickness, bin radii of 100 km, and spacing of 50 km.

The phase-weighting scheme uses the phase stack as a weight, thus acting as a filter of coherent arrivals. The CCP average was originally obtained as

$$a(z) = \frac{1}{N} \sum_{j=1}^N w_j s_j(z), \quad (3.35)$$

where w_j is the distance weight attributed to each receiver function amplitude s_j , depending on the depth z , and N is the number of piercing points (Gilbert and Sheehan, 2004). The phase-weighted CCP averages are additionally weighted with an instantaneous phase (ϕ_j) filter

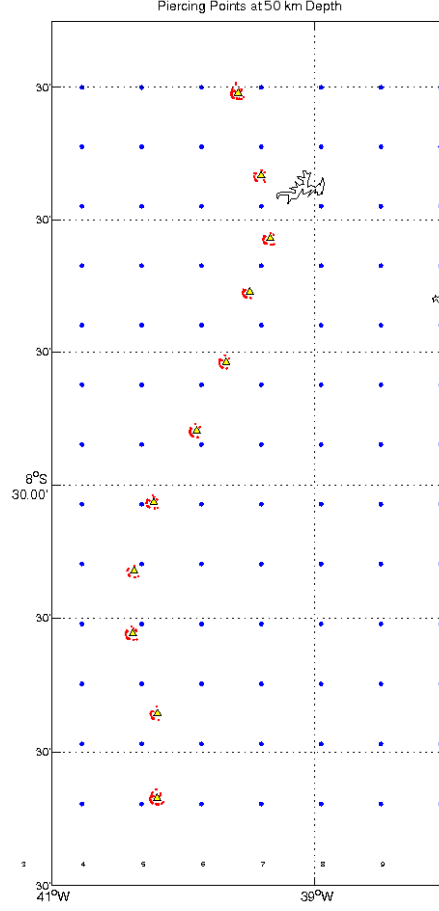


Figure 3.12: Distribution bins-grid (blue). Piercing points (red) are the geographical locations of the P-to-S conversions, which are tabulated for each depth and used to define a grid of uniformly spaced nodes with bin radii enclosing a minimum of 5 piercing points. The stations corresponds to the triangles (yellow).

$$c(z) = \frac{1}{N} \left| \sum_{j=1}^N \exp^{i\phi_j(z)} \right|, \quad (3.36)$$

and the phase-weighted CCP is defined as

$$g(z) = a(z)c(z)^v, \quad (3.37)$$

where v defines the sharpness of the filter (Schimmel and Paulssen, 1997), which is commonly selected as $v = 0.5$.

3.2.4 Joint inversion of receiver functions and dispersion

The joint inversion approach of Julià et al. (2000; 2003) builds on the iterative scheme of Ammon et al. (1990) for the inversion of receiver functions. In that approach, the root-mean-square (RMS) misfit between observed and predicted receiver functions is minimized in a least-square sense, with smoothness constraints on the inverted velocity-depth profile.

The forward problem can be expressed as

$$\mathbf{d} = \mathbf{F}[\mathbf{m}], \quad (3.38)$$

where $\mathbf{d}$ is an N -dimensional vector of data points, $\mathbf{F}[\cdot]$ is a non-linear operator that maps vectors in the model space into vectors in the data space, and $\mathbf{m}$ is an M -dimensional vector that contains shear wave velocities for fixed-thickness layers at depth. Expanding eq. (3.38) in a Taylor series

$$\mathbf{d} \simeq \mathbf{F}(\mathbf{m}_0) + \nabla \mathbf{F}|_{\mathbf{m}_0} \cdot [\mathbf{m} - \mathbf{m}_0] + \frac{1}{2} \nabla^2 \mathbf{F}|_{\mathbf{m}_0} \cdot [\mathbf{m} - \mathbf{m}_0]^2 + \dots, \quad (3.39)$$

so,

$$\mathbf{d} \simeq \mathbf{F}(\mathbf{m}_0) + \nabla \mathbf{F}|_{\mathbf{m}_0} \cdot [\mathbf{m} - \mathbf{m}_0] + \mathbf{O}[\mathbf{m} - \mathbf{m}_0]^2, \quad (3.40)$$

where $\mathbf{m}_0$ is the starting model and $\mathbf{O}$ represents the error of the expansion. To make the dependence of the final model with respect to the initial model more explicit,

$$\mathbf{d} - \mathbf{F}(\mathbf{m}_0) + \nabla \mathbf{F}|_{\mathbf{m}_0} \cdot \mathbf{m}_0 = \nabla \mathbf{F}|_{\mathbf{m}_0} \cdot \mathbf{m}, \quad (3.41)$$

or, alternatively

$$\mathbf{d}' = \nabla \mathbf{F}|_{\mathbf{m}_0} \cdot \mathbf{m}, \quad (3.42)$$

where $\mathbf{d}'$ is the modified vector of residuals. The solution is then obtained in a linearized and iterative fashion as

$$\mathbf{m}_{i+1} = (\nabla \mathbf{F}|_{\mathbf{m}_i})^{-1} \cdot \mathbf{d}'. \quad (3.43)$$

The inverse operator $(\nabla \mathbf{F}|_{m_i})^{-1} = \mathbf{G}^{-1}$ is obtained by minimizing the functional with L_2 -norm

$$\varphi = ||\Delta \mathbf{d} - \nabla \mathbf{F}|_{m_0} \cdot \Delta m||^2, \quad (3.44)$$

and with a smoothness constraint

$$\varphi = (\Delta \mathbf{d} - G\Delta m)^T (\Delta \mathbf{d} - G\Delta m) + \theta((\Delta m)^T A^T A(\Delta m) - G\Delta m), \quad (3.45)$$

where A corresponds to an approximation by finite differences of the second differences between shear velocity perturbations in adjacent layers, expressed as

$$A = \begin{bmatrix} 1 & -2 & 1 & 0 & \dots \\ 0 & 1 & -2 & 1 & \dots \\ 0 & 0 & 1 & -2 & \dots \\ 0 & 0 & 0 & 1 & \dots \\ 0 & 0 & 0 & \dots & 1 \end{bmatrix} \quad (3.46)$$

and θ is an adjustable parameter that controls the trade-off between fitting the waveform and model roughness. The smoothness parameter is computed through *L-curve* analysis.

In the joint inversion scheme, the RMS portion of the functional is altered to account for the minimization of two independent datasets: receiver functions and dispersion velocities. An influence factor p ($0 \leq p \leq 1$) is introduced that controls the trade-off between

fitting the receiver functions and the dispersion curves. Following Julià et al. (2000) the altered RMS portion of the misfit function is

$$E_{y|z} = \frac{p}{N_y} \sum_{i=1}^{N_y} \left(\frac{y_i - \sum_{j=1}^M Y_{ij} x_j}{\sigma_{yi}} \right)^2 + \frac{1-p}{N_z} \sum_{i=1}^{N_z} \left(\frac{z_i - \sum_{j=1}^M Z_{ij} x_j}{\sigma_{zi}} \right)^2, \quad (3.47)$$

where the first term accounts for the misfit in the dispersion curves and the second for the misfit in the receiver functions. N_y , N_z are the number of data points for each data set, while σ_{yi} , σ_{zi} are the standard deviations, Y_{ij} are the partial derivate matrices and y_i is the residual vector for dispersion curves, Z_{ij} are the partial derivate matrices and z_i is the residual vector for receiver functions.

The system of equations to be inverted, following Julià et al. (2003) is

$$\begin{bmatrix} \sqrt{\frac{p}{w_s^2}} D_s \\ \sqrt{\frac{q}{w_b^2}} D_b \\ \theta A \\ W \end{bmatrix} \vec{m} = \begin{bmatrix} \sqrt{\frac{p}{w_s^2}} \vec{r}_s \\ \sqrt{\frac{q}{w_b^2}} \vec{r}_b \\ \vec{0} \\ \vec{0} \end{bmatrix} + \begin{bmatrix} \sqrt{\frac{p}{w_s^2}} D_s \\ \sqrt{\frac{q}{w_b^2}} D_b \\ 0 \\ 0 \end{bmatrix} \vec{m}_0 + \begin{bmatrix} 0 \\ 0 \\ 0 \\ W \end{bmatrix} \vec{m}_a, \quad (3.48)$$

where the sub-indices s and b refer to the dispersion and receiver function estimates, respectively. D_s and D_b are the partial derivate matrices, $\vec{r}_s$ and $\vec{r}_b$ are the vectors of residuals, w_s^2 and w_b^2 are weights that equalize the data sets, contains the velocities of fixed-thickness layers overlying a half-space, and $\vec{m}_0$ is the initial model. The matrix A builds the second differences between adjacent model layers, and W is the diagonal matrix of weights for the *a-priori* velocity values in $\vec{m}_a$.

The joint inversion procedure is illustrated for station AR04 in Figures 3.13 and 3.14. The match between observations and predictions (Figure 3.14) is good for both receiver function waveforms and dispersion curves.

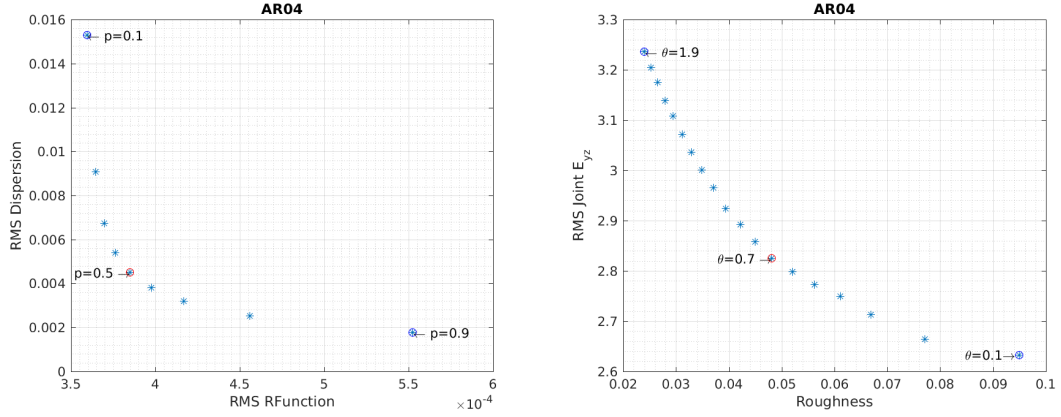


Figure 3.13: Trade-off curves for station AR04, with optimal influence p and smoothness θ parameters indicated.

3.2.5 Joint inversion of MT and seismological data

The joint inversion of MT and seismological datasets utilized here was developed by Moorkamp et al. (2010), and builds on the multi-objective, genetic algorithm approach (NSGA-II) of Deb et al. (2002). Genetic algorithms (GA's) are a class of Monte Carlo techniques based on a biological analogy (Goldberg, 1989). Although this idea of using a population of solutions to solve practical engineering optimization problems was considered since the 1950's, GAs were invented by John Holland in the 1960's (Coley., 1999). Due to the non-linear nature of geophysical inverse problems, a large variety of genetic algorithms have been applied to various types of datasets (e.g. Moorkamp et al., 2010; Pérez-Flores and Schultz, 2002; Everett and Shultz, 1993; Jervis et al., 1993). GAs work through encoded parameters, which solve the problem by performing a global search that avoids the local search of linearized methods. The earliest GAs encoded the parameters into finite-length strings of alphabets with certain cardinality (Goldberg, 1989). However, a binary representation has gained popularity due to the analogy with DNA in evolutionary processes. Additionally, it makes the two modification operators, crossover and mutation, independent from model parameter storage (Moorkamp, 2007).

The basic genetic algorithm process flow includes:

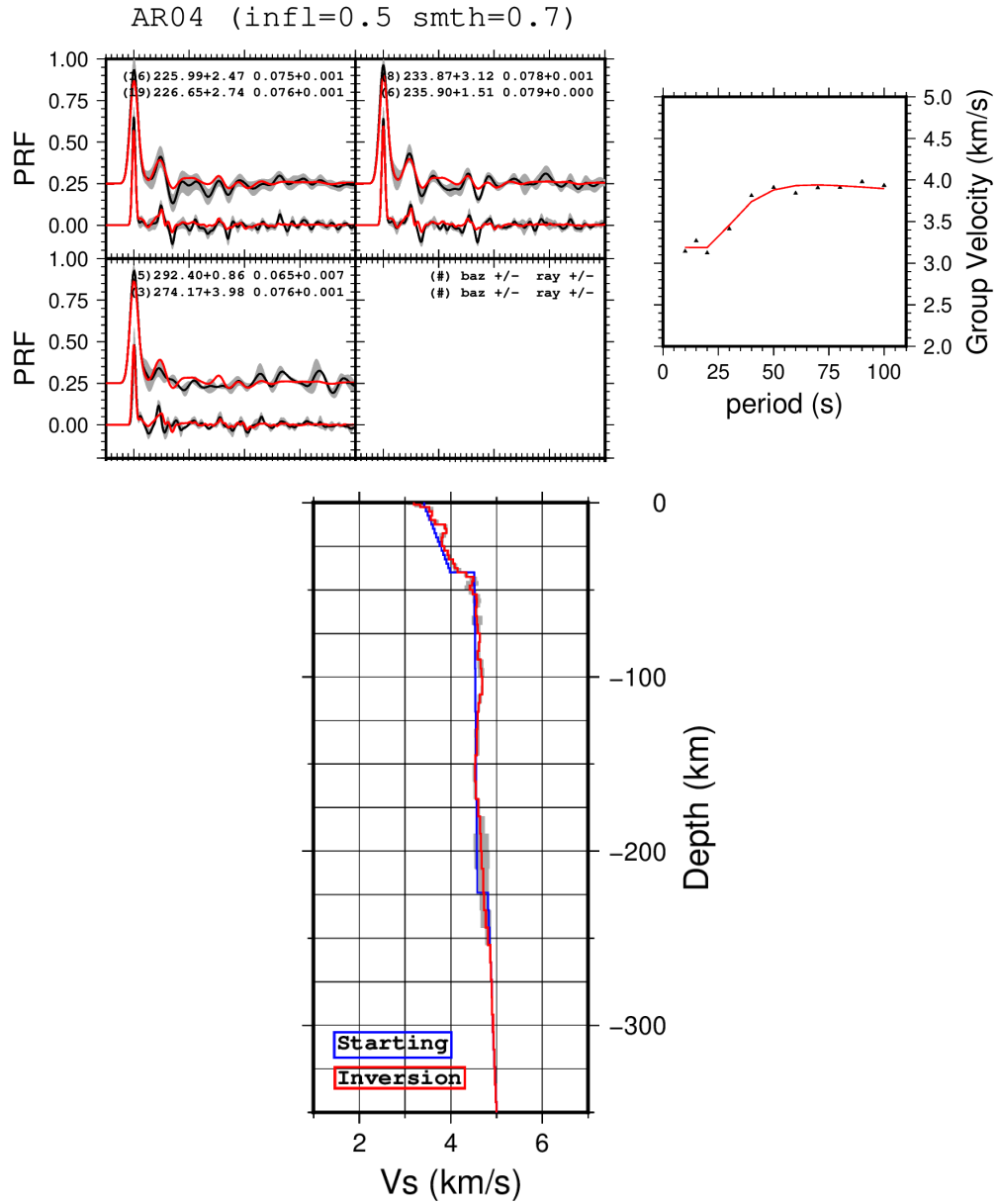


Figure 3.14: Detailed joint inversion results for AR04. The panels display: a comparison between observed (black) and predicted (red) receiver functions at both high ($f < 1.25$ Hz) and low ($f < 0.5$ Hz) frequency contents and for different ray parameter and back azimuth (the number of receiver functions, average back azimuth and average ray parameter are displayed on top of each receiver function panel); and next right panel, a comparison between observed (symbols) and predicted (red line) Rayleigh wave, group velocities. The bottom panel, displays the starting (blue) and inverted (red) velocity model, with the 2σ confidence bounds (grey shade).

1. Choose parameters.
2. Generate an initial population of individuals.
3. Evaluate the fitness of all individuals.

While termination condition not met **do**

Select fitter individuals for reproduction,

Recombine between individuals,

Mutate individuals,

Evaluate the fitness of the modified individuals and

Generate a new population.

End while

Each individual is a solution to the problem, which is characterized by a set of parameters (variables) known as 'genes'. Genes are joined into a string to form a 'chromosome' (solution), which is represented in terms of an alphabet. Usually, binary values are used (string of 1s and 0s) and we say that we encode the genes in a chromosome (see Figure 3.15).

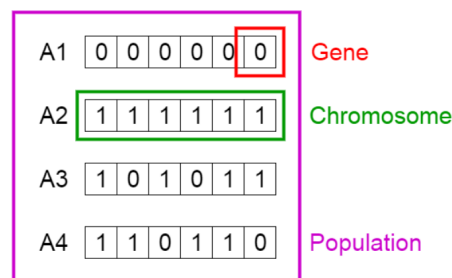


Figure 3.15: Genes: set of parameters to solve a problem. Chromosomes: a solution. Population: Size of candidate solutions.

An important concept of the notion of population is its size, which is usually specified by the user and is one of the important factors affecting the scalability and performance of

genetic algorithms. Thus, small population sizes might lead to premature convergence and low-quality solutions. In contrast, large population sizes lead to unnecessary expenditure of valuable computational time (Goldberg, 1989).

In the selection, two pairs of individuals (parents) are selected based on their fitness scores. Individuals with high fitness have more chance to be selected for reproduction. In the recombination or crossover, for each pair of parents selected, a crossover point is chosen at random from within the genes (for example, crossover point is 3 in Figure 3.16). Offspring are created by exchanging the genes of the parents among themselves until the crossover point is reached and new offspring are added to the population. Additionally, some of their genes in the newly formed offspring can be subject to mutation with low probability. In practice, this implies that some of the bits in the bit string can be flipped (see Figure 3.17). Mutation is implemented to maintain diversity within the population and prevent premature convergence.

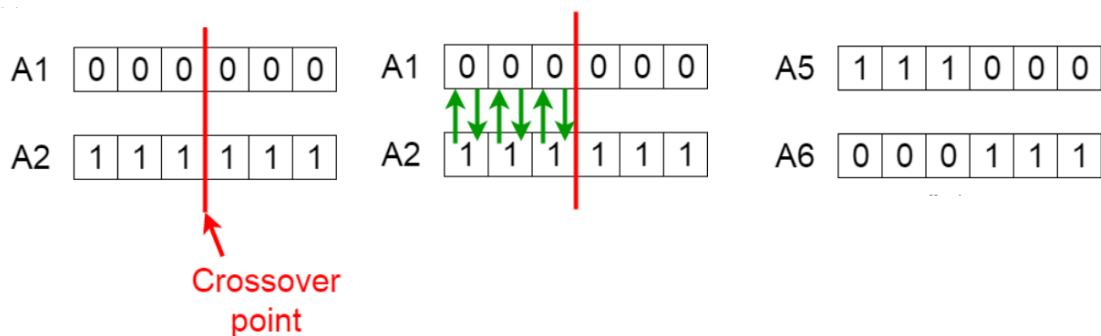


Figure 3.16: Example of crossover or recombination with crossover point =3. For each pair of parents selected, a crossover point is chosen at random from within the genes.

After selection, when both crossover and mutation have been applied to the initial population, a new population is created and a generational counter is increased by one. Again, the process of selection, crossover and mutation are continued until a fixed number of generations (specified by the user) have elapsed or some form of convergence criterion

Before Mutation

A5

1	1	1	0	0	0
---	---	---	---	---	---

After Mutation

A5

1	1	0	1	1	0
---	---	---	---	---	---

Figure 3.17: Example of Mutation, is used to randomly change (flip) the value of single bits within individual strings.

has been met (Coley., 1999).

NSGA-II (Deb et al., 2002) is a stochastic technique that yields a number of final models based on the criterion of Pareto-optimality. The selection process stage and the determination of the fitness of each individual member are specifically geared towards multi-objective optimization (Moorkamp, 2007). Thus, the optimization functional has several components that are arranged into a vector functional:

$$\mathbf{O}(m) = \begin{bmatrix} \Phi_1(m) \\ \vdots \\ \Phi_n(m) \end{bmatrix}, \quad (3.49)$$

where

$$\Phi = \sqrt{\frac{1}{N} \sum_{j=1}^N \left(\frac{d_j^{obs} - d_j^{pred}}{\Delta d_j^{obs}} \right)^2}. \quad (3.50)$$

The equation above corresponds to a normalized RMS misfit for each dataset in the algorithm (Moorkamp et al., 2010). The NSGA-II method differs from single-objective approaches in which a scalar functional is built as

$$O_{so}(m) = \sum_i^n \lambda_i \Phi_i(m), \quad (3.51)$$

where λ is called the *Lagrange multiplier*. NSGA-II does not follow this approach, and

preserves the information from the various minimization functions by ranking the models according to *dominance*.

In the specific implementation of Moorkamp et al. (2010) for the joint inversion of receiver functions, dispersion curves and MT observations, the inversion scheme is as follows:

1. *Choose parameters.* Given a model parameter $\mathbf{m}$, the algorithm creates the genetic string, encoding each individual component m_i to binary representation and concatenating them such that

$$m_i = m_i^{\min} + \delta_i \sum_{j=1}^{n_i} 2^{j-1} s_j, \quad (3.52)$$

where $m_i^{\min}$ is the minimum possible parameter value, δ_i is the discretization step and n_i the number of bits used to encode the parameter value. Those parameters are defined by the user. The s_j represent parts of the genetic string. For example:

i	$m_i^{\min}$	δ_i	n_i
1	0	0.1	3
2	2.5	0.05	6

Table 3.2: Example of binary encoding for a problem with two parameters.

$$\underbrace{010}_{m_1} \underbrace{110101}_{m_2} \rightarrow m_1 = 0.2 \quad m_2 = 4.65$$

2. *Algorithm creates a random population of fixed size P .* With the previous information about the transcription parameters, the algorithm can generate a number of random model strings, the so-called *population*, of size N_P . Each member of this population represents one possible model.
3. *For each population member $\mathbf{m}$, we transform the encoded bit string back into physical parameters and calculate the objective function value $O_j(\mathbf{m})$ for each ob-*

jective function, the misfit of each data set and possible regularization functional.

NSGA-II preserves the multi-objective nature of the problem by storing the objective function values in a vector $\mathbf{O}(\mathbf{m})$ for each population member (Moorkamp et al., 2010). This is a major difference with linearized schemes and some GAs that minimize a weighted sum of the objective function values (e.g. Julià et al., 2000; Gallardo and Meju, 2003; Linde et al., 2006), which avoids the need to specify the weighting for each dataset.

4. While the termination condition is not met:

* *Select models for the next generation.* Each population member is assigned a rank based on the concept of Pareto optimality. A population member $\mathbf{m}_i$ is said to be partially less than the member $\mathbf{m}_j$ if all objective function values are less or equal and at least one is less for $\mathbf{m}_i$. This is expressed as

$$\mathbf{m}_i \prec_P \mathbf{m}_j \Leftrightarrow \forall k : \mathbf{O}_k(\mathbf{m}_i) \leq \mathbf{O}_k(\mathbf{m}_j) \quad \text{and} \\ \exists k : \mathbf{O}_k(\mathbf{m}_i) < \mathbf{O}_k(\mathbf{m}_j), \quad (3.53)$$

where the algorithm only compares objective function values for each data set or regularization functional and, using the criterion of dominance, the algorithm ranks the population. If there is no member in the population that is partially less than a given member $\mathbf{m}$, $\mathbf{m}$ is called locally Pareto optimal and assigned a rank of one. In other words, members that are not dominated by any other, are called Pareto optimal with rank 1. If dominated by another member, they are less than one. Once all non-dominated members have been identified, they are removed from the ranking scheme. Members that were not removed (Pareto optimal front) are assigned, now, rank 2 and the process is repeated until the whole population is ranked (Figure

3.18).

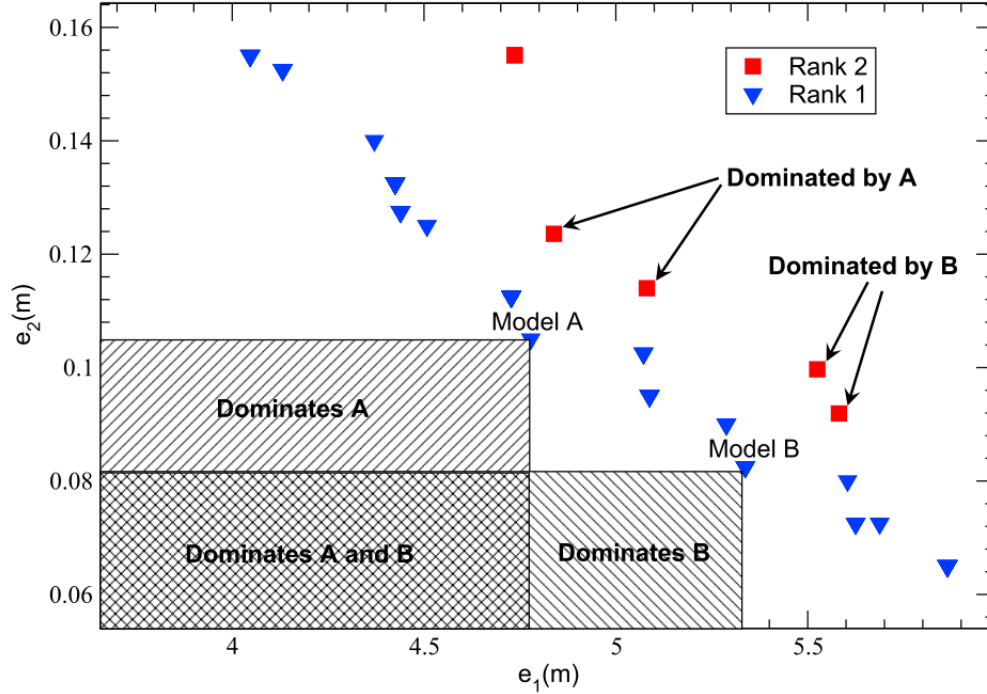


Figure 3.18: The concept of Pareto optimality. For two selected models from the Pareto optimal front, labeled model A and model B, we show the region in objective function space that these models would be dominated by and the members of the population these two model dominate. We also show the rank the population members would be assigned in the genetic algorithm. Font: Moorkamp et al. (2010)

* *Do a binary tournament selection (rank < larger distance, chosen randomly).*

Two random members from the population are compared and the one with lower rank is selected. If both have the same rank, the member with the larger crowding distance (Figure 3.19) is used as a secondary criterion. If both criteria are equal, one member is chosen randomly. The main advantage is that a single model with fitness much higher than average will not be replicated excessively in subsequent generations. A disadvantage is that there is no guarantee that the best model will be selected. To avoid this, NSGA-II uses an *elitist* selection process, where at each iteration, the new generation is combined with the last, the combined population of size $2N$ is ranked, and the N best become the next generation.

Calculation of crowding distances

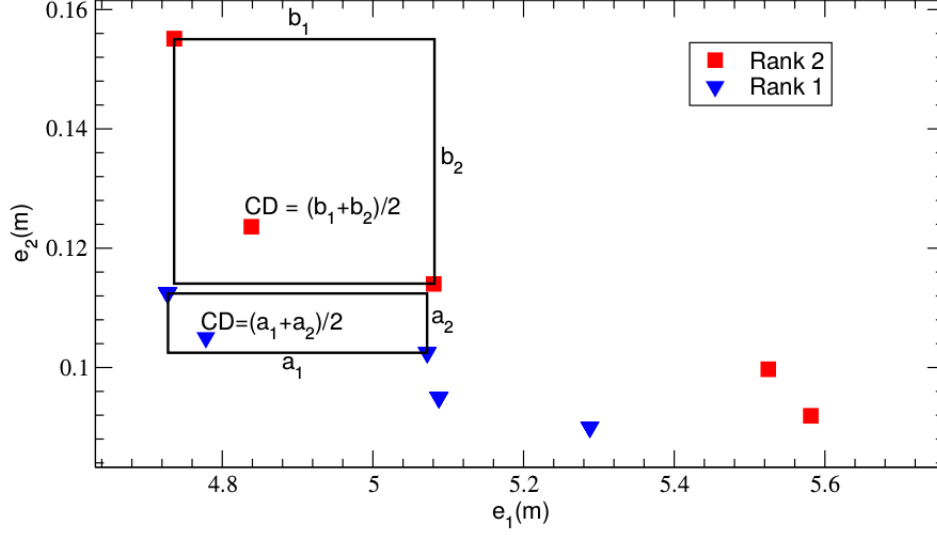


Figure 3.19: Crowding distance. $Rank(a) < Rank(b) \vee Rank(a) = Rank(b) \wedge CD(a) > CD(b)$. Within each rank the distance to the neighbouring models in objective function space is calculated and normalized by maximum distance within that rank. The average of those distances over all objective function components correspond with the crowding distance. Font: Moorkamp (2007)

* *Crossover between models.* With a probability between $0.2 \leq p_c \leq 0.7$, chosen by the user, two models are selected for crossover and exchange their binary representation at a randomly chosen location. This process is very effective at distributing successful segments of the model across the population (Goldberg, 1989).

* *Mutate individual models.* Mutation is a purely unstructured process; with a probability p_m one bit changes its value. This mechanism helps to explore completely new regions of the model space.

* *Perform a new iteration.* Once it is ensured that the best models are always preserved and selected, we repeat the steps above for a fixed number of iterations.

An additional implementation that improved the computational cost (factor 2 or higher) was that for each model a hash value is calculated, which is stored with its corresponding

objective function in an archive during one run of the algorithm. The algorithm checks this archive before the evaluation of objective functions, and whether the model was calculated before its values are taken. Once the algorithm is completed, (N_p populations $\times$ N_g generations), we obtain the *L-curve* or final Pareto front for each combination of datasets and use criteria like maximum curvature to identify the best model.

An example for station NBPN is shown next in detail. We invert for layer thickness t , logarithmic layer resistivity $\log \rho$ and S-wave velocity V_s . We run the genetic algorithm with $N_p = 1000$ and $N_g = 100$ iterations. We must first specify three values for each model parameter (m_i): The minimum possible value (m_i^{min}), the discretization step (δm_i) and the number of bits used for encoding (n_i), where our index i corresponds with $i = 1$ for thickness, $i = 2$ for resistivity, $i = 3$ for S-velocity. All values are listed in Table 3.3.

Layer	1	2	3	4	5	6	7	8	9	10
m_1^{min}	0.1	1	1	1	5	5	10	10	20	20
δ_1	0.1	0.1	0.5	0.5	0.5	0.5	0.5	0.5	1.0	1.0
n_1	6	6	6	6	6	6	6	6	6	0
m_2^{min}	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0
δ_2	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
n_2	5	5	5	5	5	5	5	5	5	5
m_3^{min}	2.5	3.0	3.2	3.2	3.2	3.5	3.5	3.5	3.5	3.5
δ_3	0.0	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05
n_3	6	6	6	6	6	6	6	6	6	6

Table 3.3: Parameter encoding for station NBPN. The index i corresponds with $i = 1$ for thickness, $i = 2$ for resistivity, $i = 3$ for S-velocity. See Equation 3.52

We search for an S-velocity range between 2.5 km/s and 5.0 km/s, which covers realistic values in the crust and upper mantle (Dziewonski and Anderson, 1981). For the MT data, we search for resistivities between $1 \Omega m$ and $10^6 \Omega m$. Even more important is the impact of the discretization step, Δm_i , as we have to make sure that this step size is small enough to avoid aliasing effects. In the inversions (see Chapter 5), we use a step size of 0.05 km/s for the seismic model and 0.1 for logarithmic resistivity in the electric

model (example in 3.3). Also, we keep the number of inversion layers sufficiently small to achieve a good coupling of the data sets. In fact, we perform a number of test with real data to confirm that, for a very large number of layers, the electric and seismic models are indeed decoupled (Moorkamp et al., 2010).

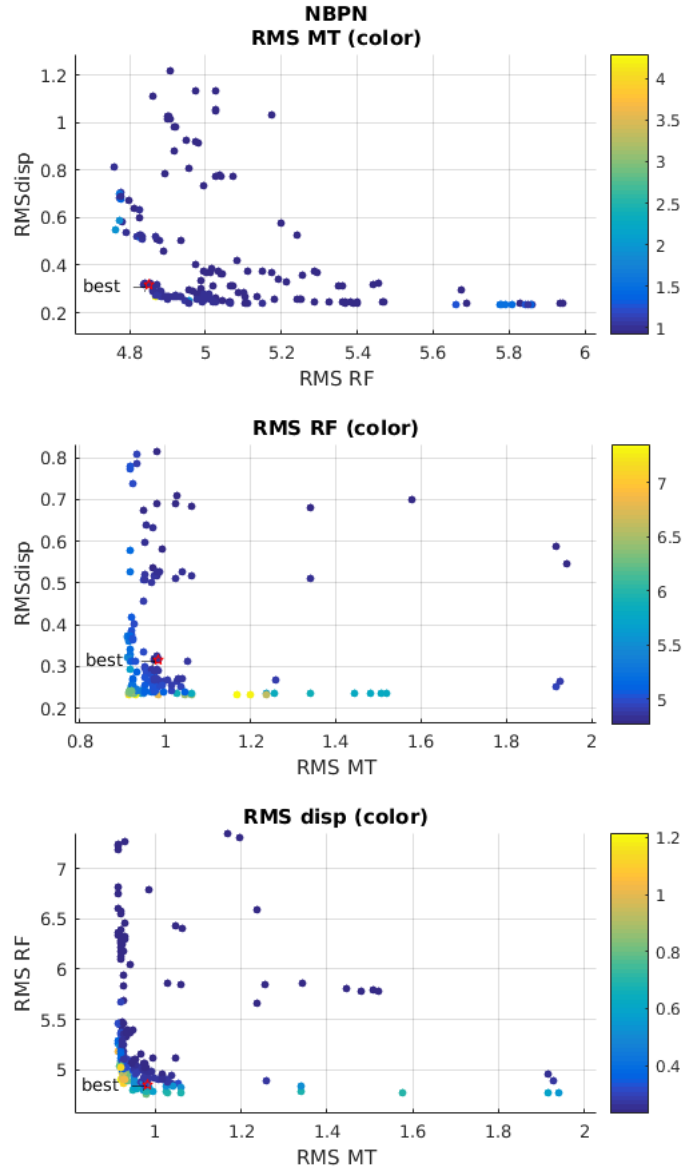


Figure 3.20: Trade-off plots for the data station NBP. Each plot show two objective function values and color the points by third objective function value. The red star is the best model selected.

The best model for site NBPN was identified at the position of highest curvature between dispersion and receiver function misfit, where both have a low misfit and where its MT misfit was around 1.5 (which is sufficiently low). Considering those misfits and our assessment of quality of the features (Figures 3.21, 3.23 and 3.22), we regard this model (red star in Figure 3.20.) as a representative model.

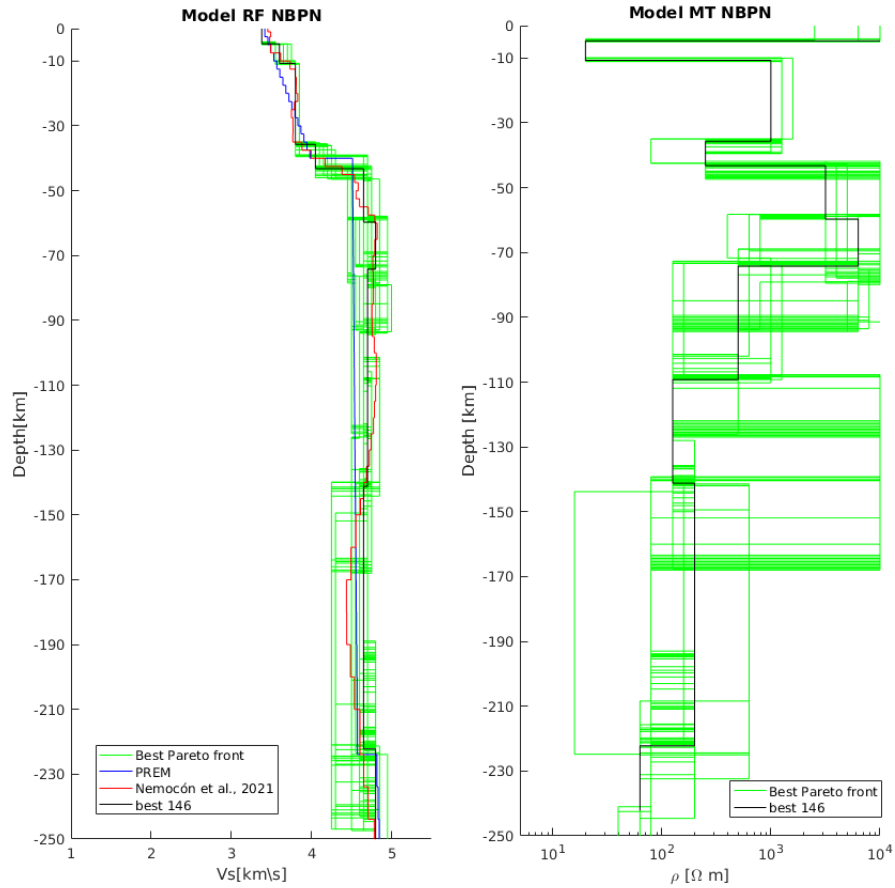


Figure 3.21: The seismic (left) and electric (right) parameters for station NBPN. Plots for all models in Pareto front (green). The best model selected in trade-off, Figure 3.20, (black); comparisons in left panels with the PREM model (blue) and joint inversion with a previous methodology (red).

In general the joint models (Figure 3.21) reproduce first-order structures, such as the Moho interface, with similar depth and velocity as in the joint inversion of dispersion curves and receiver functions calculated with the previous methodology (red line in Fig-

ure 3.21). However, changes in velocity between joint inversion with our first inversion using group velocities and the current, using phase velocities, shows an important difference at mantle levels due to the influence of the dispersion wave, which act as a type of regularization during inversion.

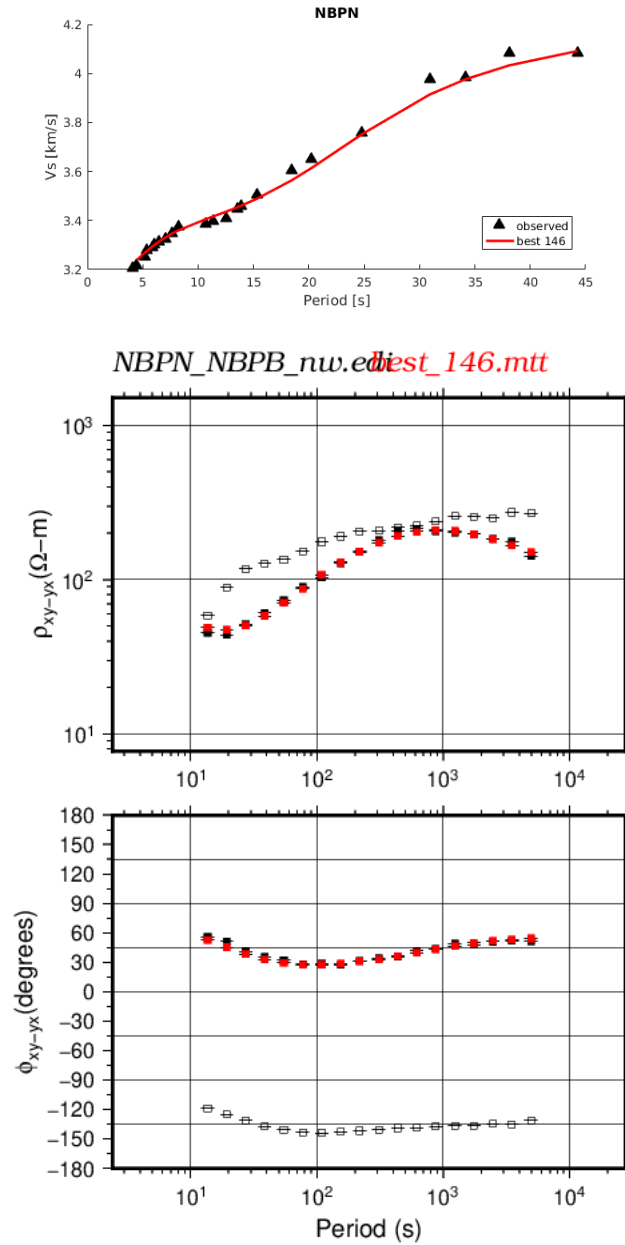


Figure 3.22: The observed (black) and predicted (red) data for site NBPB. Top panel for dispersion curves and bottom panels for resistivity and phase from magnetotelluric data.

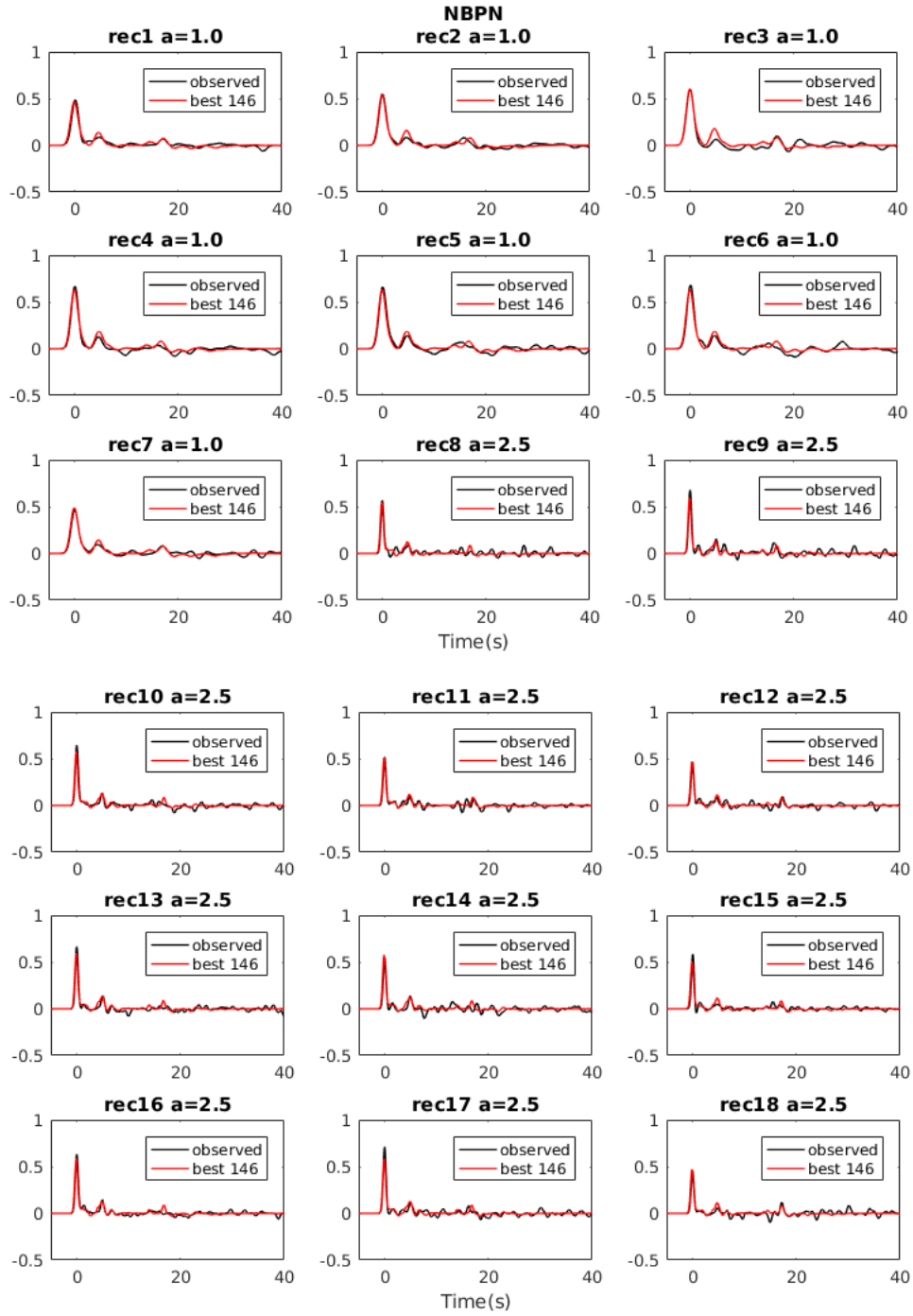


Figure 3.23: The observed (black) and predicted (red) data for Receiver functions in site NBPN.

Chapter 4

Lithospheric structure of the western Borborema Province from receiver functions and surface-wave dispersion: Implications for basin inversion. (Paper I)

The following article, entitled *Lithospheric structure of the western Borborema Province from receiver functions and surface-wave dispersion: Implications for basin inversion*, reports on the results obtained from the joint inversion of receiver functions and dispersion (group-velocity) curves. This manuscript was published in Tectonophysics, on 12th Aug 2021.

1 Lithospheric structure of the western Borborema
2 Province from receiver functions and surface-wave
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13 **Abstract**

14 The Proterozoic Borborema province of NE Brazil was under extensional
15 stresses during the Mesozoic, causing stretching and thinning of the litho-
16 sphere. During Cenozoic times, compression associated with the develop-
17 ment of the South Atlantic and collision along the Andean front dominated
18 throughout the Province. One of the main consequences of such compression
19 might have been the formation of the Chapada de Araripe, an inverted rift
20 basin located in the Western Province. We undertook a passive seismic ex-
21 periment along a NS profile crossing the Borborema province and a short EW
22 profile across the Chapada de Araripe, with the goal of providing constraints
23 on the origin of the Chapada de Araripe. A variety of mechanisms can
24 contribute to basin uplift, so understanding basin inversion is in turn impor-

25 tant for understanding the tectonic evolution and geodynamics of NE Brazil.
 26 Through joint inversion of receiver functions and surface-wave dispersion, we
 27 show that the crust is typically 34-38 km thick under the western Province; in
 28 contrast, the crust is 43-46 km thick under the northern craton. No apparent
 29 crustal thinning is observed under the Araripe Basin, due to the presence of
 30 a relatively thick (10-14 km) high-velocity (4.1 - 4.3 km/s) lowermost crust.
 31 Our velocity models also image a drop in S-velocity (from 4.6 km/s to 4.3
 32 km/s) at 135-145 km depth under the southern half of the profile, which we
 33 interpret as a mid-lithospheric discontinuity, and at 120-140 km depth under
 34 the Araripe Basin, likely representing the lithosphere-asthenosphere bound-
 35 ary. Our findings confirm the presence of thin lithosphere under the Araripe
 36 Basin - originally postulated by a collocated MT survey - that might have
 37 facilitated tectonic inversion of this basin through regional stress concentra-
 38 tion and thermal buoyancy. These findings also suggest additional buoyancy
 39 contributed by a mafic lowermost crust right under the basin.

40

41 *Keywords:* Basin inversion, Mafic lowermost crust, Araripe Basin,
 42 Receiver functions.

43 1. Introduction

44 The Borborema Province is located in the northeastern most corner of the
 45 South American continent and represents one of the Neoproterozoic mobile
 46 belts that make up Precambrian South America. To the south, the BP is
 47 bounded by the (mostly) Archean São Francisco craton, an old fragment of
 48 former supercontinent Rodinia that amalgamated to West Gondwana during
 49 the Brasiliano-Pan African orogeny at 600 Ma. To the west, the sedimentary

50 cover of the cratonic Parnaíba Basin partly conceals BP basement rocks. And
51 to the north and east, marginal basins that developed in relation to continen-
52 tal splitting between Africa and Brazil border the Province (e.g. Brito Neves
53 et al., 1984). The western Province hosts the Chapada do Araripe, an aborted
54 Mesozoic rift basin in which post-rift strata that were originally deposited
55 close to sea-level now form a plateau lying at 800-1000 m altitude (Peulvast
56 and Bétard, 2015).

57 The processes that led to the unusual topography of the Araripe basin,
58 when compared to other nearby rift basins (see Figure 1), are still debated.
59 Inversion conventionally results from application of horizontal compressive
60 stresses (compressional tectonics) following basin formation (Buchanan and
61 Buchanan, 1995); however it may also include topographic inversion (i.e.
62 regional epeirogenic uplift and exhumation) due to a lithospheric anomaly
63 (Brodie and White, 1995). Indeed, two main models have been proposed
64 for the Araripe Basin: (i) basin inversion, in which uplift resulted from
65 plate-wide horizontal compression that reactivated inherited faults and shear
66 zones into the main inversion faults (Marques et al., 2014), perhaps aided by
67 thinned lithosphere (Garcia et al., 2019), and (ii) regional epeirogenic up-
68 lift, in which the elevated basin would have resulted from regional uplift and
69 subsequent differential erosion (Peulvast and Bétard, 2015).

70 Inversion of an intra-continental basin is thus not a trivial problem (e.g.
71 Sandiford (1999); Bonini et al. (2011); Hansen and Nielsen: 2002; 2003)
72 and, yet, little is known about the deep structure of the crust and litho-
73 sphere surrounding this enigmatic basin. To fill this gap, this study provides
74 new constraints on crustal and lithospheric thickness, and deep S-velocity

75 structure under and around the Araripe Basin. More specifically, we use
76 teleseismic P-wave receiver functions to investigate crustal and lithospheric
77 mantle structure under 13 seismic stations located in the western Province
78 and northern São Francisco craton (see Figure 1).

79 Our results show that the crust in the northern São Francisco craton is
80 43-46 km thick with a bulk V_P/V_S ratio around 1.70, while the crust in the
81 western province is 34-38 km thick with a bulk V_P/V_S ratio around 1.73.
82 The crust in both regions display relatively slow S-velocities (≤ 4.0 km/s)
83 down to Moho depths. Crustal thickness under the Araripe Basin is similar
84 to that immediately North and South, but displays a thick (~ 10 km) layer
85 of fast-velocity material ($4.1 - 4.3$ km/s) right above the Moho, which we
86 interpret as mafic lowermost crust. Our velocity models also reveal that the
87 lithosphere is 120-140 km thick under the Araripe Basin, which is broadly
88 consistent with lithospheric thicknesses imaged by a collocated magnetotel-
89 luric survey (Garcia et al., 2019). We argue that our results support models
90 invoking tectonic inversion for the Araripe Basin, and that the anomalously
91 high topographies of this basin might be the result of stress concentration
92 and thermal buoyancy due to a thin lithosphere, aided by compositional
93 buoyancy due to local mafic intrusion of the basin’s lower crust.

94 **2. Geology and tectonic setting**

95 *2.1. Regional framework*

96 The BP was defined by Almeida et al. (1981) as a complex orogenic sys-
97 tem that developed at the end of the Neoproterozoic (Brasiliano Orogeny)
98 after several deformational, metamorphic and magmatic episodes. Its struc-

99 turation has been traditionally attributed to the accretion of allochthonous
100 tectonic blocks (Jardim de Sá et al., 1999) - mostly microplates and other
101 small continental fragments scattered among the main cratonic units - that
102 collided along with the Amazonian, Congo-São Francisco and West African
103 cratons, followed by extrusion tectonics (Ganade de Araujo et al., 2013). The
104 suture zones related to the tectonic collisions were thus reactivated to accom-
105 modate shear deformation, giving rise to the extensive network of shear zones
106 (e.g. Patos and Pernambuco lineament) that now pervades the Province
107 (Ganade de Araujo et al., 2013). Some of these shear-zones - such as the
108 EW trending Patos and Pernambuco lineaments or the NE-SW trending
109 Transbrasiliiano Lineament - are large-scale structures that continue into the
110 African continent in paleogeographic reconstructions (de Matos, 1992, 1999;
111 Ganade de Araujo et al., 2013).

112 This network of shear zones was partly reactivated in the Mesozoic to
113 help accommodate the extensional stresses that eventually led to the sepa-
114 ration between Africa and South America. Deformation developed within
115 the BP along three principal axes or trends: Cariri-Potiguar, Recôncavo-
116 Tucano-Jatobá and Gabon-Sergipe-Alagoas (Matos, 1992; 1999). With the
117 opening of the Equatorial Atlantic, rifting aborted along the Cariri-Potiguar
118 and Recôncavo-Tucano-Jatobá trends and continued along the Gabon-Sergipe-
119 Alagoas trend, eventually leading to continental break-up. Marginal basins
120 developed along the Brazilian (and African) conjugate margins, while a num-
121 ber of aborted syn-rift basins were left in the continental interior along the
122 two first axes of deformation. In particular, the 500 km-long, NE-SW trend-
123 ing Cariri-Potiguar trend hosts a number of rift basins and half-grabens

124 within the western Province. The location and architecture of these basins
125 appear to have been strongly controlled by upper crustal heterogeneities (de
126 Matos, 1999) and include the Araripe, Rio do Peixe, Potiguar, Iguatu, Mal-
127 hada Vermelha and Lima Campos basins (see Figure 1).

128 After continental breakup, the evolution of the Province was character-
129 ized by episodes of intra-plate volcanism and uplift. Volcanism concentrated
130 along two othogonal magmatic alignments in the Eastern Province: Fer-
131 nando de Noronha-Mecejana, trending EW and located off-shore; and Macau-
132 Queimadas, trending NS and located on-shore. K-Ar dating determined ages
133 of 26-34 My for Mecejana volcanism (Mizusaki et al., 2002) and 22-2 My for
134 volcanic rocks at the Fernando de Noronha archipelago (e.g Knesel et al.,
135 2011), displaying a clear age progression from West to East. K-Ar and Ar-
136 Ar ages for the Macau-Queimadas Alignment, on the other hand, range from
137 80 to 30 My (Mizusaki et al., 2002) or 93-7 My (Knesel et al., 2011)), with
138 no clear age progression. Magmatic activity, therefore, appears to have been
139 restricted to Late Cretaceous-Ceneozoic times and north of the Patos Linea-
140 ment, the lack of Late Jurassic magmatism along the GSA and RTJ trends
141 suggesting a cold, non-volcanic rifting style (de Matos, 1999).

142 Uplift was mainly expressed through the high elevations of the Borborema
143 Plateau to the East, with maximum elevations of ~ 1200 m, and the Chapada
144 do Araripe to the West, with elevations of ~ 1000 m. Uplift in the Borborema
145 Plateau has been reported to be coeval with Cenozoic volcanism (e.g. Oliveira
146 and de Medeiros, 2012), although consensus has not yet been reached (Moraes
147 Neto et al., 2009; Luz et al., 2015).

148 *2.2. Stratigraphy and basin architecture*

149 The Araripe Basin is the most extensive of all the intra-continental Meso-
150 zoic rift basins of the Brazilian northeast. Although it manifests as a geomor-
151 phological feature elongated in the EW direction (Figure 1), its underlying
152 structure consists of asymmetric grabens oriented in the NE-SW direction
153 (Marques et al., 2014). Regional gravity, magnetic and seismic surveys have
154 revealed the Araripe Basin comprises two sub-basins, the Feira Nova sub-
155 basin to the west and the Cariri sub-basin to the East, separated by a central
156 Dom Leme horst (Assine, 2007). The eastern portion of the basin stands at
157 an average altitude of about 1000 m a.m.s.l., forming a plateau known as the
158 Chapada do Araripe. The Chapada dips gently to the west and is separated
159 by the adjacent Vale do Cariri through deep erosive scarps (Assine, 2007).

160 The crystalline basement of the Araripe Basin is composed of gneiss-
161 magmatitic units and granitoids, and metamorphic rocks of Precambrian
162 age deformed during the Brasiliano orogeny (750-540 My). Its stratigraphic
163 record includes up to 7 lithostratigraphic units and 5 super-sequences, sep-
164 arated by regional unconformities, which resulted from fragmentary deposi-
165 tion during the pre-, syn-, and post-rifting stages (Assine, 2007). Pre-rift
166 sedimentation started in Late Jurassic times (Brejo Santo and Missão Velha
167 Formations), with deposition of clays, siltites and sandstones on top of pre-
168 existing Paleozoic sandstones (Cariri Formation) that covered Precambrian
169 basement rocks (Assine, 2007). Pre-rift sedimentation likely followed exten-
170 sional deformation related to the South Atlantic opening, which extended
171 as far north as the Patos Lineament forming a larger Afro-Brazilian Depres-
172 sion (Matos 1992; 1999). Syn-rift sedimentation progressed during the lowest

173 stages of the Cretaceous period (Neocomian), with deposition of silty shales
174 and siltstones with fine sandstones interbedded (Abaíara Formation). This
175 deposition coincided with the main rifting phase in the BP, which involved
176 stretching and thinning of the crust along all three linear trends (GSA, RTJ,
177 and CP) and, eventually, continental breakup along the GSA trend.

178 Sedimentation continued during the Early Cretaceous with post-rift de-
179 position of the Barbalha and Santana Formation sediments, and until Late
180 Cretaceous (Cenomanian) with deposition of the Araripina and Exu Forma-
181 tion sediments (Assine, 2007). The Barbalha Formation is dominated by
182 sandstones, which appear interbedded with shales and conglomerates, while
183 the Santana Formation includes limestones, gypsum, sandstones, conglom-
184 erates, siltstones and shales. These sedimentary sequences can be correlated
185 with sequences found in the marginal basins of the eastern continental mar-
186 gin (Assine, 2007). The Araripina Formation is present only in the western
187 portion of the basin and consists of conglomerates and sandstones, which are
188 overlain by fluvial sandstones from the Exu formation. According to Peulvast
189 and Bétard (2015), this sedimentary layer ensures, where present, a “strong
190 contrast in mechanical resistance to erosion between the sedimentary cover
191 and the crystalline rocks of the basement”. The Exu sandstones are also
192 present in the eastern portion of the basin, although with coarser grains.

193 A striking feature of the crust underlying the Araripe Basin is the ap-
194 parent lack of evidence for crustal thinning and stretching. According to De
195 Castro & Medeiros (1997), crustal thinning related to NW-SE extension along
196 the the Cariri-Potiguar trend did not affect the crust south of the Patos Lin-
197 eament, including that under the Araripe Basin. Indeed, crustal thicknesses

198 of 30-32 km were reported for the Potiguar basin and other marginal basins
 199 around the BP by Luz et al. (2015) and Almeida et al. (2015), which con-
 200 trasted with thicker values of 35-39 km in the southern Borborema Plateau
 201 and easternmost tip of the Araripe Basin. Interestingly, Garcia et al. (2019)
 202 did report evidence for lithospheric thinning under the Araripe Basin. Anal-
 203 ysis of long-period electromagnetic data collected during a magnetotelluric
 204 survey crossing the Araripe Basin (collocated with the seismic stations con-
 205 sidered in this study) revealed a marked conductive anomaly ($\sim 25 \Omega m$) start-
 206 ing at 120 km depth under the basin flanked by more resistive ($> 65 \Omega m$)
 207 anomalies immediately north and south. High conductivities were interpreted
 208 as enriched asthenospheric material, involving small amounts of fluids and
 209 melt, that were preserved until today by lateral asthenospheric flow through
 210 a thin lithospheric channel under the basin inherited from Mesozoic extension
 211 (Simões Neto et al., 2019).

212 *2.3. Basin inversion*

213 A most striking feature of the Araripe Basin is that a significant por-
 214 tion of its post-rift cover presently stands at 1000 m above mean sea level
 215 (Assine, 2007). Two main models have been proposed to explain the anoma-
 216 lous elevation of the basin's sediments. On one hand, Peulvast and Bétard
 217 (2015) analyzed topographic, geologic, morphostructural, and morphopedo-
 218 logical data to conclude that the present elevation of the Chapada do Araripe
 219 is related to differential erosion and morphological resistance in response to
 220 epeirogenic uplift. On the other hand, Marques et al. (2014) considered
 221 topographic, gravity and structural/tectonic data to propose that basin in-
 222 version resulted from plate-wide compression and reactivation of inherited

223 Precambrian shear zones to form the main inversion faults.

224 The first mechanism, labeled as topographic inversion by Peulvast and
225 Bétard (2015), postulates that tectonic uplift of regional proportions would
226 have affected the entire BP in Early Cenozoic times, and that differential
227 erosion - controlled through the strong lithological contrast imposed by the
228 Araripina caprock - would have then shaped the landscape. According to this
229 model, the Chapada do Araripe would represent a remnant of a much larger
230 sandstone plateau that might have extended all the way to the equatorial
231 margin. Although no regional uplift mechanism was identified, the authors
232 related the postulated crustal upwarp to Cenozoic uplift of the neighbouring
233 Borborema Plateau and to ongoing epeirogeny identified through gravimetric
234 and apatite fission-track analysis (AFTA) data (Oliveira and de Medeiros,
235 2012; Morais Neto et al., 2009). Moreover, the authors emphasize that the
236 same dataset yields no evidence for local tectonic inversion.

237 Indeed, AFTA on rock samples from the Chapada do Araripe (Hegarty
238 et al., 2002) suggest basin inversion was coincident with two regional cooling
239 events: one at ca. 100–90 Ma, and another from 30 to 0 Ma, which were also
240 identified in the Borborema Plateau. This coincidence suggests that these
241 events are necessarily regional in nature, and therefore likely to be related to
242 epeirogenic exhumation. Moreover, according to those same AFTA results,
243 the present topography of the Chapada do Araripe would be attributed to
244 the Late Tertiary event and resulted from removal of at least 1.3 km of the
245 original sedimentary cover.

246 The topographic inversion model contrasts with that proposed by Mar-
247 ques et al. (2014), by which inversion of the Araripe Basin would be the

248 result of local tectonic inversion. This model postulates that horizontal com-
 249 pressive stresses originating from the development of the South Atlantic to
 250 the east and the Andean chain to the west would have been responsible
 251 for oblique reactivation of rift-related, high-angle normal faults and the cre-
 252 ation of new, low-angle thrust faults. Moreover, the authors emphasize that
 253 inversion would have been possible because precursor normal faults were sig-
 254 nificantly oblique to the inversion-related stresses, and because of lubrication
 255 by clay-rich diapirs or walls. They also acknowledge that their model would
 256 work best if local necking of the lithosphere were present under the basin, as
 257 that would induce differential crustal thinning and increase post-rift crustal
 258 weakness (Hansen and Nielsen, 2003). As mentioned earlier, local necking of
 259 the lithosphere under the Araripe Basin was reported by Garcia et al. (2019)
 260 from analysis of magnetotelluric data.

261 **3. Data, receiver function calculation and dispersion**

262 *3.1. Seismic deployment*

263 The dataset utilized in this work was acquired by the Universidade Fed-
 264 eral do Rio Grande do Norte mostly as part of the BORborema Deep Electro-
 265 magnetic and Seismic (BODES) experiment, a temporary array of colocated
 266 seismic and magnetotelluric stations sampling the western BP and northern
 267 São Francisco craton. The seismic network consisted of nine broadband sta-
 268 tions from the *Pool de Equipamentos Geofísicos do Brasil* (PEGBR) flanked
 269 by two permanent broadband stations belonging to the *Rede Sismográfica do*
 270 *Nordeste* (RSISNE). Overall, the 11 stations formed an approximately 600
 271 km-long linear profile trending in the NS direction, with an interstation spac-

ing of ~ 60 km (see Figure 1). Two short-period stations, also belonging to the PEGBR, were additionally deployed on top of the Araripe Basin between two broadband stations.

The broadband stations were equipped with RefTek 151-120 sensors, with flat response in velocity in the 0.008-50 Hz frequency band, feeding 24-bit digitizers and sampling at 100 Hz. The short-period stations consisted of Sercel L4A-3D sensors, with flat velocity response above 2 Hz and also sampling at 100 Hz. GPS antennas were used for all stations for time keeping. The broadband network was initially deployed in January 2015 with 8 sites (AR01 - AR08), and in April 2016 the equipment in site AR01 was moved South to site AR09 to complete the line. The deployment was dismantled in March 2017, yielding an effective time of operation of over 2 years for the central stations (AR02 - AR08) and about 1 year for the two remaining stations (AR01, AR09). The two short-period stations were deployed between November 2015 and May 2017 (AR50), and from August 2016 and May 2017 (AR51), respectively.

In October 2018 three additional stations were deployed on top of the Araripe Basin, as part of a larger effort funded by the national oil company Petrobras to investigate the crustal structure under the syn-rift basins of NE Brazil. Along with an additional RSISNE station, the new deployment formed a shorter, ~ 170 km-long EW line that complemented the previous, longer NS line. The EW line consisted of two short-period PEGBR stations flanked by a RSISNE station to the East and the newly deployed broadband station to the West (see Figure 1). The newly deployed broadband station belongs to the Laboratório Sismológico (LabSis/UFRN) and was equipped

297 with a GeoTech broadband sensor (120 s) and a high-gain SMART-24 digi-
298 tizer sampling at 100 Hz. Again, GPS antennas were utilized for time keep-
299 ing.

300 Station coordinates and recording time windows are listed in Table 1.
301 Note that RSISNE stations have been in operation since 2013, and that the
302 EW profile stations did not overlap in time with the BODES experiment.

303 *3.2. Receiver functions*

304 Seismic sources for receiver function calculation were selected considering
305 epicentral distances between 30° and 90° , with no depth restrictions, and
306 magnitude greater than 5.5 mb. A total of 2536 waveforms were selected for
307 processing along the NS profile, with most stations having over 150 wave-
308 forms. Due to shorter recording times and operational problems, stations
309 AR01 (143 waveforms), AR09 (55 waveforms), AR50 (120 waveforms), and
310 AR51 (69 waveforms), recorded less data. Along the EW profile, a total of
311 649 waveforms were selected for processing: SERO (142 waveforms), TBCA
312 (166 waveforms), BBLH (168 waveforms) and NBMA (173 waveforms). The
313 location of the selected seismic sources is displayed in Figure 2.

314 Receiver functions were computed from deconvolution of the vertical com-
315 ponent from the corresponding horizontal components to produce radial and
316 transverse receiver functions, respectively (Langston, 1979). Radial receiver
317 functions consist of a leading P-wave at zero lag-time followed by a num-
318 ber of P-to-S converted waves resulting from the interaction of the leading
319 P-wave with sub-surface discontinuities. Analysis of the corresponding S-P
320 travel-times and amplitudes are then utilized to reveal the S-velocity struc-
321 ture under the recording station (e.g. Owens et al., 1984; Ammon et al., 1990;

322 Julià et al., 1998). Transverse receiver functions should be identically zero
323 for isotropic, laterally homogeneous media, and large transverse amplitudes
324 are considered diagnostic of subsurface anisotropy, dipping interfaces and/or
325 complex 3D structures (e.g. Levin and Park, 1997; Bianchi et al., 2010).

326 In order to compute receiver function estimates from the selected wave-
327 forms, we applied the iterative, time-domain deconvolution approach of Lig-
328 orría and Ammon (1999), using 500 iterations. First, the teleseismic wave-
329 forms were windowed 10 s before and 110 s after the P-wave, demeaned, de-
330 trended and tapered. The resulting time series were then high-pass filtered
331 above 0.025 Hz to remove long-period noise, and low-pass filtered below 4
332 Hz to avoid aliasing before resampling to 10 Hz. The deconvolution oper-
333 ation was next performed on the decimated waveforms, after rotation into
334 the great-circle-path. The resulting time series were finally low-pass filtered
335 with acausal Gaussian filters of width $a=1.0$ and 2.5 , which produced re-
336 ceiver function estimates at overlapping frequency bands of $f < 0.5$ and $f <$
337 1.2 Hz. This frequency overlap helps discriminate sharp discontinuities from
338 gradational boundaries in receiver function analysis (see e.g. Julià, 2007).

339 A strict quality control was applied to the resulting receiver functions.
340 First, receiver functions that did not reproduce at least 85% of the original
341 radial waveform when convolved back with the vertical component were re-
342 jected. Second, receiver functions were visually inspected and those exhibit-
343 ing unstable behaviour and/or exceedingly large amplitudes in the transverse
344 receiver function were excluded. Third, radial receiver functions were super-
345 imposed to identify outliers, which were also excluded. In total, for the 17
346 seismic stations, 513 receiver functions at high-frequency contents ($a=2.5$)

347 and 580 receiver functions at low-frequency contents ($a=1.0$) were selected
348 for further analysis.

349 Sample radial and transverse receiver functions are displayed in Figure
350 3. The radial receiver functions contains a large peak at zero lag time, cor-
351 responding to the direct P-wave, followed by secondary peaks and troughs
352 corresponding to P-to-S conversions. In particular, we note prominent peaks
353 at lag times of about 5 and 14 s that are likely related to the P-to-S re-
354 fraction (Ps) and first reverberation (PpPs) at the crust-mantle boundary,
355 respectively. Also note that the transverse receiver functions have relatively
356 small amplitudes when compared to the radial amplitudes, indicating that
357 the medium under the recording station can be approximated as isotropic
358 and laterally homogeneous.

359 *3.3. Rayleigh-wave group velocity*

360 Long-period dispersion measurements were obtained from a continental-
361 scale tomographic study with focus on the BP (Luz et al., 2015), which
362 in turn built on the surface-wave tomography study of Feng et al. (2004).
363 In those studies, Rayleigh-wave group velocity estimates were obtained for
364 periods between 10 and 140 s, at 10 s period intervals, using the multiple
365 filter analysis of Dziewonski et al. (1969), after isolation of the fundamental
366 mode with a phase-match filter (Herrin and Goforth, 1977). The single-
367 station measurements were inverted tomographically with the procedure of
368 Feng et al. (2004) using a grid of $1.0^\circ \times 1.0^\circ$ tomographic cells covering a
369 square area enclosing the entire continent.

370 Short-period dispersion velocities, on the other hand, were taken from
371 the ambient noise tomography study of Dias et al. (2015). In that study, a

total of 333 empirical Green’s functions were developed from stacked cross correlations of 1 month of seismic ambient noise recorded in the vertical component of broadband stations in the BP. The Fast Marching Surface-wave Tomography (FMST) scheme of Rawlinson and Sambridge (2005) was utilized to produce maps of group velocity variation for the fundamental-mode of reconstructed Rayleigh waves over a grid of $0.5^\circ \times 0.5^\circ$ tomographic cells in the BP and surrounding regions. The model was parameterized with 551 velocity nodes, and 2D group-velocity maps were produced for periods of 5, 10, 15, and 20 s. Short-period dispersion, however, may conflict with zero-lag receiver function amplitudes, as those amplitudes are sensitive to absolute S-velocity at shallow depths (Ammon, 1990; Julià, 2007). Thus, in order to keep our velocity models as local to the recording station as possible, we opted for not using dispersion measurements at 5 s.

4. Crustal Architecture

In this section, we investigate variations in crustal thickness and bulk V_P/V_S ratio from the H- κ stacking technique of Zhu and Kanamori (2000), variations in S-velocity with depth from 1D S-velocity profiles at individual stations developed from the joint inversion with dispersion velocities (Julià et al., 2000; 2003), and lateral variations of seismic discontinuities from a migrated receiver function cross-section developed with the Common Conversion Point (CCP) stacking of Frassetto et al. (2010).

4.1. Crustal thickness and bulk V_p/V_s ratio

The H- κ stacking technique of Zhu and Kanamori (2000) makes use of the moveout of the secondary phases generated at the Moho to infer basic prop-

396 erties of the crust such as thickness and bulk V_P/V_S ratio. The algorithm
 397 stacks receiver function amplitudes along predicted arrival time curves for
 398 the Moho Ps and crustal multiples (PpPs and PpSs+PsPs) calculated from
 399 a range of crustal thicknesses (H) and V_P/V_S ratios (κ), without the need to
 400 identify phases and pick arrival times. The best estimates of crustal thick-
 401 ness and bulk V_P/V_S ratio are found when the three phases stack coherently
 402 along the phase-moveout curves, resulting in a well-defined maximum in the
 403 H- κ stacking surface (Zhu and Kanamori, 2000).

404 The approach requires the assumption of a P-wave velocity value, which
 405 was taken as 6.5 km/s from the global average for continental areas (Chris-
 406 tensen and Mooney, 1995). It also requires *a priori* weights for the three
 407 crustal phases (Ps, PpPs, PpSs+PsPs). We adopted the criterion of Julià
 408 and Mejia (2004) and assign weighted values of 0.4, 0.3, and 0.3, respectively,
 409 when all three phases were clearly observed; values of 0.5, 0.5, and 0.0, when
 410 the second crustal multiple was not observed; and values of 0.5, 0.0, and
 411 0.5, when the first multiple was not observed. Recent work by Ogden et al.
 412 (2019) showed that H- κ stacking results can be quite dependent on parame-
 413 ters - and even break down - in places with complex crustal structure. Our
 414 approach to overcoming complexity was to limit the crustal thickness range
 415 around values inferred from the joint inversion (see section 4.3) in order to
 416 seek results that were consistent with the inverted S-velocity models. Con-
 417 fidence levels for the estimates were obtained through a bootstrap approach
 418 (Efron and Tibshirani, 1991), from the distribution of H and κ values of 200
 419 replications.

420 Examples illustrating the performance of the H- κ stacking procedure are

421 given in Figure 4. The analysis was applied to high-frequency receiver func-
 422 tions in order to obtain the best resolution possible (see Table 1). The figure
 423 displays the H- κ stacking surface in the left panels and the receiver function
 424 waveforms, sorted by ray parameter, in the right panels. In both cases a
 425 single, well-defined maximum is observed in the H- κ stacking surface, yield-
 426 ing crustal thicknesses of 39.5 ± 0.5 km and 44.6 ± 1.1 km and V_P/V_S ratios of
 427 1.75 ± 0.02 and 1.68 ± 0.02 for stations AR04 and NBPN, respectively.

428 The results for the entire seismic line are summarized in Figure 5 and Ta-
 429 ble 1. Crustal thickness ranges from 34.6 ± 0.4 km in the Borborema Province
 430 (station AR02) to 44.6 ± 1.1 in the São Francisco craton (station NBPN),
 431 while V_P/V_S ratios vary between 1.64 ± 0.03 (station AR01) and 1.80 ± 0.03
 432 (station AR51). Although no clear progression in either crustal thickness
 433 or V_P/V_S ratio is observed along the line, we do notice a somewhat grad-
 434 ual thickening of the crust, from the Tatajuba-Jaguaribe shear zone in the
 435 northern end, to the São Francisco craton. The results along the EW profile
 436 beneath Araripe basin shows V_P/V_S ratios of 1.75 ± 0.06 and 1.80 ± 0.02 . Most
 437 interestingly, no apparent crustal thinning is observed under the basin.

438 4.2. Crustal cross-section

439 Another way to analyze the receiver function dataset is to apply the Com-
 440 mon Conversion Point (CCP) migration and stacking procedure of Frassetto
 441 et al. (2010), which combines the CCP stacking of Gilbert and Sheehan (2004)
 442 with the phase-weighting scheme of Schimmel and Paulssen (1997). First,
 443 receiver functions are migrated from time to depth by ray-tracing through an
 444 average 1D velocity model and assuming Ps conversions at given depth in-
 445 tervals. The geographical location of the Ps conversions (piercing points) are

446 next calculated at each depth and binned around a line of uniformly spaced
447 nodes. Receiver function amplitudes associated to the piercing points within
448 each bin and depth are then averaged with phase-weighting to produce the
449 final receiver function stacks for each node. The phase-weighting scheme acts
450 as a filter of coherent arrivals, effectively improving the signal-to-noise ratio,
451 while the stacking procedure smooths azimuthal heterogeneities at shallow
452 depths (Frassetto et al., 2010).

453 According to Frassetto et al. (2010), the size and spacing of the bins used
454 for stacking depend on the aperture and density of stations in the array. In
455 our case, we considered nodes spaced at 50 km and bins with radii of 100 km
456 that enclosed a minimum of 5 piercing points. This choice of bin radii and
457 node spacing provided an optimal balance between lateral resolution and
458 spatial coverage along the BODES Experiment seismic line. The piercing
459 points were calculated for depths between 0 and 100 km, at 1 km depth
460 intervals. For the back-projection, the 1D velocity model by Almeida et al.
461 (2015) was utilized. The resulting CCP cross-section is displayed in Figure
462 6, showing clear P to S conversions suggestive of a crust-mantle boundary
463 at around 34-44 km depth. A mid-crustal discontinuity is also observed at
464 10-20 km depth.

465 4.3. *S-wave velocity-depth profiles*

466 The last analysis that was applied to our dataset is the joint inversion
467 of receiver functions and surface-wave dispersion velocities. We follow the
468 scheme outlined in Julià et al. (2003), in which the difference between ob-
469 servations and predictions is minimized in a least-squares sense. As the ve-
470 locity model is over-parameterized through a stack of thin layers, the scheme

471 incorporates smoothness constraints that minimize the second derivative of
472 velocity between adjacent layers. The contribution of the regularization term
473 to the total norm is controlled by an *a priori* smoothness parameter θ , while
474 the relative contribution of each data set is set through an *a priori* influence
475 parameter ($0 \leq p \leq 1$), after normalizing for the corresponding number of
476 data points and physical units (variance). Both the influence and smoothness
477 parameters are determined through analysis of trade-off curves.

478 The starting model consisted of a perfectly elastic, laterally homogeneous
479 and isotropic medium defined through a stack of thin layers of fixed thickness
480 and uniform velocity. The crust was defined through a gradational increase
481 in S-velocity from 3.4 km/s at the surface to 4.0 km/s at 40 km depth.
482 Uppermost mantle velocities were set to 4.5 km/s, increasing smoothly to
483 match Preliminary Reference Earth Model (PREM) velocities at 220 km
484 depth. Layer thickness increased with depth to reflect the corresponding
485 decrease in resolution of dispersion velocity with period. Layers were 2.5 km
486 thick down to 60 km depth, 5.0 km thick between 60 and 150 km depth, and
487 10 km thick below 150 km depth. P velocities were obtained from an assumed
488 V_P/V_S ratio of 1.75 down to 40 km depth, and densities were calculated from
489 P velocity through the empirical relationship of Berteussen (1977). The
490 starting model was parameterized down to the top of the transition zone,
491 but only structure down to ~ 250 km depth was inverted. S velocities for the
492 deeper layers were forced to keep PREM-like values during the inversion in
493 order to account for partial sensitivity of long-period dispersion velocities to
494 deep Earth structure (Julià et al., 2003).

495 In order to perform the joint inversion for the seismic stations, we first

grouped the receiver functions within bins of maximum variation in back-azimuth of 10° and maximum variation in ray parameter of 0.01 s/km. If more than three receiver function waveforms were included in the bin, a point-to-point average was computed. Next, a local group-velocity curve was built for the selected station. First, the tomographic cell containing the seismic station was identified in the tomographic grids of Luz et al. (2015) and Dias et al. (2015), and the short- and long-period group-velocities were extracted. Next, the group-velocity curves were merged into a single, continuous curve. Both tomographic studies overlap in the (10-20 s) frequency band, and we opted for regarding the short-period group-velocity measurements from ambient noise, that are generally more reliable (Dias et al., 2015).

S-velocity models for each seismic station were then produced by jointly inverting all receiver function averages (at both high- and low-frequency contents) with the local group-velocity curve. To determine values for the influence (p) and smoothness (θ) parameters, we initially fixed the influence factor to $p = 0.5$ (equal importance to each data set) and varied the smoothness factor between $0.1 \leq \theta \leq 1.9$. The best value was then chosen through analysis of the trade-off curve between RMS misfit and model roughness. Posteriorly, we fixed the smoothness factor and varied the influence factor between $0.1 \leq p \leq 0.9$. Velocity models change little for influence parameters around the selected value, emphasizing the complementarity between the two datasets (Julià et al., 2000). Six iterations were generally enough to converge to a stable model.

The joint inversion procedure is illustrated through select stations in the BP (Figure 7) and the São Francisco craton (Figure 8). In both cases, the

521 match between observations and predictions is good for both receiver function
 522 waveforms and dispersion curves. In the BP, the velocity-depth profile for
 523 station NBPB displays a ~ 40 km thick crust with velocities ranging from
 524 3.5 km/s near the surface to 4.0 km/s at 35 km depth, where they rapidly
 525 increase to 4.4 km/s near the Moho. Upper mantle, velocities are around 4.7
 526 km/s down to 100 km depth. In the São Francisco craton, the velocity-depth
 527 profile for station NBPN shows ~ 44 km thick crust with velocities ranging
 528 from 3.7 km/s near the surface to 3.85 km/s at 35 km depth, where they
 529 gradually increase to 4.4 km/s down to Moho depths.

530 Confidence bounds were developed from velocity-depth profiles after sepa-
 531 rately inverting high- and low-frequency receiver functions within each group
 532 with the dispersion curves for the corresponding seismic station (see e.g. Julià
 533 et al., 2009). For station NBPB, for example, a total of 7 receiver function
 534 groups were available (Figure 7), so up to seven velocity models could be
 535 developed. The standard deviation for each layer (S-velocity) was then com-
 536 puted with respect to the average velocity model developed from jointly
 537 inverting all receiver function group-averages with the dispersion curve. For
 538 station NBPB, this resulted in confidence bounds of ± 0.1 km/s in most
 539 crustal layers and ± 0.2 km/s in the mantle. Uncertainties may grow larger if
 540 lateral variations in Earth velocity around the station are required to explain
 541 features particular to a given receiver function group. We realize that this
 542 approach might be statistically non-rigorous, but we think it still succeeds in
 543 conveying informative ranges in crustal velocity variation due to azimuthal
 544 changes in Earth structure. Finally, depth uncertainties must be at least one
 545 layer thickness.

546 The velocity-depth profiles for the NS transect (BODES stations) are
 547 displayed in Figure 10. Within confidence bounds, our velocity-depth pro-
 548 files display crustal thicknesses consistent with those obtained from the H- κ
 549 stacking analysis (see Figure 9), thus confirming a gradual thickening of the
 550 crust towards the São Francisco craton. The velocity-depth profiles also re-
 551 veal that the crust is generally structured into two layers: (1) a 12-20 km
 552 thick upper crust with S-velocities of 3.5-3.7 km/s; and (2) a 16-23 km thick
 553 lower crust with S-velocities of 3.8-3.9 km/s. In the upper mantle, veloci-
 554 ties are generally 4.5-4.7 km/s immediately below the Moho. We note that
 555 a sediment-bedrock interface at 1.5-2.0 km depth is identified for stations
 556 AR01 in the Iguatu basin and for stations AR50 and AR51, in the Araripe
 557 Basin, with velocities in the 2.4-3.2 km/s range.

558 At mantle depths, we identified a drop in S-velocity (see Figure 12 and
 559 13) that we later interpret as either the lithosphere-asthenosphere boundary
 560 (LAB) or a mid-lithospheric discontinuity (MLD) (Yuan and Romanowicz,
 561 2010). The velocity drop is characterized by a negative velocity gradient over
 562 a finite depth-interval, and a unique depth was attributed at the center of
 563 the interval. We observe that the depth of the gradient changes along the
 564 profile: under the northern tip, in the BP, is at around 135-150 km depth with
 565 minimum S-velocities of 4.5 km/s; in the central portion of the profile, around
 566 the Araripe Basin, it shallows to 120-140 km depth with minimum velocities
 567 of 4.4-4.7 km/s; and under the southern portion, in the São Francisco craton,
 568 it is at 135-145 km depth with S-velocities in the 4.6-4.7 km/s range.

569 Stations AR50 and AR51 deserve special attention, as they are located
 570 within the Araripe Basin. Those stations show the presence of a 10-14 km

571 thick layer of fast (> 4.0 km/s) velocity material at the bottom of the crust
572 (see Figure 10). The velocity-depth profiles for the stations along the EW
573 transect are displayed in Figure 11, and confirm the presence of a 10-14 km
574 thick layer of fast velocity material (~ 4.1 km/s) under the Araripe Basin. As
575 discussed in the next section, S-velocities larger than 4.0 km/s are commonly
576 interpreted as mafic material (e.g. Julià et al., 2005). Moreover, those stations
577 are generally accompanied by a similar high-velocity layer at 5-10 km depth,
578 which might represent shallower intrusions. This might suggest the original
579 crust under the basin was as thin as ~ 30 km, and that it thickened through
580 mafic intrusions to more than 40 km.

581 5. Discussion

582 The most important findings reported in this study are: (i) The pres-
583 ence of a negative velocity gradient at mantle depths of 120-140 km under
584 the Araripe Basin and 135-145 km under the BP and São Francisco craton,
585 and (ii) the presence of a high-velocity (> 4.0 km/s) lower crust under the
586 short-period stations in the Araripe Basin. Next, we discuss the implications
587 of these findings in regard to crustal and lithospheric evolution and basin
588 inversion.

589 5.1. Evidence lacking for secular variation of Precambrian crust

590 Our results on S-velocity structure demonstrate consistent differences at
591 crustal levels between the BP (mainly Proterozoic) and the São Francisco
592 craton (mostly Archean). These show that the crust in the northern São
593 Francisco craton is 41-44 km thick with a mean bulk V_P/V_S ratio around
594 1.7 and an average S-velocity of 3.82 km/s, while the crust in the western

Province is 34-38 km thick with a bulk V_P/V_S ratio around 1.73 and an average S-velocity of 3.79 km/s (see Table 1). The average for the Province does not include the velocity models for the stations located atop the Araripe Basin. Slightly higher V_P/V_S ratios under the Province, mainly beneath Araripe basin, might be suggestive of more mafic lithologies making up the basement crust (Christensen, 1996).

According to the global compilation of Precambrian crustal structure of Durrheim and Mooney (1994), Archean crust is generally thinner (27 - 40 km) than Proterozoic crust (40 - 55 km), and lacks a high-velocity (mafic) layer in the lower crust. The authors argued that higher heat-flow rates in the early Earth resulted in a depleted (low FeO) and refractory lithospheric mantle that prevented the formation of mafic melts to underplate Archean crust; in contrast, a more fertile (FeO rich) and less refractory Proterozoic lithosphere would have allowed the formation of such melts and the underplating of Proterozoic crust. Indeed, a number of studies support the secular evolution of Precambrian crust: in northern Canada, Thompson et al. (2010) reported a structurally simple and felsic-to-intermediate crust for the Rae domain (3.9-2.7 Ga) and a more mafic and structurally complex crust for the slightly younger Hearne domain (2.7 Ga); in eastern Canada Darbyshire et al. (2017) and Petrescu et al. (2016) observed the Proterozoic orogenic crust is generally thicker than Archean crust, invoking mafic underplating of the lower crust to explain the observed gradational Moho; in Australia, Kennett et al. (2011) reported thin crust beneath the Archean cratons in the Pilbara and the southern part of the Yilgarn and thick crust beneath parts of the Proterozoic terrains of Central Australia. Thompson et al. (2010)

620 suggest that secular crustal variation might reflect an evolution from non-
621 plate tectonic style during Paleo- to Mesoarchean times towards modern-style
622 plate tectonics during the Paleoproterozoic. On the other hand, some stud-
623 ies have reported evidence lacking for such a secular evolution. In Africa, an
624 S-velocity model compilation from the joint inversion of receiver functions
625 and dispersion velocities (Tugume et al., 2013) displayed no clear difference
626 in crustal structure between Archean and Proterozoic terranes; and, glob-
627 ally, Rudnick and Fountain (1995) found no significant variation in crustal
628 thickness among Archean and Proterozoic crust.

629 Our results demonstrate that the crust of the (northern) São Francisco cra-
630 ton is thicker than that of the neighboring (Proterozoic) Borborema Province,
631 and that a basal, high-velocity layer is thin to non-existent in both domains.
632 This contrasts with the tectonic age correlation reported in Central Brazil
633 by Assumpção et al. (2002), in which Archean crust (including that of the
634 southern São Francisco craton) ranges in thickness from 37 to 43 km while
635 Proterozoic crust ranges from 40 to 47 km. However, we consider those thick-
636 ness variations questionable, as the thickness ranges seem to be quite large
637 and overlapping and the reported absence of mafic underplating beneath the
638 southern craton conflicts with a $V_p > 7.0$ km/s in the lower crust displayed
639 in some of the corresponding velocity models.

640 Nonetheless, Phanerozoic processes might have significantly altered crustal
641 thickness and composition in the Borborema Province. The formation of
642 several syn-rift basins (including the Araripe Basin) during Mesozoic times
643 clearly attests towards extensional stresses stretching the crust and litho-
644 sphere (e.g. de Matos, 1999), and Cenozoic volcanism in the eastern Province

645 (Mizusaki et al., 2002) might also have altered crustal composition. Luz et al.
646 (2015) suggested that the crust under the Borborema Plateau might have
647 been relatively unaffected by Mesozoic extension, but Phanerozoic tectonics
648 renders it questionable whether the Borborema crust as a whole is a good
649 proxy for remnant Proterozoic crust.

650 5.2. *Lithospheric delamination under Araripe?*

651 Our results also allow for an investigation of changes in lithospheric struc-
652 ture along the profile, although the definition of the lithosphere-asthenosphere
653 boundary (LAB) in continental cratons is not straight-forward (Eaton et al.,
654 2009; Darbyshire and Eaton, 2010). As noted above, a drop in S-velocity
655 is observed at depths of 120-140 km under the Araripe Basin; and, if the
656 anomalous lithosphere under stations NBPB and AR01 is not considered,
657 we observe a similar velocity drop beneath both Archean and Proterozoic
658 domains at 135-145 km depth (Figure 14). Interpreting the depth of the neg-
659 ative gradient as the LAB under the entire profile makes our results broadly
660 consistent with those of the collocated MT experiment (Garcia et al., 2019).
661 Although no LAB was observed along the entire profile, craton-like resistiv-
662 ities were reported under both Archean and Proterozoic domains down to
663 140-150 km depth and, more importantly, a prominent conductive anomaly
664 starting at about 120 km depth under the Araripe Basin was interpreted as
665 evidence of thin lithosphere.

666 This interpretation, however, conflicts with lithosphere thicknesses of
667 more than 200 km postulated in other cratons worldwide (Afonso et al.,
668 2019; Steinberger and Becker, 2018; Bastow et al., 2015). Although some
669 compilations suggest South American cratons might not have a well-formed

LAB (Bastow et al., 2015), another possibility would be that the negative gradient corresponds to a mid-lithospheric discontinuity (MLD). According to Yuan and Romanowicz (2010), the MLD would be separating an old, highly depleted chemical layer of variable thickness from a younger, less depleted deeper layer accreted at a later stage and bounded at the bottom by a relatively flat LAB. This could indeed be the case under the São Francisco craton, where the fast mantle lid and the marked gradient would be consistent with a depleted (refractory), old lithosphere and a sharp transition to a younger lithosphere, respectively. A similar interpretation would also be possible for the Borborema Province (north of the Araripe Basin), as long as the existence of a MLD were extended to Paleoproterozoic lithosphere. Indeed, relatively fast velocities at lower lithospheric levels are observed in the tomographic study of Simões Neto et al. (2019) immediately north of the Araripe basin, which would be consistent with this idea. Detailed investigations of the Canadian shield (Darbyshire et al., 2013; Liddell et al., 2017), however, show that a layered cratonic lithosphere is confined to the Archean domains of Arctic Canada. Moreover, fast velocities at a lower lithosphere level do not seem apparent under Proterozoic domains globally (e.g. Ritsema et al., 2010), implying that this extension might be unlikely on a global scale and quite specific to the Borborema Province.

Finally, Garcia et al. (2019) postulated the presence of a small-scale convection cell under the Araripe basin coinciding with the locally thin lithosphere imaged through analysis of collocated MT observations. They further postulated that such small-scale convection cell would have triggered a roll-down process (Liu and Shen, 1998) by which passive upwelling of the

asthenosphere under the basin would have thinned the overlying lithosphere and accreted to the lithosphere on each side. The accretion of the eroded material would have effectively thickened the lithosphere immediately north and south off the basin, giving the impression of craton-like lithosphere under the Proterozoic domains of the BP. Moreover, the proximity of the 120-140 km LAB under the Araripe Basin to the depth for the postulated MLD immediately to the north and south, further suggests that the younger lithospheric layer could have been eroded away from under the Araripe Basin, bringing the original Archean lithosphere back into contact with the underlying asthenosphere.

5.3. *Implications for basin inversion*

Taken at face value, our estimates of crustal thickness display a general thickening trend towards the São Francisco craton with no apparent thinning under the Araripe Basin. Interestingly, our S-velocity models demonstrate that this lack of apparent thinning is due to the presence of a relatively thick ($\sim 10\text{-}14$ km), high-velocity layer (> 4.1 km/s) at the bottom of the crust (Figures 10, 11 and 14). These high velocities might represent mafic material in the lowermost crust under the basin (White et al., 2008). Indeed, S-velocities of 4.0 km/s or higher have been interpreted in other Precambrian terrains as diagnostic of mafic material (e.g. Julià et al., 2005; Tugume et al., 2013), and a V_P/V_S of 1.80 for the crust under most of the stations within the basin boundaries is suggestive of a bulk mafic composition (Christensen, 1996) (see Table 1).

A recent study based on aeromagnetic data (Camacho and Ferreira, 2017) proposed an updated structural framework for the Araripe Basin, consistent

720 with previous works (e.g. Ponte, 1996; de Matos, 1992; Assine, 2007; de Cas-
721 tro and Branco, 1999), that includes up to four internal domains: Araripe-
722 Araripina, Feitora, Crato-Juazeiro and Mauriti. Perhaps more importantly,
723 the survey showed four magnetic anomalies of high amplitude and semicircu-
724 lar form localized very close to station AR51, and just a few kilometers away
725 from stations AR50 and AR03. These anomalies were considered robust by
726 the authors, because even the overlying sedimentary cover (~ 1.500 m) was
727 unable to attenuate the magnetic signal. The magnetic anomalies represent
728 evidence of bodies with high magnetic susceptibility at depth, which may be
729 associated with magmatic intrusions (Camacho and Ferreira, 2017).

730 As mentioned in the previous section, Garcia et al. (2019) had reported
731 a locally thin lithosphere of ~ 120 km under the Araripe Basin, between the
732 Patos and Pernambuco lineament. A marked resistivity anomaly of $25 \Omega m$
733 was interpreted as indication of shallow, enriched asthenospheric material,
734 involving small amounts of fluids and melt, with temperatures around 1350
735 $^{\circ}C$. Moreover, the authors proposed that the buoyancy of the asthenospheric
736 material would be providing a vertical push to the overlying lithosphere that
737 might be aiding the tectonic inversion of the basin driven by the regional
738 stresses. The presence of a lowermost crust under the basin's sedimentary
739 cover reported here could be the result of melts formed in the underlying
740 lithospheric mantle, triggered by heating from this thermal anomaly, and
741 could have provided additional buoyancy for local uplift of the basin.

742 Our results thus support and complement local tectonic inversion of the
743 Araripe Basin, as proposed in Marques et al. (2014). These authors pro-
744 posed that the inversion of the Araripe Basin was the result of horizontal

745 compressive stresses originating from the development of the South Atlantic
746 to the east, and the Andean chain to the west, reactivating old high-angle
747 normal faults and creating new, long-angle thrust faults. Moreover, the au-
748 thors acknowledge that their model would work best if local necking of the
749 lithosphere was present under the basin, as that would induce differential
750 crustal thinning and increase post-rift crustal weakness. As mentioned ear-
751 lier, local necking of the lithosphere was reported in Garcia et al. (2019) and
752 is consistent with depths to the negative seismic velocity gradient imaged in
753 this study.

754 Nonetheless, we think that a contribution from regional epeirogeny and
755 differential erosion - as suggested in Peulvast and Bétard (2015) - cannot be
756 completely ruled out, as that would make it difficult to explain the present
757 morphology of the regional landscape. We do rule out, however, that the
758 postulated regional epeirogeny of the Province be permanent in character
759 and related to mafic underplating. Our results demonstrate that a mafic
760 lowermost crust is localized under the basin, which is unlikely to provide
761 enough buoyancy to generate a crustal upwarp of regional dimensions. Recent
762 tomographic images (Simões Neto et al., 2019) have revealed the presence of a
763 broad, NE-SW trending sublithospheric channel crossing the entire BP, which
764 might represent a more plausible source of buoyancy. Indeed, a non-zero free-
765 air anomaly of the regional gravity field reported in Oliveira and de Medeiros
766 (2012) would be consistent with a more dynamic (uncompensated) character
767 for the regional elevation of the Province.

768 6. Conclusions

769 We have investigated the lithospheric structure under the Araripe Basin
770 and neighboring regions by jointly inverting receiver functions and surface-
771 wave dispersion in the region. Our results have demonstrated the lithosphere
772 under the basin is thin (~ 120 km) when compared to the regions flanking the
773 basin and that the basin's underlying lower crust includes 10-14 km of mafic
774 intrusions. Local tectonic inversion of the Araripe Basin is thus likely to
775 have resulted from a combination of stress concentration on thin lithosphere,
776 thermal buoyancy from a shallow asthenosphere - as pointed out by Garcia
777 et al. (2019) - and compositional buoyancy from a local mafic lowermost
778 crust. Regional uplift and differential erosion, however, cannot be ruled
779 out. Tectonic (Marques et al., 2014) and topographic (Peulvast and Bétard,
780 2015) inversion of the Araripe Basin should then be regarded as end-member
781 models, the relative contribution of which is yet to be assessed.

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References

- Afonso, J., Salajeghegh, F., Szwillus, W., Ebbing, J., Gaina, C., 2019. A global reference model of the lithosphere and upper mantle from joint inversion and analysis of multiple data sets. *Geophys. J. Int.* 169. doi:10.1093/gji/ggz094.
- Almeida, F.F.M., Hasui, Y., Brito Neves, B.B., Fuck, R.A., 1981. Brazilian structural provinces: An introduction. *Earth Science Reviews* 17, 1–29. doi:10.1016/0012-8252(81)90003-9.
- Almeida, Y.B., Julià, J., Frassetto, A., 2015. Crustal architecture of the Borborema Province, NE Brazil, from receiver function CCP stacks: Implications for Mesozoic stretching and Cenozoic uplift. *Tectonophysics* 649, 68–80. doi:10.1016/j.tecto.2015.03.001.
- Ammon, C.J., Randall, G.E., Zandt, G., 1990. On the nonuniqueness of receiver function inversions. *J. Geophys. Res.* 95, 15303–15318. doi:10.1029/JB095iB10p15303.
- Ganade de Araujo, C.E., Weinberg, R., Cordani, U., 2013. Extruding the Borborema Province (NE-Brazil): a two-stage Neoproterozoic collision process. *Terra Nova*, 1–12. doi:10.1111/ter.12084.
- Assine, M., 2007. Araripe basin [Bacia do Araripe]. *Boletim de Geociências da Petrobras* 15, 371–389.
- Assumpção, M., James, D., Snoke, A., 2002. Crustal thicknesses in SE Brazilian Shield by receiver function analysis: Implications for isostatic compensation. *J. Geophys. Res.* 107. doi:10.1029/2001JB000422.

- 815 Bastow, I., Julia, J., do, N.A., Fuck, R., Buckthorp, T., McClellan, J., 2015.
816 Upper mantle anisotropy of the Borborema Province, NE Brazil: Impli-
817 cations for intra-plate deformation and sub-cratonic asthenospheric flow.
818 *Tectonophysics* 657, 81–93. doi:10.1016/j.tecto.2015.06.024.
- 819 Berteussen, K.A., 1977. Moho depth determinations based on spectral ratio
820 analysis of NORSAR long-period p waves. *Phys. Earth Planet. Inter.* 15,
821 13–27. doi:10.1016/0031-9201(77)90006-1.
- 822 Bianchi, I., Park, J., Agostinetti, N., Levin, V., 2010. Mapping seismic
823 anisotropy using harmonic decomposition of receiver functions: An ap-
824 plication to Northern apennines, Italy. *J. Geophys. Res.*, 115, 1–14.
825 doi:10.1029/2009JB00706.
- 826 Bonini, M., Sani, F., Antonielli, B., 2011. Basin inversion and contractional
827 reactivation of inherited normal faults: A review based on previous and
828 new experimental models. *Tectonophysics* 522–523, 55–88. doi:10.1016/
829 j.tecto.2011.11.014.
- 830 Brito Neves, B., Fuck, R., Cordani, U., Thomaz, F., 1984. Influence of
831 basement structures on the evolution of the major sedimentary basins of
832 Brazil: A case of tectonic heritage. *Journal of Geodynamics* 1, 495–510.
833 doi:10.1016/0264-3707(84)90021-8.
- 834 Brodie, J., White, N., 1995. The link between sedimentary basin inversion
835 and igneous underplating. Geological Society, London, Special Publica-
836 tions 88, 21–38. doi:10.1144/GSL.SP.1995.088.01.03.

- 837 Buchanan, J., Buchanan, P., 1995. Basin Inversion. Geological Society,
838 London, Special Publications 88, 7–9. doi:10.1144/GSL.SP.1995.088.
839 01.30.
- 840 Camacho, C., Ferreira, F.R., 2017. The structural framework of the Araripe
841 Sedimentary Basin, Borborema Province, based on aeromagnetic data.
842 Geol. USP, Sér. cient. 17, 149–161. doi:10.11606/issn.2316-9095.
843 v17-393.
- 844 Christensen, N.I., 1996. Poisson’s ratio and crustal seismology. J. Geophys.
845 Res. 101, 3139–3156. doi:10.1029/95JB03446.
- 846 Christensen, N.I., Mooney, W.D., 1995. Seismic velocity structure and com-
847 position of the continental crust: A global view. J. Geophys. Res.: Solid
848 Earth 100, 9761–9788. doi:10.1029/95JB00259.
- 849 Darbyshire, F.A., Bastow, I.D., Petrescu, L., Gilligan, A., Thompson, D.A.,
850 2017. A tale of two orogens: Crustal processes in the Proterozoic Trans-
851 Hudson and Grenville Orogens, eastern Canada. Tectonics. 36, 1633–1659.
852 doi:10.1002/2017TC004479.
- 853 Darbyshire, F.A., Eaton, D.W., 2010. The lithospheric root beneath Hudson
854 Bay, Canada from Rayleigh wave dispersion: No clear seismological dis-
855 tinction between Archean and Proterozoic mantle. Lithos. 120, 144–159.
856 doi:10.1016/j.lithos.2010.04.010.
- 857 Darbyshire, F.A., Eaton, D.W., Bastow, I.D., 2013. Seismic imaging of the
858 lithosphere beneath Hudson Bay: Episodic growth of the Laurentian man-

- 859 tle keel. *Earth Planet. Sci. Lett.* 373, 179–193. doi:10.1016/j.epsl.2013.
860 05.002.
- 861 de Castro, D., Branco, R.M., 1999. Caracterização da arquitetura in-
862 terna das bacias do Vale do Cariri (NE do Brasil) com base em
863 modelagem gravimétrica 3-D. *Rev. Bras. Geof.* 17. doi:10.1590/
864 S0102-261X1999000200003.
- 865 de Castro, D., Medeiros, W., 1997. Afinamento crustal associado à implan-
866 tação das bacias intracratônicas do trend Cariri-Potiguar. XVII Simp Geol
867 do Nordeste, Fortaleza, Anais. , 335–338.
- 868 de Matos, R.M.D., 1992. The Northeast Brazilian rift system. *Tectonics* 11,
869 766–791. doi:10.1029/91TC03092.
- 870 de Matos, R.M.D., 1999. History of the northeast Brazilian rift system:
871 kinematic implications for the break-up between Brazil and West Africa.
872 Geological Society, London, Special Publications 153, 55–73. doi:10.1144/
873 GSL.SP.1999.153.01.04.
- 874 Dias, R.C., Julià, J., Schimmel, M., 2015. Rayleigh-wave, group-velocity
875 tomography of the Borborema Province, NE Brazil, from ambient
876 seismic noise. *Pure Appl. Geophys.* 172, 1429–1449. doi:10.1007/
877 s00024-014-0982-9.
- 878 Durrheim, R.J., Mooney, W.D., 1994. Evolution of the precambrian litho-
879 sphere: Seismological and geochemical constraints. *J. Geophys. Res.* 99,
880 15359–15374. doi:10.1029/94JB00138.

881 Dziewonski, A., Bloch, S., Landisman, M., 1969. A technique for analysis of
882 transient seismic signals. *Bull. Seismol. Soc.* 59, 427–444.

883 Eaton, D., Darbyshire, F., Evans, R., Grütter, H., Jones, A., Yuan, X., 2009.
884 The elusive lithosphere-asthenosphere boundary (LAB) beneath cratons.
885 *Lithos.* 109, 1–22. doi:10.1016/j.lithos.2008.05.009.

886 Efron, B., Tibshirani, R., 1991. Statistical Data Analysis in the Computer
887 Age. *Science* 253, 390–395. doi:10.1126/science.253.5018.390.

888 Feng, M., Assumpção, M., Van der Lee, S., 2004. Group-velocity to-
889 mography and lithospheric S-velocity structure of the South American
890 continent. *Physics of the Earth and Planetary Interiors* 147, 315–331.
891 doi:10.1016/j.pepi.2004.07.008.

892 Frassetto, A., Zandt, G., Gilbert, H., Owens, T.J., Jones, C.H., 2010.
893 Improved imaging with phase-weighted common conversion point stacks
894 of receiver functions. *Geophys. J. Int.* 182, 368–374. doi:10.1111/j.
895 1365-246X.2010.04617.x.

896 Garcia, X., Julià, J., Nemocón, A.M., Neukirch, M., 2019. Lithospheric
897 thinning under the Araripe Basin (NE Brazil) from a long-period magne-
898 totelluric survey: Constraints for tectonic inversion. *Gondwana Research*
899 68, 174–184. doi:10.1016/j.gr.2018.11.013.

900 Gilbert, H.J., Sheehan, A.F., 2004. Images of crustal variations in the
901 intermountain west. *J. Geophys. Res.* 109, B03306. doi:10.1029/
902 2003JB002730.

- 903 Hansen, D.L., Nielsen, S.B., 2003. Why rifts invert in compression. 373, 5–24.
904 doi:10.1016/S0040-1951(03)00280-4.
- 905 Hegarty, K., Morais Neto, J., Karner, G., 2002. The enigma of the Araripe
906 plateau: new constraints on its uplift and tectonic history using AFTA.
907 SBG, Congresso Brasileiro de Geologia, 41, João Pessoa, Anais. , 495.
- 908 Herrin, E., Goforth, T., 1977. Phase-matched filters: Application to the
909 study of Rayleigh waves. Bull. Seismol. Soc. 67, 1259–1275.
- 910 Julià, J., 2007. Constraining velocity and density contrasts across the crust
911 mantle boundary with receiver function amplitudes. Geophys. J. Int. 171,
912 286–301. doi:10.1111/j.1365-2966.2007.03502.x.
- 913 Julià, J., A. NYBLADE, A., Durrheim, R., Linzer, L., Gok, R., Walter,
914 W., Spottiswoode, S., Dirks, P., 2009. A Wadati filter for mine-induced
915 seismicity. South African Journal of Geology - S AFR J GEOL 112, 371–
916 380. doi:10.2113/gssajg.112.3-4.371.
- 917 Julià, J., Ammon, C.J., Herrmann, R.B., 2003. Lithospheric structure
918 of the arabian shield from the joint inversion of receiver functions and
919 surface-wave group velocities. Tectonophysics 371, 1–21. doi:10.1016/
920 S0040-1951(03)00196-3.
- 921 Julià, J., Ammon, C.J., Herrmann, R.B., Correig, A.M., 2000. Joint inversion
922 of receiver function and surface wave dispersion observations. Geophys. J.
923 Int. 143, 99–112. doi:10.1046/j.1365-246x.2000.00217.x.
- 924 Julià, J., J. Ammon, C., A. Nyblade, A., 2005. Evidence for mafic lower
925 crust in Tanzania, East Africa, from joint inversion of receiver functions

926 and Rayleigh wave dispersion velocities. *Geophys. J. Int.* 162, 555–569.
927 doi:10.1111/j.1365-246X.2005.02685.x.

928 Julià, J., Mejia, J., 2004. Thickness and V_p/V_s ratio variation in the Iberian
929 crust. *Geophys. J. Int.* 156, 59–72. doi:10.1111/j.1365-246X.2004.
930 02127.x.

931 Julià, J., Vila, J., Macià, R., 1998. The receiver structure beneath the Ebro
932 basin, Iberian Peninsula. *Bull. Seismol. Soc. Am.* 88, 1538–1547.

933 Kennett, B., Salmon, M., Saygin, E., 2011. AusMoho: the variation of
934 Moho depth in Australia. *Geophys. J. Int.* 187, 946–958. doi:10.1111/j.
935 1365-246X.2011.05194.x.

936 Knesel, K.M., Souza, Z.S., Vasconcelos, P.M., Cohen, B.E., , Silveira., F.V.,
937 2011. Young volcanism in the Borborema Province, NE Brazil, shows no
938 evidence for a trace of the Fernando de Noronha plume on the continent.
939 *Earth Planet. Sci. Lett.* 302, 38–50. doi:10.1016/j.epsl.2010.11.036.

940 Langston, C.A., 1979. Structure under Mount Rainier, Washington, inferred
941 from teleseismic body waves. *J. Geophys. Res.* 84, 4749–4762. doi:10.
942 1029/JB084iB09p04749.

943 Levin, V., Park, J., 1997. P-SH conversions in a flat-layered medium with
944 anisotropy of arbitrary orientation. *Geophys. J. Int.* 131, 253–266. doi:10.
945 1111/j.1365-246X.1997.tb01220.x.

946 Liddell, M.V., Bastow, I.D., Darbyshire, F.A., Gilligan, A., S., P., 2017. The
947 formation of Laurentia: Evidence from shear wave splitting. *Earth Planet.*
948 *Sci. Lett.* 479, 170–178. doi:10.1016/j.epsl.2017.09.030.

- 949 Ligorria, J.P., Ammon, C.J., 1999. Iterative deconvolution and receiver-
950 function estimation. *Bulletin of the Seismological Society of America* 89,
951 1395–1400.
- 952 Liu, M., Shen, Y., 1998. Sierra Nevada uplift: A ductile link to mantle
953 upwelling under the Basin and Range province. *Geology* 26, 299–302.
954 doi:10.1130/0091-7613(1998)026<0299:SNUADL>2.3.CO;2.
- 955 Luz, R.M.N., Julià, J., do Nascimento, A.F., 2015. Bulk crustal properties of
956 the Borborema Province, NE Brazil, from P-wave receiver functions: Im-
957 plications for models of intraplate Cenozoic uplift. *Tectonophysics* 644–645,
958 81–91. doi:10.1016/j.tecto.2014.12.017.
- 959 Marques, F.O., Nogueira, F.C.C., Bezerra, F.H.R., de Castro, D., 2014. The
960 Araripe Basin in NE Brazil: An intracontinental graben inverted to a
961 high-standing horst. *Tectonophysics* 630, 251–264. doi:10.1016/j.tecto.
962 2014.05.029.
- 963 Mizusaki, A.M.P., Thomaz-Filho, A., Milani, E.J., Césero., P., 2002. Meso-
964 zoic and Cenozoic igneous activity and its tectonic control in northeastern
965 Brazil. *J. South Am. Earth Sci.* 15, 183–198.
- 966 Morais Neto, J., Hegarty, K., Karner, G., Alkmim, F., 2009. Timing and
967 mechanisms for the generation and modification of the anomalous to-
968 pography of the Borborema Province, northeastern Brazil. *Marine and*
969 *Petroleum Geology* 26, 1070–1086. doi:10.1016/j.marpetgeo.2008.07.
970 002.

- 971 Simões Neto, F., Julià, J., Schimmel, M., 2019. Upper-mantle structure
972 of the Borborema Province, NE Brazil, from P-wave tomography: im-
973 plications for rheology and volcanism. *Geophys. J. Int.* 216, 231–250.
974 doi:10.1093/gji/ggy421.
- 975 Ogden, C.S., Bastow, I.D., Gilligan, A., Rondenay, S., 2019. A Reappraisal
976 of the H- κ Stacking Technique: Implications for Global Crustal Structure.
977 *Geophys. J. Int.* 219, 1491–1513. doi:10.1093/gji/ggz364.
- 978 Oliveira, R.G.d., de Medeiros, W.E., 2012. Evidences of buried loads in
979 the base of the crust of Borborema Plateau (NE Brazil) from Bouguer
980 admittance estimates. *Journal of South American Earth Sciences* 37, 60–
981 76. doi:10.1016/j.jsames.2012.02.004.
- 982 Owens, T.J., Zandt, G., Taylor, S.R., 1984. Seismic evidence for an an-
983 cient rift beneath the Cumberland Plateau, Tennessee: A detailed analysis
984 of broadband teleseismic P waveforms. *J. Geophys. Res.* 89, 7783–7795.
985 doi:10.1029/JB089iB09p07783.
- 986 Petrescu, L., Bastow, I., Darbyshire, F., Gilligan, A., Bodin, T., Menke, W.,
987 Levin, V., 2016. Three billion years of crustal evolution in eastern Canada:
988 Constraints from receiver functions. *J. Geophys. Res. Solid Earth.* 121.
989 doi:10.1002/2015JB012348.
- 990 Peulvast, J.P., Bétard, F., 2015. A history of basin inversion, scarp retreat
991 and shallow denudation: The Araripe basin as a keystone for understand-
992 ing long-term landscape evolution in NE Brazil. *Geomorphology* 233, 20–
993 40. doi:10.1016/j.geomorph.2014.10.009.

- 994 Ponte, F.C. and Ponte Filho, F., 1996. Estrutura geológica e evolução tec-
995 tônica da Bacia do Araripe. Recife: DNPM. 17, 68.
- 996 Rawlinson, N., Sambridge, M., 2005. The Fast Marching Method: An Ef-
997 fective Tool for Tomographic Imaging and Tracking Multiple Phases in
998 Complex Layered Media . Exploration Geophysics 36, 341–350. doi:10.
999 1071/EG05341.
- 1000 Ritsema, J., Deuss, A., van Heijst, H.J., Woodhouse, J.H., 2010. S40RTS:
1001 a degree-40 shear-velocity model for the mantle from new Rayleigh wave
1002 dispersion, teleseismic traveltime and normal-mode splitting function mea-
1003 surements. Geophys. J. Int. 184, 1223–1236. doi:10.1111/j.1365-246X.
1004 2010.04884.x.
- 1005 Rudnick, R., Fountain, D., 1995. Nature and composition of the continental
1006 crust: A lower crustal perspective. Reviews of Geophysics. 33, 267–309.
1007 doi:10.1029/95RG01302.
- 1008 Jardim de Sá, E.F., Da Trindade, R.I.F., de Hollanda, M.H.B.M., Araújo,
1009 J.M.M., Galindo, A.C., Amaro, V.E., de Souza, Z.S., Vignerese, J.L.,
1010 Lardeaux, J.M., 1999. Brasiliano Syntectonic Alkaline Granites Emplaced
1011 in a Strike Slip/Extensional Setting (Eastern Seridó Belt, NE Brazil).
1012 Anais da Academia Brasileira de Ciencias 71, x2–27.
- 1013 Sandiford, M., 1999. Mechanics of basin inversion. Tectonophysics 305, 109–
1014 120. doi:10.1016/S0040-1951(99)00023-2.
- 1015 Schimmel, M., Paulssen, H., 1997. Noise reduction and detection of weak,

- coherent signals through phase-weighted stacks. *Geophys. J. Int.* 130, 497–505. doi:10.1111/j.1365-246X.1997.tb05664.x.
- Steinberger, B., Becker, T.W., 2018. A comparison of lithospheric thickness models. *Tectonophysics* 746, 325–338. doi:10.1016/j.tecto.2016.08.001.
- Thompson, D., Bastow, I., Helffrich, G., Kendall, J.M., Wookey, J., Snyder, D., Eaton, D., 2010. Precambrian crustal evolution: Seismic constraints from the Canadian Shield. *Earth Planet. Sci. Lett.* 297. doi:10.1016/j.epsl.2010.07.021.
- Tugume, F., Nyblade, A., Julià, J., van de Meijde M., 2013. Precambrian crustal structure in Africa and Arabia: evidence lacking for secular variation. *Tectonophysics* 609, 250–266.
- White, R.S., Smith, L.K., Roberts, A.W., Christie, P.A., Kusznir, N.J., 2008. Lower-crustal intrusion on the North Atlantic continental margin. *Nature*. 452, 460–464. doi:10.1038/nature06687.
- Yuan, H., Romanowicz, B., 2010. Lithospheric layering in the North American craton. *Nature*. 466, 1063–1069. doi:10.1038/nature09332.
- Zhu, L., Kanamori, H., 2000. Moho depth variation in southern California from teleseismic receiver functions. *J. Geophys. Res.* 105, 2969–2980. doi:10.1029/1999JB900322.

STATION	LAT	LON	ALT(m a.s.l.)	# RF	a	Vp (km/s)	W1	W2	W3	H	$\epsilon(H)$	Vp/Vs	$\epsilon(Vp/Vs)$	Corr%
NBPB	-5.5432	-39.5836	260	40	2.5	6.5	0.5	0.5	0	37.1	0.8	1.73	0.02	93.1
AR01	-6.1682	-39.4098	305	30	2.5	6.5	0.4	0.3	0.3	38.3	1.1	1.64	0.03	85.9
AR02	-6.6419	-39.3402	341	37	2.5	6.5	0.5	0.5	0	34.6	0.4	1.75	0.02	85.8
AR03	-7.0466	-39.4958	479	40	1.0	6.5	0.4	0.3	0.3	36.6	0.8	1.71	0.03	95.4
AR50	-7.1957	-39.5463	896	20	1.0	6.5	0.4	0.3	0.3	38.1	0.6	1.73	0.03	93.6
AR51	-7.3861	-39.6255	890	10	1.0	6.5	0.4	0.3	0.3	38.3	1.2	1.8	0.03	97.6
AR04	-7.5758	-39.6746	493	38	2.5	6.5	0.4	0.3	0.3	39.5	0.5	1.75	0.02	86.8
AR05	-8.0932	-39.8983	441	40	2.5	6.5	0.4	0.3	0.3	41	0.8	1.73	0.03	97.2
AR06	-8.6305	-40.2249	454	39	2.5	6.5	0.4	0.3	0.3	44.2	0.4	1.68	0.01	92.7
AR08	-9.1450	-40.3727	394	9	2.5	6.5	0.4	0.3	0.3	44.5	1.3	1.7	0.03	76.9
AR07	-9.6154	-40.3837	405	40	2.5	6.5	0.4	0.3	0.3	43.4	0.7	1.72	0.02	91.9
AR09	-10.2136	-40.1954	678	12	2.5	6.5	0.4	0.3	0.3	43.6	1.0	1.73	0.04	87.3
NBPN	-10.8468	-40.1988	387	40	2.5	6.5	0.4	0.3	0.3	44.6	1.1	1.68	0.02	88.5
SERO	-7.5000	-40.2555	749	21	1.0	6.5	0.4	0.3	0.3	36.2	2.2	1.8	0.06	98.6
TBCA	-7.4090	-39.8395	861	20	1.0	6.5	0.5	0.5	0.0	39.6	2.2	1.75	0.06	83.1
BBLH	-7.3672	-39.3447	924	17	2.5	6.5	0.5	0.5	0.0	39.8	0.8	1.78	0.04	89.1
NBMA	-7.3650	-38.7640	437	23	2.5	6.5	0.4	0.3	0.3	37.2	0.7	1.78	0.03	91.9

Table 1: Station coordinates and H - κ stacking parameters.

Figure Captions

Fig. 1. Geologic map of the BP with the seismic stations of the BODES experiment (red triangles) and the Precambrian network of shear zones (solid lines) superimposed.

Fig. 2. Global map seismic sources selected for this study (red circles) and stations of the BODES Experiment (black triangles).

Fig. 3. Sample receiver function stacks for station NBP. Each panel displays two traces corresponding to the radial (top) and transverse (bottom) receiver functions, respectively. The stacks are presented as a black line within a grey shade representing 1σ -confidence bounds for the stacked amplitudes. The number of waveforms, average back-azimuth, and average ray parameter (in s/km) is indicated on top of each panel.

Fig. 4. Sample H- κ stacking results for stations AR04 (top) and NBP (bottom). The right panels show the corresponding radial receiver functions used during the stacking procedure, sorted by ray parameter, while the left panels show the H- κ stacking surfaces. The thin red lines in the right panels are the phase moveout curves for the Ps, PpPs and PpSs + PsPs phases associated with the maximum in the H- κ stacking surface. The thick red line in the left panels is the 1σ confidence ellipse from bootstrapping. Depth windows are 10 km wide and centered around the selected maximum in the H- κ stacking surface to facilitate comparison.

Fig. 5. Summary of H- κ results for the BODES Experiment stations. Crustal thickness is indicated in the middle of the blue bands (in km) and V_P/V_S ratios are indicated at the bottom. Detailed H- κ stacking parameters can be found in Table 1.

1061 **Fig. 6.** Receiver function CCP-stack along the BODES Experiment
1062 seismic line, showing clear P to S conversions at the crust-mantle boundary,
1063 as delineated by positive receiver function amplitudes.

1064 **Fig. 7.** Detailed joint inversion results for station NBPB. Observed
1065 (black) and predicted (red) receiver functions at high- and low-frequency
1066 contents are displayed on the left-hand panels. Observed (red) and predicted
1067 (triangles) dispersion velocities are displayed in the upper-right panel. The
1068 inverted (red) and starting (blue) velocity models are displayed in the lower-
1069 right corner, along with the 2σ confidence bounds (grey shade).

1070 **Fig. 8.** Same legend as in Figure 7, for station NBPB.

1071 **Fig. 9.** Comparison of crustal thicknesses from H- κ stacking and joint
1072 inversion. Crustal thickness from the joint inversion models was obtained
1073 from the top of the first layer above a threshold S-velocity of 4.3 km/s;
1074 crustal thickness from the H- κ stacking was taken from the Gaussian width
1075 that gave the closest estimate to that from the corresponding joint inversion
1076 model. Confidence bounds for the joint inversion estimates were set to ± 2.5
1077 km (i.e. $\pm$ one layer thickness). Note the general consistency between both
1078 sets of estimates.

1079 **Fig. 10.** Velocity-depth profiles for each of the BODES Experiment
1080 NE stations. Each panel displays the inverted model (black) with the 2σ -
1081 confidence bounds (grey shade) and the starting PREM model (blue). Moho
1082 depth (red line) is noted with ± 2.5 km confidence bounds (red shade). H-
1083 κ stacking results for crustal thickness are superimposed as a black error
1084 bar. Note the presence of a thick layer of fast-velocity material at the top
1085 and bottom of the crust under station AR51. The relatively high-velocity at

1086 10-20 km depth under station AR09 is likely to be an artifact.

1087 **Fig. 11.** Velocity-depth profiles for each of the BODES Experiment WE
1088 stations. Same legend as in Figure 10. Note the presence of a thick layer of
1089 fast-velocity material at the bottom of the crust under all stations of Araripe
1090 basin, which is generally accompanied by a similar fast-velocity layer at 5-10
1091 km depth.

1092 **Fig. 12.** Velocity-depth profiles for each of the BODES Experiment
1093 stations down to sub-lithospheric depths. Each panel displays the inverted
1094 model (black) with the 2σ -confidence bounds (grey shade) and PREM model
1095 (blue). The crust-mantle boundary (red) and the LVZ depth (magenta) are
1096 indicated for each velocity-depth profile.

1097 **Fig. 13.** Velocity-depth profiles for each of the BODES Experiment
1098 WE-stations down to sub-lithospheric depths. Same legend as in Figure 12.

1099 **Fig. 14.** Pseudo-2D cross-section constructed through cubic interpo-
1100 lation of all 1D BODES Experiment velocity-depth profiles, down to sub-
1101 lithospheric depths. Moho depths and interpreted MLD and/or LAB discon-
1102 tinuities are superimposed.

1103 Figures

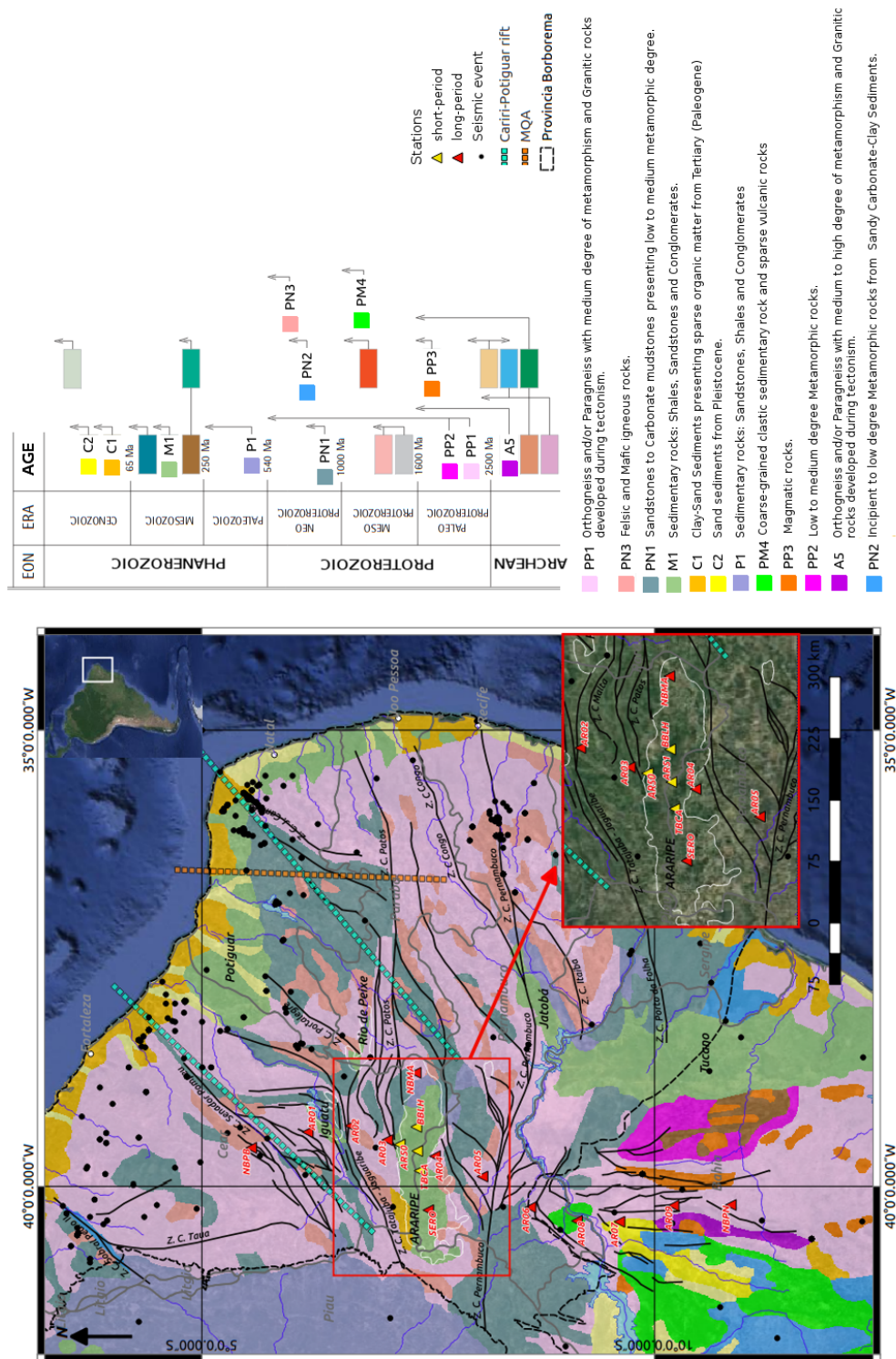


Figure 1: Geologic map of the BP (font, IBGE 2019) with the seismic stations of the BODES Experiment (red triangles) and the Precambrian network of shear zones (solid lines) superimposed.

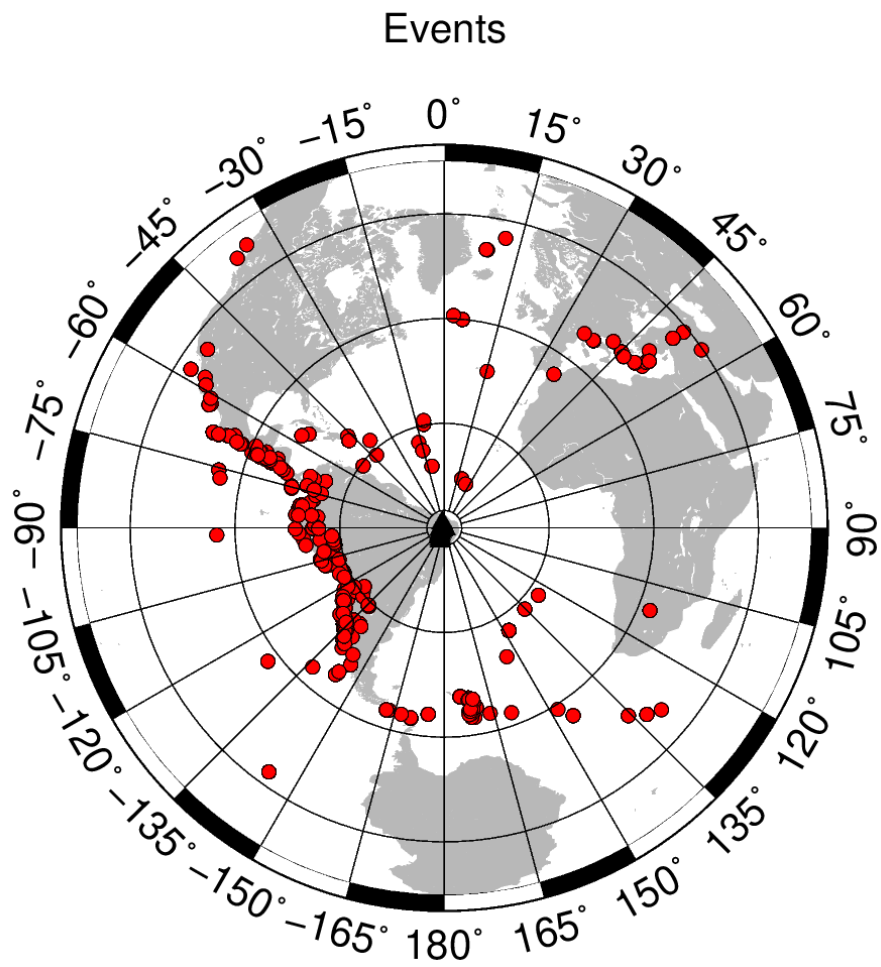


Figure 2: *Global map seismic sources selected for this study (red circles) and stations of the BODES Experiment (black triangles).*

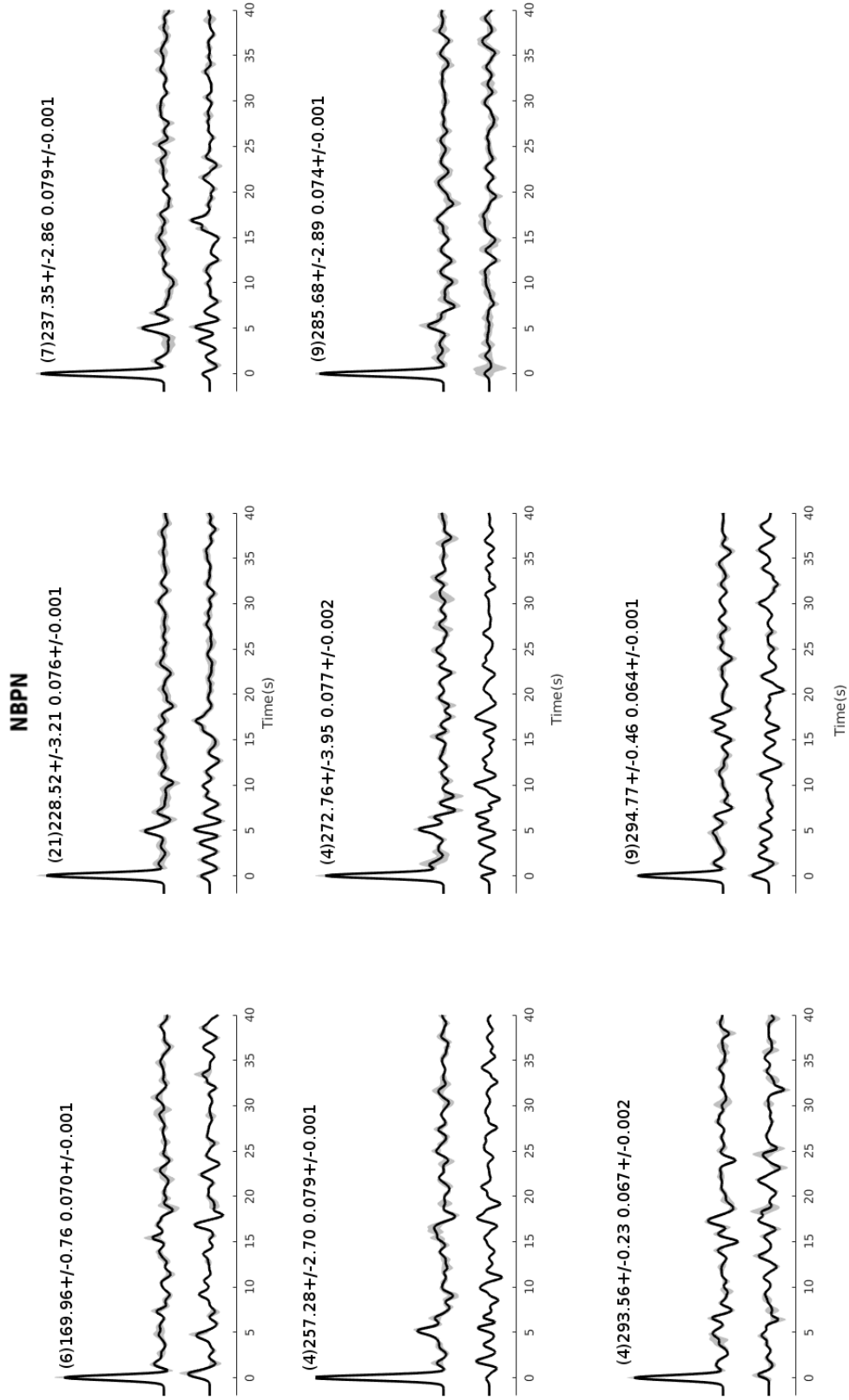


Figure 3: Sample receiver function stacks for station NBPB. Each panel displays two traces corresponding to the radial (top) and transverse (bottom) receiver functions, respectively. The stacks are presented as a black line within a grey shade representing 1σ -confidence bounds for the stacked amplitudes. The number of waveforms, average back-azimuth, and average ray parameter (in s/km) is indicated on top of each panel.

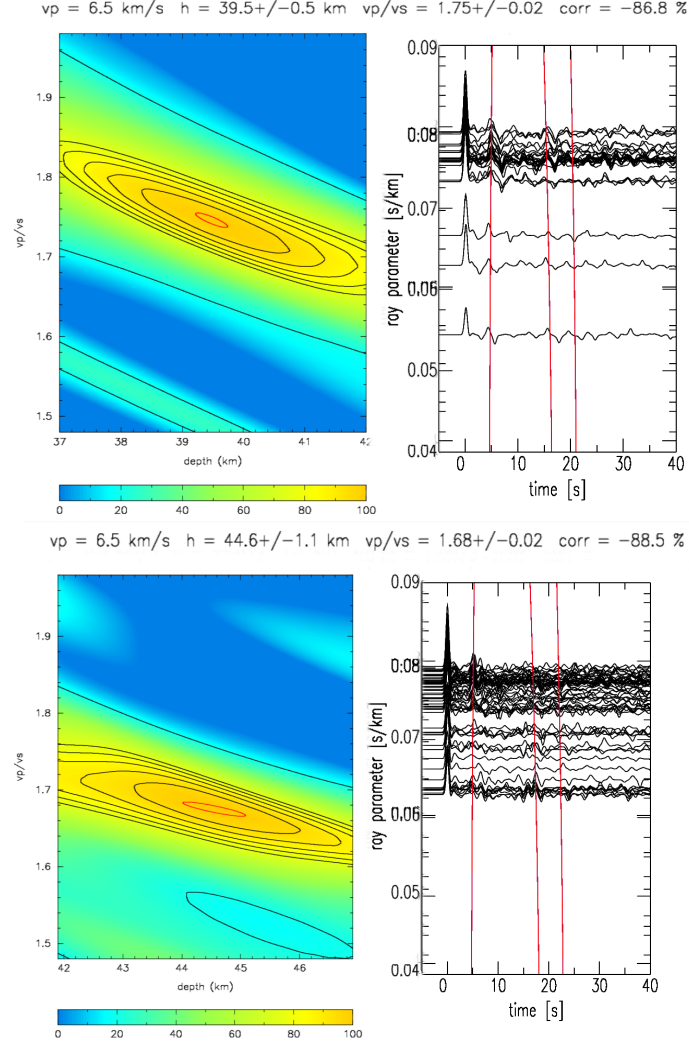


Figure 4: Sample H - κ stacking results for stations AR04 (top) and NBPB (bottom). The right panels show the corresponding radial receiver functions used during the stacking procedure, sorted by ray parameter, while the left panels show the H - κ stacking surfaces. The thin red lines in the right panels are the phase moveout curves for the Ps, PpPs and PpSs + PsPs phases associated with the maximum in the H - κ stacking surface. The thick red line in the left panels is the 1σ confidence ellipse from bootstrapping.

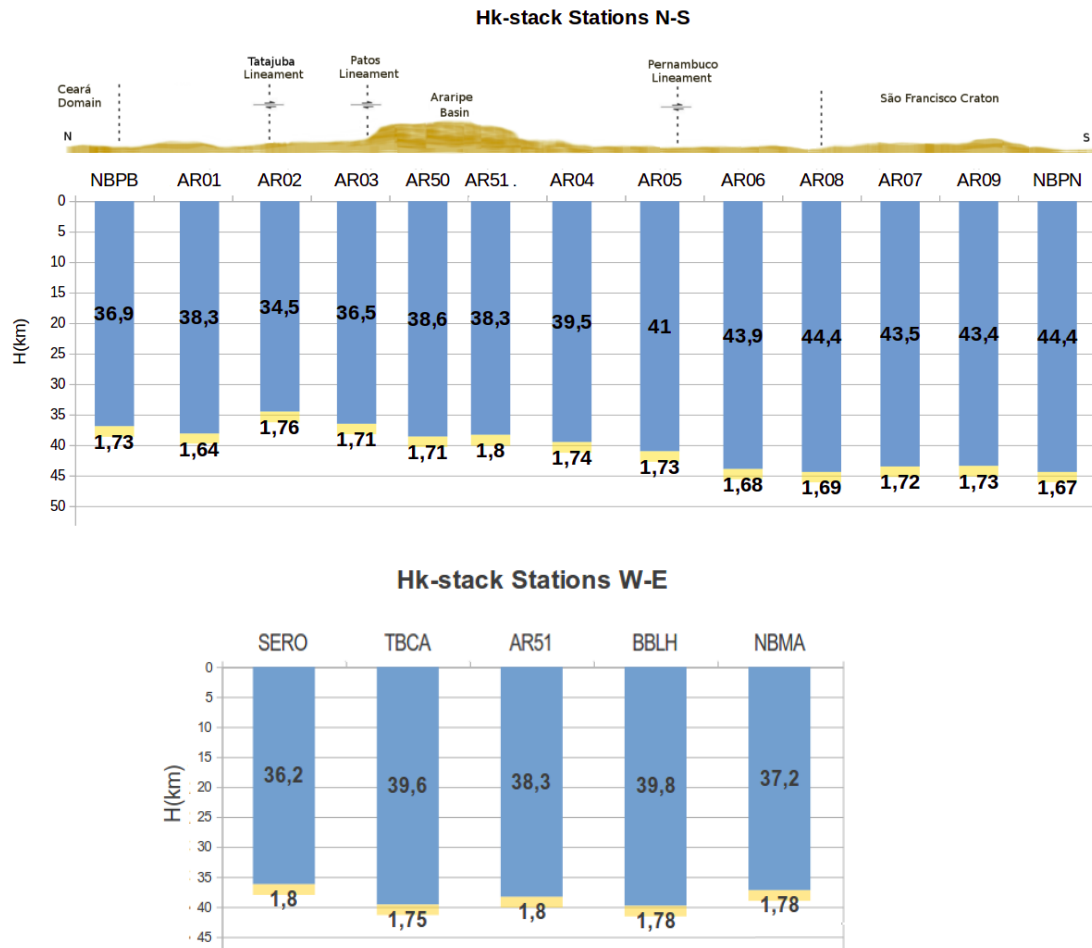


Figure 5: Summary of H - κ results for the BODES Experiment stations. Crustal thickness is indicated in the middle of the blue bands (in km) and V_P/V_S ratios are indicated at the bottom. Detailed H - κ stacking parameters can be found in Table 1.

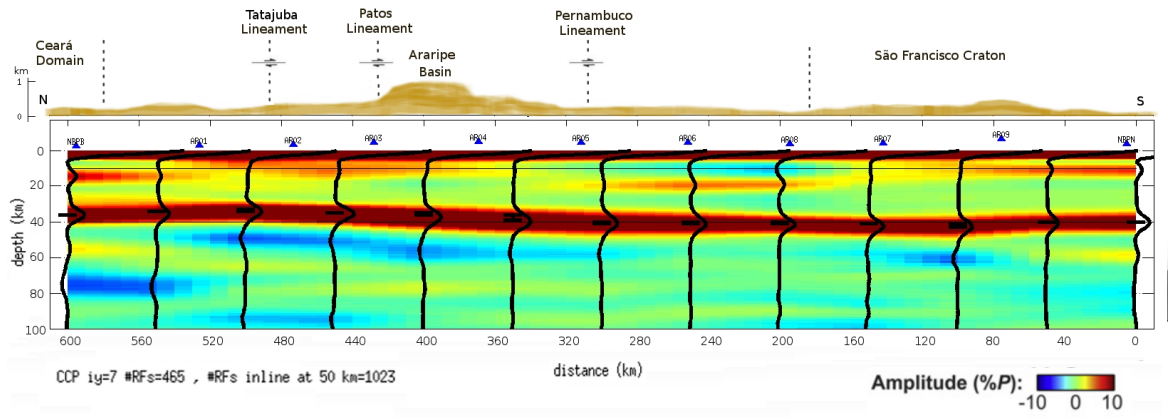


Figure 6: Receiver function CCP-stack along the BODES Experiment seismic line, showing clear *P* to *S* conversions at the crust-mantle boundary, as delineated by positive receiver function amplitudes.

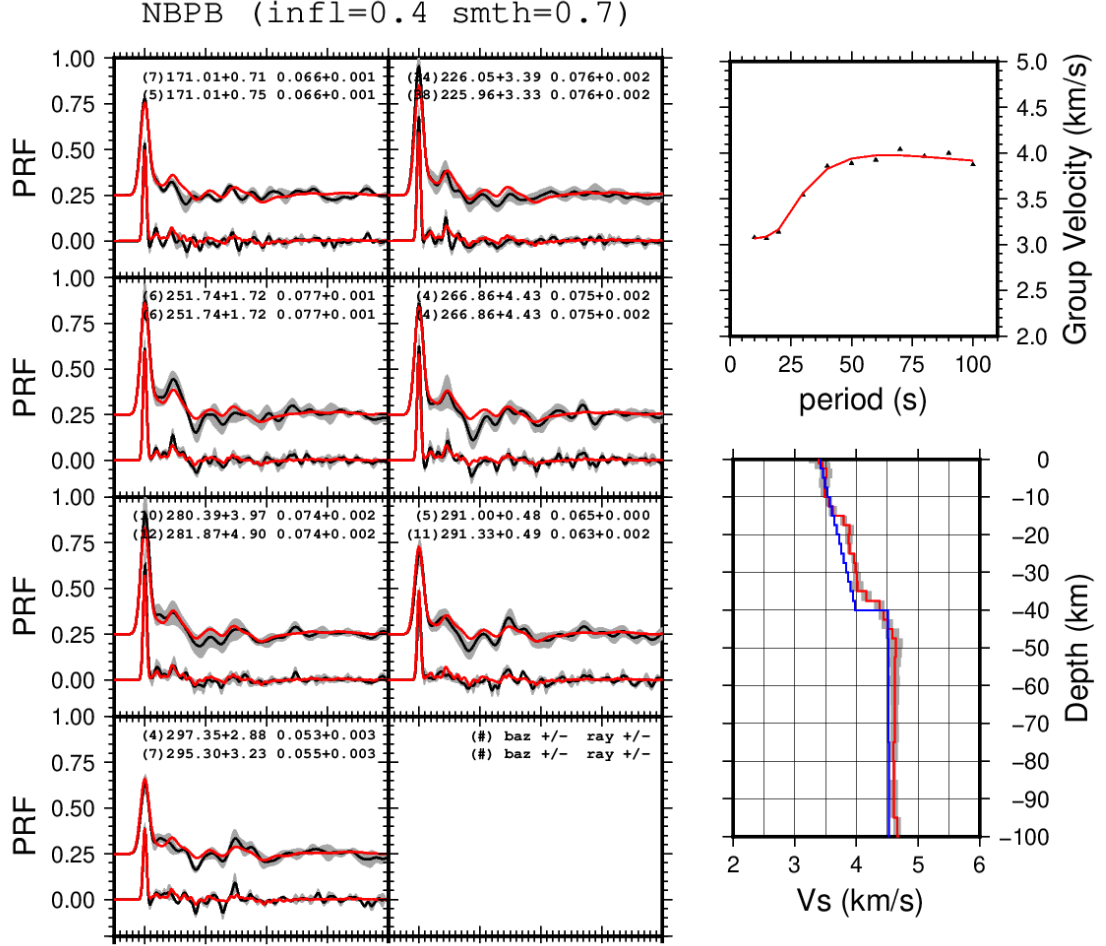


Figure 7: Detailed joint inversion results for station NBPB. Observed (black) and predicted (red) receiver functions at high- and low-frequency contents are displayed on the left-hand panels. Observed (red) and predicted (triangles) dispersion velocities are displayed in the upper-right panel, phase (green) and group (red) dispersions. The inverted (red) and starting (blue) velocity models are displayed in the lower-right corner, along with the 2σ confidence bounds (grey shade).

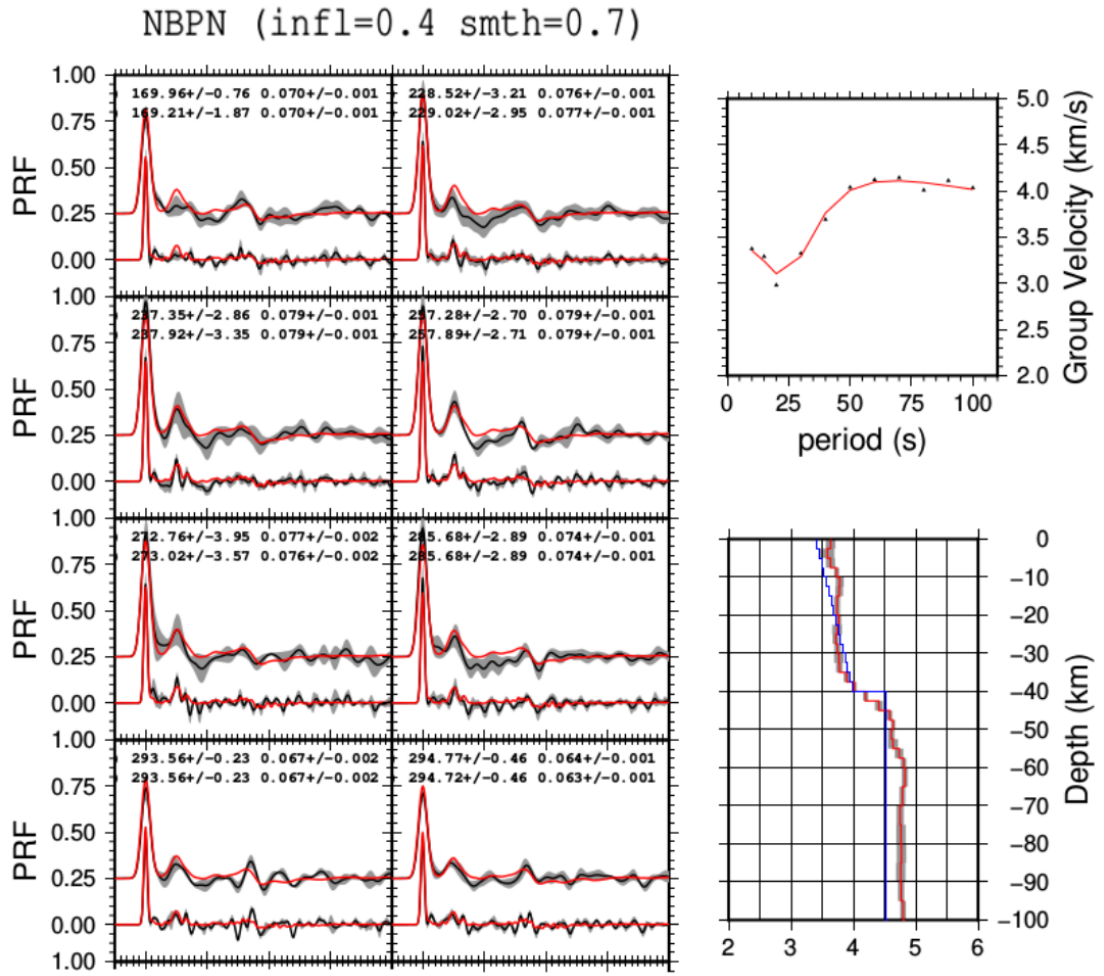


Figure 8: Same legend as in Figure 7, for station NBPN.

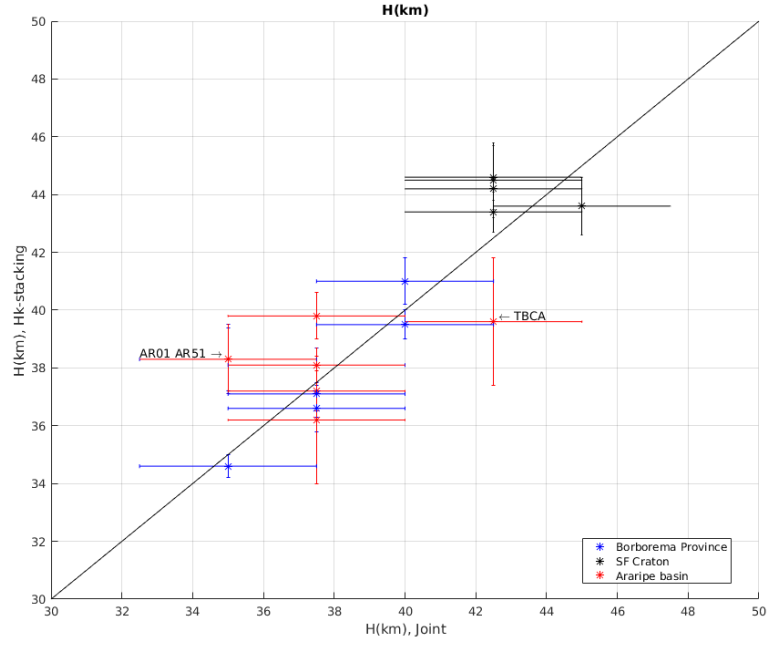


Figure 9: Comparison of crustal thicknesses from $H-\kappa$ stacking and joint inversion. Crustal thickness from the joint inversion models was obtained from the top of the first layer above a threshold S -velocity of 4.3 km/s; crustal thickness from the $H-\kappa$ stacking was taken from the Gaussian width that gave the closest estimate to that from the corresponding joint inversion model. Confidence bounds for the joint inversion estimates were set to ± 2.5 km (i.e. $\pm$ one layer thickness). Note the general consistency between both sets of estimates.

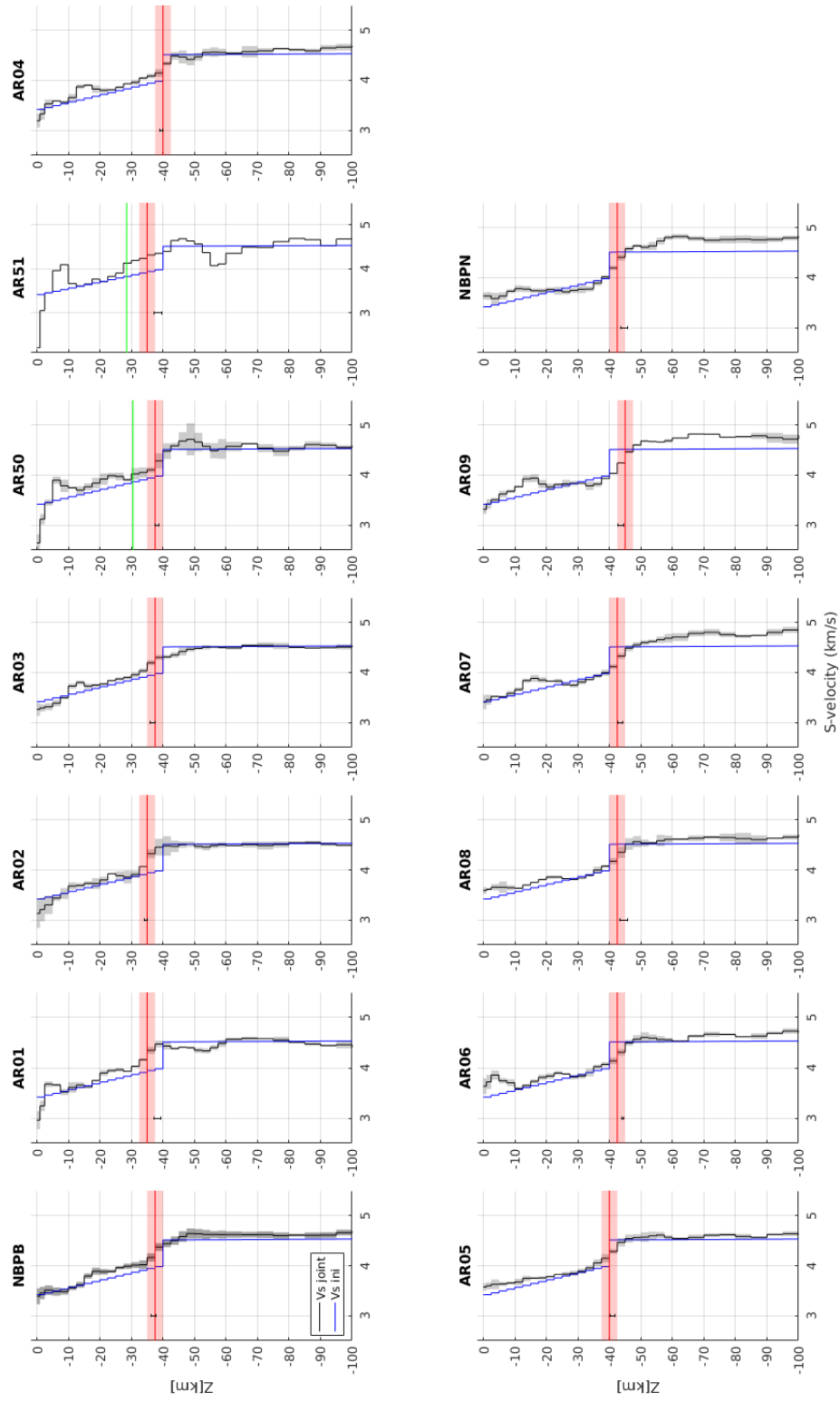


Figure 10: Velocity-depth profiles for each of the BODES Experiment NE stations. Each panel displays the inverted model (black) with the 2σ -confidence bounds (grey shade) and the starting PREM model (blue). Moho depth (red line) is noted with ± 2.5 km confidence bounds (red shade). H - κ stacking results for crustal thickness are superimposed as a black error bar. Note the presence of a thick layer of fast-velocity material at the top and bottom of the crust under station AR51. The relatively high-velocity at 10-20 km depth under station AR09 is likely to be an artifact.

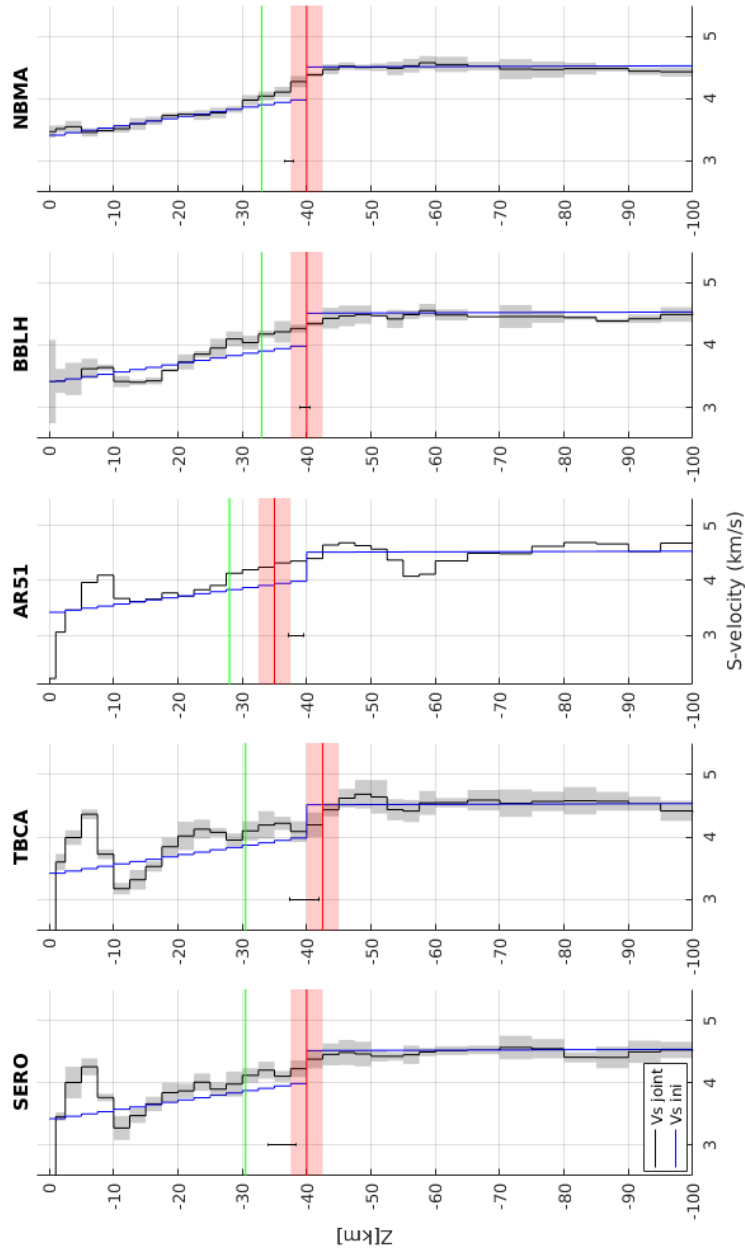


Figure 11: Velocity-depth profiles for each of the BODES Experiment WE stations. Same legend as in Figure 10. Note the presence of a thick layer of fast-velocity material at the bottom of the crust under all stations of Araripe basin, which is generally accompanied by a similar fast-velocity layer at 5-10 km depth

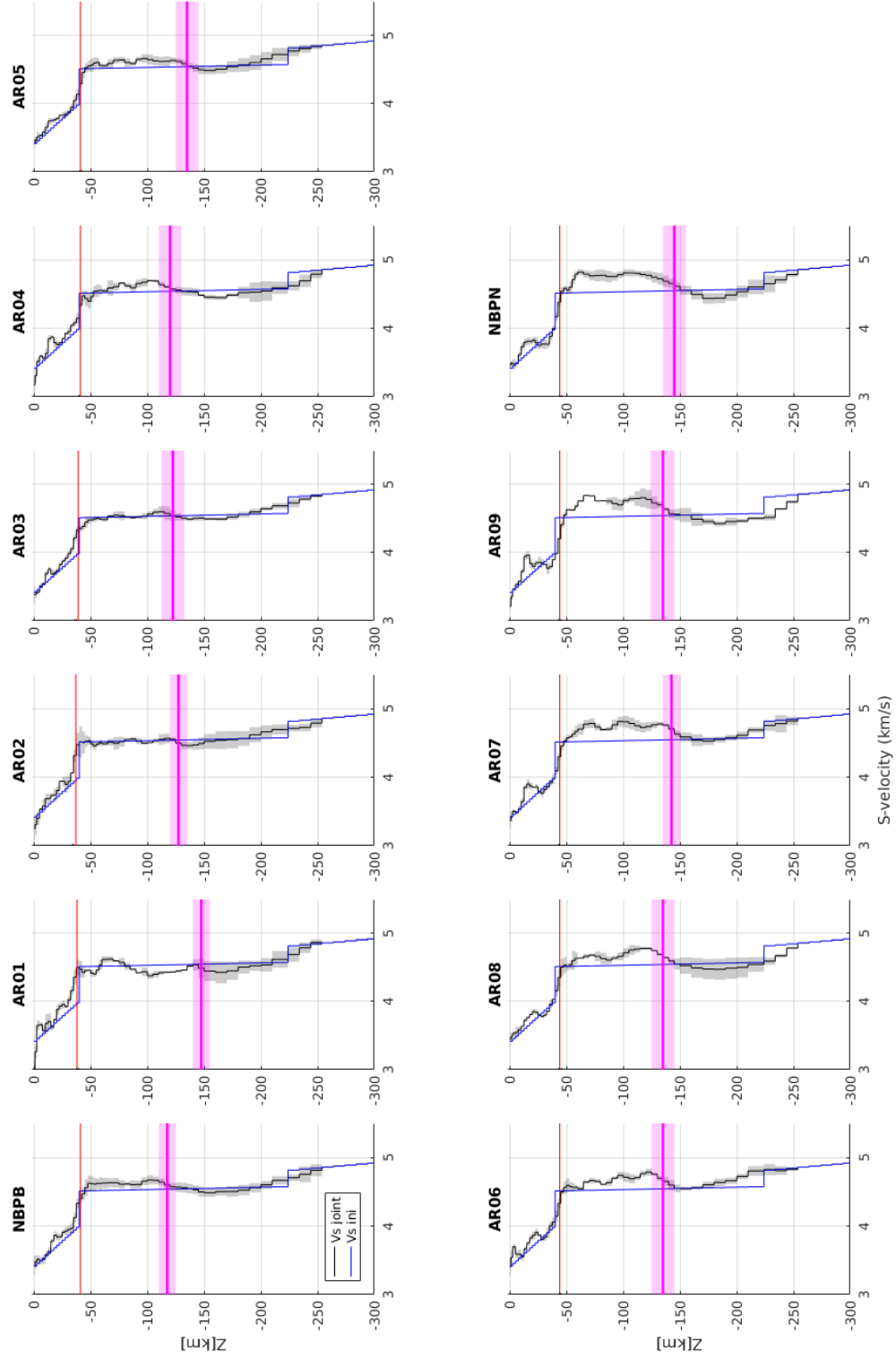


Figure 12: Velocity-depth profiles for each of the BODES Experiment stations down to sub-lithospheric depths. Each panel displays the inverted model (black) with the 2σ -confidence bounds (grey shade) and PREM model (blue). The crust-mantle boundary (red) and the LVZ depth (magenta) are indicated for each velocity-depth profile.

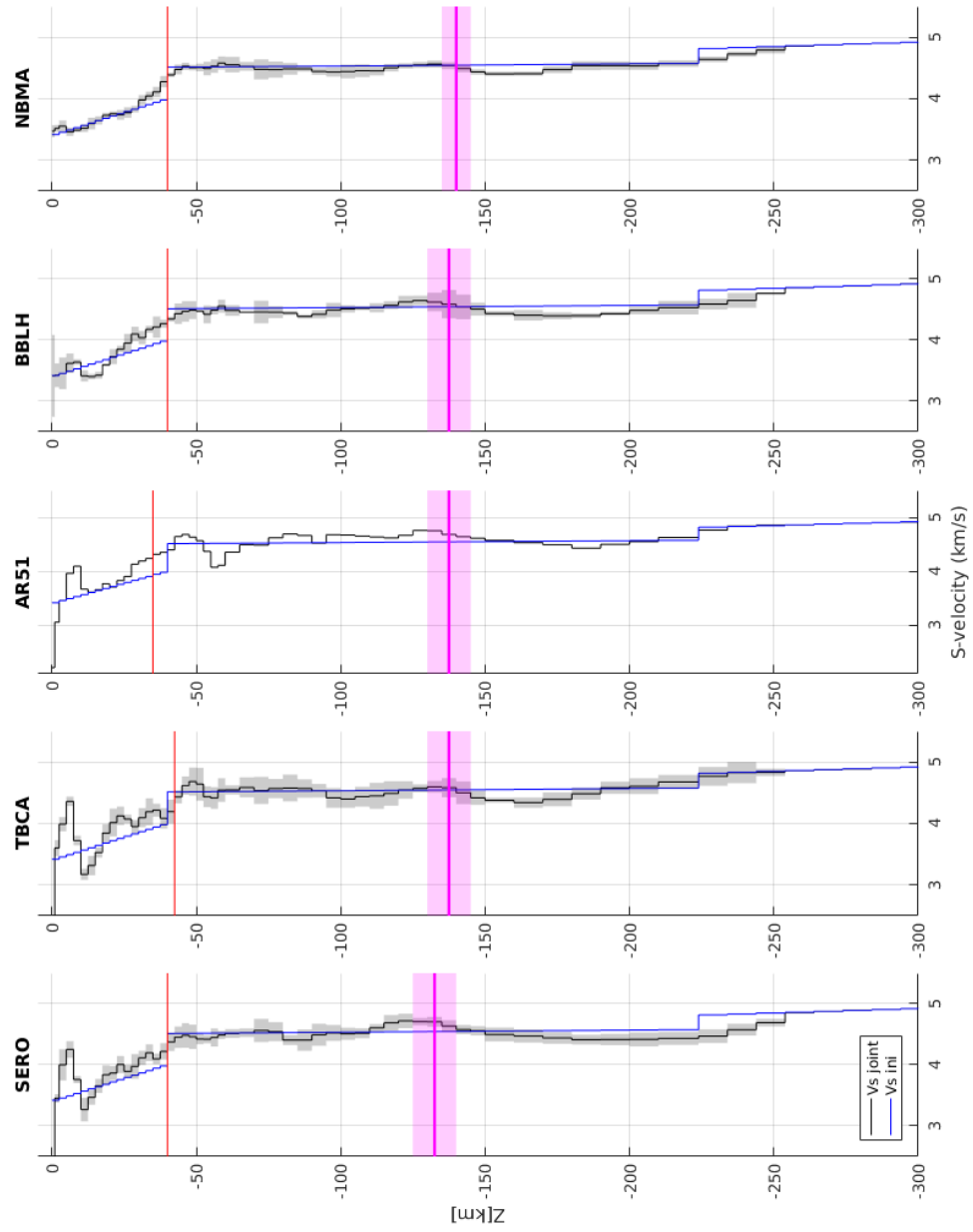


Figure 13: Same legend as in Figure 12.

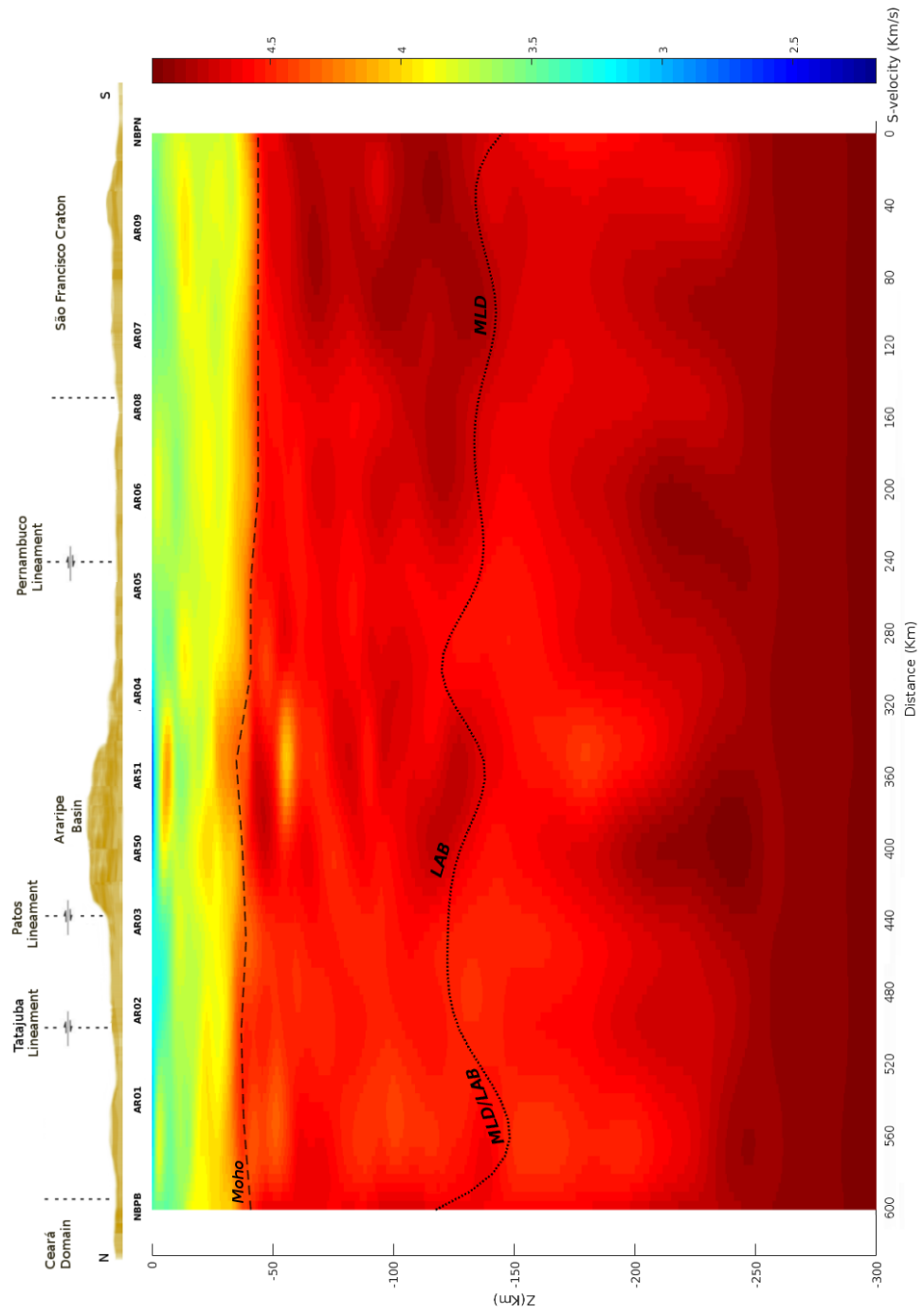


Figure 14: *Pseudo-2D cross-section constructed through cubic interpolation of all 1D BODES Experiment velocity-depth profiles, down to sub-lithospheric depths. Moho depths and interpreted MLD and/or LAB discontinuities are super-imposed.*

Chapter 5

Joint inversion of magnetotelluric and seismological observations:

Imaging the electric Moho in NE Brazil. (Paper II)

The following manuscript, entitled *Joint inversion of magnetotelluric and seismological observations: Imaging the electric Moho in NE Brazil*, presents results from joint inversion of magnetotelluric data, receiver functions and dispersion (phase velocity) curves. This manuscript was submitted for publication in Geophysical Journal International on 22nd June 2023.

1 Joint inversion of magnetotelluric and seismological
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12 **Abstract**

13 We assess the performance of a joint inversion of collocated magnetotel-
14 luric and seismological data acquired along a NS-trending profile centered
15 at the Araripe Basin of NE Brazil, and extending into the neighboring Bor-
16 borema Province and São Francisco Craton. Previously, separate seismic
17 (1D) and resistivity (3D) models had been developed from the same datasets
18 by jointly inverting teleseismic receiver functions and group-velocity disper-
19 sion curves, and magnetotelluric apparent resistivity curves and phases, re-
20 spectively. Our joint inversion utilizes the same site-specific receiver function
21 and apparent resistivity amplitudes and phases of the independent inversions,
22 a newly developed phase-velocity dispersion curve representing a network av-
23 erage, and assumes a common seismic and resistivity layered structure (1D)
24 under any given station. Moreover, the inversion is achieved through a multi-

25 objective genetic algorithm (GA) scheme that effectively explores the entire
26 parameter space. Our findings demonstrate that, in spite of significant geo-
27 logical complexity, it is possible to determine structurally consistent velocity
28 and resistivity layered structures for both the Borborema Province and São
29 Francisco craton. Although the accuracy of our simple (1D) inverted models
30 prevents us from developing empirical relationships between resistivity and
31 seismic S-velocity at crustal or mantle depths, they succeed in imaging an
32 electric Moho under most of the sites. Indeed, most models display a resis-
33 tivity increase at depths coincident with a seismic velocity increase identified
34 as the seismic Moho. Moreover, the resistivity increase occurs immediately
35 under a portion of conductive lower crust, roughly consistent with global
36 compilations of Precambrian resistivity structure. The electric Moho was
37 not observed in the 3D resistivity model, so we suspect that the removal of
38 smoothness constraints in our layered resistivity profiles, along with struc-
39 tural help from the collocated seismic data, guided the identification.

40

41 *Key words:* Joint inversion, Seismological Data, Magnetotellurics, Electric
42 Moho, NE Brazil

43 1. Introduction

44 Joint inversion (JI) of geophysical datasets is commonly regarded as an
45 effective tool for integrating constraints on subsurface Earth structure in an
46 optimal, self-consistent manner. When the datasets are sensitive to a com-
47 mon geophysical parameter, JI results benefit from complementary datasets,
48 as constraints from one dataset may bridge resolution gaps in the other and

49 vice-versa. This is the case, for instance, of the JI of receiver functions and
50 surface-wave dispersion (e.g. Julià et al., 2000). In that case, receiver func-
51 tions constrain relative S-P travel-times and velocity contrasts while disper-
52 sion velocities constrain average S-velocity over frequency-dependent depth-
53 ranges, resulting in velocity-depth profiles for which average S-velocities are
54 constrained by one dataset while small-wavelength details are superimposed
55 by the other. When, on the other hand, the datasets are sensitive to differ-
56 ent physical parameters, some sort of additional relationship that link those
57 physical parameters is required. Those relationships may range from purely
58 structural (e.g. physical parameters may change according to a commonly pa-
59 rameterized structure), to empirical (i.e. linking parameters through empir-
60 ically derived relationships), to petrological (i.e. linking parameters through
61 theoretical, microscopic-scale models). Among those, structural relationships
62 are the least restrictive and easiest to satisfy during an inversion process.

63 The JI of electromagnetic (EM) and seismic (passive- or active-source)
64 datasets has been given particular consideration in the literature, as both
65 data sets are commonly utilized for constraining the continental crust and
66 lithosphere. It is important to realize that while seismic data constrain seis-
67 mic velocity, which is mainly sensitive to temperature and composition, EM
68 data constrain electrical conductivity, which is more sensitive to fluid content
69 (i.e. water, partial melts). A variety of approaches are available for the joint
70 inversion of those datasets. Gallardo and Meju [2003], for instance, developed
71 a 2D joint inversion scheme for collocated dc resistivity and seismic refraction
72 experiments that links both sets of physical parameters (seismic velocity and
73 electrical conductivity) through their cross-gradients. The cross-gradient,

74 defined as the cross product of the gradients, enforces a structural control
 75 on velocity and conductivity variations without coercing the two models to
 76 converge when there is no data to justify it. Moorkamp et al. [2010], on the
 77 other hand, implemented a 1D joint inversion of receiver functions, disper-
 78 sion curves and MT observations that uses a genetic algorithm (GA) scheme
 79 to avoid a localized search characteristic of linearized approximations. A
 80 structural link is implicitly assumed through a common layered structure for
 81 both sets of physical parameters. In addition, their scheme incorporates a
 82 multi-objective approach, NSGA-II [Deb et al., 2002], which provides a set
 83 of best final models ranking according to principle of dominance. This al-
 84 lows to inspect the trade-off between data sets and their compatibility on the
 85 common layered structure.

86 Other approaches have utilized empirical relationships between seismic
 87 velocity and resistivity to constrain the JI. Gao et al. [2012], most notably,
 88 utilize Archie’s law (EM) and rock-physics fluid-substitution equations (seis-
 89 mic) to link electrical resistivity and seismic velocity to porosity and fluid
 90 saturation in reservoir rocks. Their work shows that, whereas seismic data
 91 are more sensitive to porosity, electromagnetic data are more sensitive to
 92 fluid-saturation. Lastly, Abubakar et al. [2012] compare the performances
 93 of the petrophysical and structural JI of controlled-source electromagnetic
 94 (CSEM) and full seismic waveform data in reservoirs. The joint petrophysi-
 95 cal inversion is similar to that of Gao et al. [2012], and is carried out by linking
 96 porosity and fluid saturation to seismic velocity and electrical conductivity.
 97 The best results, nonetheless, are obtained from the joint structural inver-
 98 sion, due to the higher degree of unknown parameters and the computation

99 complexity on the joint petrophysical inversion.

100 Here, we assess the performance of the JI algorithm of Moorkamp et al.
101 [2010] to invert MT, receiver function and Rayleigh wave dispersion data
102 acquired along collocated NS-trending profiles in NE Brazil. The approach
103 was successfully applied to data sets in the Slave Craton of North America
104 and the Kaapvaal Craton of Southern Africa, for which the assumption of
105 an underlying 1D medium could be safely assumed [Moorkamp et al., 2007].
106 Indeed, receiver functions are generally 1D, as they are obtained from near-
107 vertically propagating seismic waves from distant sources (e.g. Ammon et
108 al., 1990), with off-azimuth energy becoming important in the presence of
109 strongly dipping structures and/or anisotropy right under the receiver (Sav-
110 age, 1998; Schulte-Pelkum and Mahan, 2010). MT data, on the other hand,
111 can seldom be regarded as 1D, as the presence of electrical charges in local,
112 near-surface heterogeneities close to electrical conductivity transitions can
113 cause sizeable galvanic distortion of magnetotelluric data [Garcia and Jones,
114 2002]. Thus, further testing of this approach in more complex tectonic set-
115 tings is required in order to assess its general applicability.

116 Our datasets consist of teleseismic and MT data acquired at 11 collocated
117 broadband seismic and long-period MT stations. The stations were deployed
118 along a ~ 600 km long line crossing the Proterozoic Borborema Province
119 of NE Brazil and the Archean São Francisco craton, with an inter-station
120 spacing of ~ 70 km. Separate inversions of the seismic and magnetotelluric
121 datasets already exist, providing reliable models against which our JI results
122 can be benchmarked. The 3D MT model was developed by Garcia et al.
123 [2019] employing ModEM [Kelbert et al., 2014], and found a marked resis-

124 tivity anomaly of 25 Ωm under the Araripe Basin that was interpreted as ev-
125 idence for a thin (120 km) lithosphere. Moreover, northward and southward
126 of the conductive anomaly, two thick resistive structures reaching more than
127 200 km depth were found, the southern one coinciding with the São Francisco
128 craton. The S-velocity model was obtained by Nemocón et al. [2021] from the
129 joint inversion of receiver functions and dispersion curves, and demonstrated
130 the presence of mafic material at the bottom of the crust under the Araripe
131 basin. The seismic model showed a crustal thickness between 34 and 44 km
132 along the profile and variations of lithosphere thickness ranging from ~ 200
133 along the craton to 120 km under the basin. The results from separate in-
134 versions thus suggest some degree of correlation at lithospheric mantle levels
135 between seismic and resistivity structures, which justifies our attempt at a
136 joint inversion of both data sets.

137 As mentioned above, however, seismological methods are typically uti-
138 lized to obtain subsurface seismic velocity, while MT techniques are a pow-
139 erful tool to recover electrical resistivity. One important challenge that we
140 faced was to sensibly model the MT dataset - which is markedly 3D [Garcia
141 et al., 2019] - through a stack of resistivity layers structurally consistent with
142 those in the seismic velocity model. In spite of those difficulties, our results
143 show great structural compatibility between electrical and seismic parame-
144 ters, as evidenced through trade-off analysis of ~ 350 optimal models from
145 the final Pareto front for each station. Moreover, this compatibility enabled
146 us to assess the viability of empirical relationships between conductivity and
147 resistivity in our study area, and to map the elusive electric Moho under
148 most of the study region.

149 2. Data and processing

150 2.1. The BODES experiment

151 The data sets utilized in this study were acquired by the Universidade
152 Federal do Rio Grande do Norte as part of the BORborema Deep Electro-
153 magnetic and Seismic (BODES) experiment, a temporary array of collo-
154 cated seismic and magnetotelluric stations crossing the western Borborema
155 Province and northern São Francisco craton. The seismic network consisted
156 of 9 temporary broadband stations that were deployed in 2015, and two per-
157 manent broadband stations belonging to the *Rede Sismográfica do Nordeste*
158 (RSISNE). The magnetotelluric (MT) component of the experiment con-
159 sisted in 11 long-period stations that were installed in May and October of
160 2015 at the same locations as the seismic instruments. Instrument location
161 coordinates are listed in Table 3.

162 The MT stations and the temporary broadband seismic stations were
163 borrowed from the *Pool de Equipamentos Geofísicos do Brasil* (PEGBR).
164 Each MT station included a three-axis magnetometer (FluxGate) and five
165 non-polarizable Cu-CuSO₄ electrodes. Those receivers were connected to a
166 LEMI-417M digitizer sampling at 1 Hz, which stores the time series for up
167 to seven channels: three magnetic and two or four electric fields. The LEMI
168 instruments included a GPS receiver that provided satellite synchronization
169 to the internal clock and records the coordinates. The temporary seismic
170 stations consisted of RefTek 120A sensors, with flat velocity response in the
171 0.01 - 50 Hz frequency range, attached to high-gain RT-130 digitizers and a
172 GPS antenna for timekeeping. This setup matched that of the two permanent
173 seismic stations. Seismic stations recorded continuously for about 2 years,

174 except for stations AR01 and AR09 which recorded for about just 1 year. MT
175 instruments simultaneously recorded between 21 and 34 days, which allowed
176 us to obtain remote reference robust estimates of the resistivity distribution
177 down to LAB depths.

178 *2.2. Receiver functions*

179 We considered the same receiver function dataset as in Nemocón et al.
180 [2021], so we will only give a brief description of the major processing steps.
181 Seismic events were selected with magnitude greater than 5.5 mb and epi-
182 central distances between 30° and 90° , with no depth restrictions. The loca-
183 tion of the seismic sources were mainly along the subduction zones of North
184 and Central America and along the Mid-Atlantic Ridge [Nemocón et al.,
185 2021]. Before the deconvolution process was carried out some additional pre-
186 processing steps were taken: a) Teleseismic waveforms were windowed 10 s
187 before and 110 s after the P-wave, demeaned, detrended and tapered; b) a
188 high-pass filter above 0.025 Hz was applied to remove long-period noise; c)
189 a low-pass filter below 8 Hz was applied to avoid aliasing; and d) record-
190 ings were decimated to 10 s.p.s. Receiver functions were estimated from the
191 selected waveforms applying the time-domain deconvolution approach of Lig-
192 orría and Ammon [1999], using 500 iterations. The resulting time series were
193 low-pass filtered with Gaussian filters of width $a=1.0$ and 2.5 , producing re-
194 ceiver function estimates at overlapping frequency bands of $f < 0.5$ and $f <$
195 1.2 Hz. Additionally, a quality control was applied, which included: a) rejec-
196 tion of receiver functions that do not reproduce at least 85% of the original
197 radial waveform when convolved back with the vertical component; b) re-
198 jection of receiver functions with a large transverse component or oscillatory

behaviour; c) identification and rejection of outliers after superimposition of radial receiver functions. In total, 2319 waveforms were examined for the 11 seismic stations, resulting in 477 receiver functions at high-frequency contents ($a=2.5$) and 546 receiver functions at low-frequency contents ($a=1.0$).

In order to improve the signal-to-noise ratio, we performed a point-to-point stack of the selected receiver functions within bins defined by ray parameter, with variations of less than 0.01 s/km, and back-azimuth, with variations under 8° . These relatively narrow bins were chosen to account for the phase moveout of multiple reverberations. An additional inspection of the receiver function averages was necessary before inversion, which consisted of a visual inspection and a check on P-to-S conversion arrival times. Receiver function averages were required to: (i) display a direct P-wave with zero lag time and a large amplitude; (ii) display P-to-S conversions on lag times about 5 and 14 s, corresponding to the Ps and PsPs conversions at the crust-mantle boundary; and (iii) display transverse receiver functions with small amplitudes relative to the corresponding radial amplitudes, which indicates an isotropic and laterally homogeneous medium under the recording station. For both components, the receiver functions that did not meet these criteria were excluded. In total, 46 receiver function averages were selected, each with an average of ~ 18 receiver functions.

Finally, a new visual inspection was carried out within each bin to select individual RFs for inversion. This additional culling of our dataset resulted in a final selection of 202 receiver functions.

2.3. Rayleigh wave dispersion data

Phase velocities were obtained through the array-processing approach of McMechan and Yedlin [1981], as implemented in the Computer Programs in Seismology package of Herrmann and Ammon [2002]. This approach is a technique to obtain phase velocity dispersion from an array of seismic traces. It is based on an initial transformation from the distance-time (x, t) domain into the ray parameter-time intercept (p, τ) domain through slant-stack integration. A second transformation into the (p, ω) domain then follows, through 1D Fourier integration over the τ variable, which effectively yields the phase velocity dispersion relation through its equivalence in the (c, T) domain (recall that $p=1/c$, $\omega = 2\pi/T$). Rayleigh-wave phase velocities thus obtained represent an average dispersion curve for the medium beneath the recording network.

A total of 18 events located along the South American subduction zone and Mid-Atlantic ridge, with magnitudes between 5.0 and 5.9 Mw and epicentral distances ranging from 1050 to 3412 km, were selected. Data preparation consisted of the following steps: 1) demean, detrend and taper the waveforms with a 2% cosine-window; 2) apply a low-pass filter at 1.0 Hz; 3) cut the waveforms between 400-1600 s after the origin time to isolate the surface-wave signal; and 4) resample to 0.5 seconds. Rayleigh-wave phase velocities were thus obtained for periods in the 12 - 80 s period range, with velocities ranging between 3.39 and 3.93 km/s (see Figure 2).

2.4. Apparent resistivity and phase

As mentioned above, the magnetotelluric survey consisted of 11 MT stations that were installed between May and October of 2015 and collocated

with the seismic sensors. Details on the recording times, instrument specifications, and processing can be found in Garcia et al. [2019]. Here, we used the same impedances as in Garcia et al. [2019], after eliminating the highest frequencies (<10 s) and other spurious or noisy points or sections.

3. JI of MT and seismological data

3.1. Joint Inversion Details

We parameterize the medium under a given station as a stack of layers, each defined through layer thicknesses, isotropic shear-wave velocity (V_s) and isotropic logarithmic resistivity ($\log(\rho)$). As a GA approach (NSGA-II) is considered, a vector of model parameters ($\mathbf{m}$) is constructed so that each individual component m_i is concatenated into a binary string. A random population of fixed size P , where each member represents one possible model, is then created. The encoded binary strings is transformed back into vectors of physical parameters and the corresponding objective function values are computed for each of them.

An advantage of the NSGA-II algorithm is that it stores the objective function values in a vector $\mathbf{O}(\mathbf{m})$ for each population member, so that the algorithm can select models for the next generation while assigning a rank to each population member based on the concept of Pareto optimality. Pareto optimality in this context means that a model is only considered superior to another model if all objective function values are equal or smaller and at least one objective function value is smaller. For example, let's consider a model for which the misfit to MT data is smaller, but the receiver function misfit is larger; and a different model for which the misfit to MT data is larger, while

the corresponding to the receiver function is smaller. The two models would be considered equally valid and part of the Pareto front. The algorithm actively tries to find as many of these equally valid models as possible, while preserving the best fitting models. A full mathematical description of the algorithm can be found in Deb et al. [2002] and Moorkamp et al. [2010].

Once the best Pareto-optimal models are preserved and selected within a given generation, a new iteration is performed through repetition of the previous steps. Thus, after a fixed number of iterations, we have obtained the final Pareto front and we can use a variety of criteria, such as maximum curvature [Hansen, 1992], to identify the best model within the front. As we invert three datasets (objective functions), the Pareto front corresponds to a multidimensional trade-off surface that can be used to analyze the structural compatibility between the datasets and their noise levels. All objective functions - MT misfit, receiver function misfit and surface wave misfit - are given identical weight to avoid penalizing any of them.

During inversion the RMS misfit Φ used for each dataset is defined as

$$\Phi = \sqrt{\frac{1}{N} \sum_{j=1}^N \left(\frac{d_j^{obs} - d_j^{synth}}{\Delta d_j^{obs}} \right)^2}, \quad (1)$$

where N is the number of measurements, d_j^{obs} the individual observations, d_j^{synth} the corresponding model predictions and Δd_j^{obs} the estimated data uncertainties. When inverting individual datasets we also use a regularization functional θ as an additional objective function, which is defined as

$$\theta = \sum_{i=0}^{N-1} \left(\frac{m_{i+1} - m_i}{\Delta z_i} \right)^2. \quad (2)$$

291 Finally, as the coupling between seismic and electrical models is based
 292 on coincident layer interfaces and the number of layers in the inversion is a
 293 fixed parameter (which determines the degrees of freedom), it is important to
 294 keep the number of inversion layers sufficiently small to ensure good coupling
 295 among the data sets. To obtain the appropriate number of layers, we perform
 296 a number of inversions by varying the number of layers and investigate their
 297 effects in the inversion of MT data alone. Although an appropriate number
 298 of layers could be guessed from the velocity models developed in Nemocón
 299 et al. [2021] for the seismic dataset, we still prefer to test for the optimum
 300 number of layers required by MT data. These experiments are described in
 301 more detail below.

302 *3.2. Static shift corrections*

303 Galvanic distortion analysis by [Garcia et al., 2019] of the MT dataset
 304 utilized in our study revealed large galvanic distortion values indicative of a
 305 complex subsurface geo-electrical structure. Thus, besides using a galvanic
 306 distortion removal algorithm, static shift corrections must be applied prior
 307 to 1D modeling of our MT dataset. Static shift corrections were determined
 308 empirically through comparison of the inverted 1D resistivity profiles with
 309 analogous resistivity-depth profiles extracted from the 3D inversion results
 310 of [Garcia et al., 2019] (Figure 3) and using MTPy [Krieger and Peacock,
 311 2014; Kirkby et al., 2019] to perform the shift for each site.

312 *3.3. A detailed example*

313 We illustrate the JI procedure through a detailed inversion of MT and
 314 seismic data acquired at station AR01. The first step is the selection of

315 the number of inversion layers using MT data only. Before that, however, we
 316 inspect the MT data, reject any spurious noisy values and select a component
 317 of the impedance tensor to fit (either xy or yx). The selection procedure
 318 is shown in Figure 4 for station AR01, where the trade-off curves between
 319 model roughness and RMS misfit and the inverted resistivity models obtained
 320 through models with 7, 8, 9, 10, 11, 12 and 14 layers are displayed along with
 321 the change in RMS misfit with the number of layers. The RMS misfit initially
 322 decreases as the the number of layers increases, to then slightly increase with
 323 the number of layers. Overall, however, the variation in RMS misfit is small
 324 and, in all cases, the inverted models exhibit the same features at crustal
 325 and mantle levels: a crustal resistor at the bottom of the crust and a second
 326 resistor between 100 and 140 km depth, approximately. Based on these plots
 327 and the variation in misfit, we consider 10 layers as the minimum and most
 328 suitable number of layers to describe the main features in the MT data (Table
 329 1).

330 Once the number of inversion layers is selected, other parameters need
 331 to be set up. Recall that we invert for layer thickness (t), logarithmic layer
 332 resistivity ($\log \rho$) and S-wave velocity (V_s). Thus, in order to perform the
 333 joint inversion, we must first specify three values for each model parameter
 334 (m_i): The minimum possible value (m_i^{min}), the discretization step (δm_i),
 335 and the number of bits used for encoding (n_i). Setup values for station
 336 AR01 are given in Table 2. We then run the genetic algorithm inversion
 337 with $N_p = 1000$ and $N_g = 100$. We search for an S-velocity range between
 338 2.5 km/s and 5.0 km/s, which covers realistic values in the crust and upper
 339 mantle [Dziewonski and Anderson, 1981], and a resistivity range between 1

340 Ωm and $10^4 \Omega m$. Even more important is the discretization step, δm_i , as
 341 we have to make sure that the step size is small enough to avoid aliasing
 342 effects. In the inversions, we use a step size of 0.05 km/s for seismic velocity
 343 and 0.1 for logarithmic resistivity. Also, remember that we keep the number
 344 of inversion layers sufficiently small (10 layers) to achieve a good coupling
 345 among all data sets (Table 1).

346 The next step is to proceed with the JI of all three datasets. The inversion
 347 results are summarized through the trade-off curves shown in (Figure 5),
 348 where each panel displays two objective functions while color-coding the third
 349 one. The best model for site AR01 (red star) was selected in two steps:
 350 First, the trade-off curve between dispersion velocity and receiver function
 351 was visually inspected, and the region of highest curvature identified; then,
 352 the solution with the smallest MT misfit within that high-curvature region
 353 was selected. Considering the trade-off curves and a visual quality assessment
 354 of the features from seismic and electric models (black lines in Figure 6), we
 355 validate this model as our preferred solution.

356 Figure 6 reveals that the Moho is found at ~ 35.4 km depth, when defined
 357 as the top of the first layer above a threshold S-velocity of 4.3 km/s. Note
 358 that this depth corresponds to both an increase of S-velocity and to the base
 359 of a thin crustal conductor. Previously, Nemocón et al. [2021] had reported a
 360 crustal thicknesses of 38.3 ± 1.1 km from H- κ stacking and of 35.0 ± 2.5 km
 361 from JI of receiver functions and dispersion velocities at this site, very close
 362 to our JI result. At upper mantle levels we identify overall high-resistivities
 363 and high-velocities throughout. A comparison to our JI model (Figure 6)
 364 with a the model presented in Nemocón et al. [2021] using seismic data only

for the same station, however, displays noticeable changes, likely due to the use of an average phase-velocity curves in our JI approach when compared to station-specific group-velocity curves in Nemocón et al. [2021]. First-order features, such as depth to the Moho, are nonetheless preserved, and the match between observations and predictions is good for surface-wave dispersion, receiver functions amplitudes, and electric resistivity and phase (Figure 8).

3.4. Inversion results

The JI models for all the BODES experiment stations are displayed in Figure 9. Our velocity-depth profiles show that the Moho coincides with a clear increase in velocity from ~ 4.3 km/s to ~ 4.5 km/s at depths between ~ 34 km and ~ 44 km, in good agreement with purely seismological results reported in Nemocón et al. [2021]. The resistivity models show a sharp increase in resistivity for stations where a conductor exists in the middle-lower crust, whereas at the other stations shows a sharp decrease. High conductivity in the middle-lower crust is found under station AR01 ($\sim 12 \Omega m$ at ~ 19 km and ~ 36 km depth) and station AR08 ($\sim 10 \Omega m$ at ~ 20 km and ~ 45 km depth), which coincides with the C3 and C2 conductivity anomalies reported by Garcia et al. [2019]. A resistive crust is found for stations NBPB, AR04 and AR09.

At upper mantle levels, our velocity-depth profiles also agree reasonably well with results reported in Nemocón et al. [2021]; however, a low velocity zone starting at 120 - 150 km depth - as reported in that study - is not observed in our JI models. Recall that we utilized phase velocity curves with a longest period of 45 s, while Nemocón et al. [2021] considered group-

390 velocity curves with periods up to 100 s. The longer periods are critical to
391 constrain deep lithospheric structure and, as those periods are missing in our
392 phase-velocity curves, the resolution at lithospheric mantle depths becomes
393 poor. Our 1D resistivity model under station AR03, flanking the Araripe
394 Basin, does show a clear decrease in resistivity of 10-19 Ωm for depths larger
395 than > 96 km, however, which coincides with the C1 conductivity reported
396 by Garcia et al. [2019].

397 4. Discussion

398 The main result of our JI is that it is possible to determine structurally
399 consistent velocity and resistivity models under collocated seismic and mag-
400 netotelluric stations in the Western Borborema Province and northern São
401 Francisco craton. Moreover, the comparison in Figure 9 demonstrates that
402 the coarse, 1D layered structures assumed in our JI successfully account for
403 long-wavelength variations observed in the more detailed separate inversions
404 of the same seismological and magnetotelluric datasets. Recall, however, that
405 the joint inversion of receiver functions and group-velocity dispersion curves
406 of Nemocón et al. [2021] utilized station-specific dispersion curves for each
407 station, while here we just assumed an average phase-velocity dispersion
408 curve for the entire study area. Also recall that the MT inversion of [Garcia
409 et al., 2019] considered a full 3D resistivity model to account for lateral vari-
410 ations in the impedance tensor, and that we utilized those 3D results to pro-
411 duce static shift corrections during our 1D modeling. The long-wavenumber
412 variations in our resistivity models, nonetheless, are fully constrained by the
413 dataset during the (1D) JI procedure.

4.1. Resistivity-Velocity relationships

The broad consistency of the seismological and resistivity models obtained in our JI encourages us to investigate average resistivity-velocity relationships in the study region. The S-velocity models of Nemocón et al. [2021] reported a crust ~ 38 km thick under the Araripe Basin, as well as the presence of mafic underplate (10 – 14 km thick) in its lowermost crust resulting from melts formed in the underlying lithospheric mantle. They also reported crustal variations between 34 – 44 km depth along the entire profile and accompanying variations in lithospheric thickness. The magnetotelluric inversion of Garcia et al. [2019], on the other hand, found a significant thinning of the lithosphere (120 km) beneath the Araripe basin inferred from the presence of a marked conductive anomaly of 25 Ωm . This feature was interpreted as enriched asthenospheric material that, through a roll-down flow process, erodes material from under the basin.

Correlation plots between seismic S-velocity and $1/\log(\text{resistivity})$ from our (1D) JI results at select crustal and upper mantle levels are given in Figure 10. The plots are color-coded by terrain age, in order to assess potential secular variations of the inferred parameters. At crustal levels (20-40 km) there seems to be an overall decrease of S-velocity with age of the terrain, as well as a slight general increase in resistivity. However, the large scatter in the inferred velocity and resistivity values prevents from working out a clear resistivity-velocity relationship for the crust. At upper mantle levels (50-100 km) variations in S-velocity seem to be losing the age dependency observed at crustal levels (except, perhaps, at 100 km depth) and display a wide range of velocity variation (4.4 - 4.9 km/s) throughout. In contrast,

439 variations in conductivity seem to be restricted to a narrower range (0.3-0.7)
 440 when compared to the wider range (0.1-0.9) range observed at crustal levels,
 441 again making it difficult to infer a clear relationship between conductivity
 442 and seismic S-velocity. Similar correlation plots are provided in Figure 11
 443 using S-velocity and $1/\log(\text{resistivity})$ values from Nemocón et al. [2021] and
 444 Garcia et al. [2019], respectively. As before, there seems to be an overall
 445 decrease of S-velocity with terrain age, and perhaps a reversed trend of in-
 446 creased conductivity with age. Moreover, the range of $1/\log(\text{resistivity})$ is
 447 much narrower than in our previous plots, suggesting a linear relationship be-
 448 tween S-velocity and $1/\log(\text{resistivity})$, especially at larger depths. Note that
 449 the resistivity-depth profiles from Garcia et al. [2019] - and, to some extent,
 450 the velocity-depth profiles of Nemocón et al. [2021] - are smoother than those
 451 obtained from the JI (recall Figure 9) considered in our study, likely due to a
 452 less restrictive regularization required by our 10-layer parameterization dur-
 453 ing inversion. At upper mantle levels, a clear-cut increase of S-velocity and a
 454 subtle increase of resistivity with terrain age can be observed, especially at 75
 455 and 100 km depth, again suggesting a linear relationship between S-velocity
 456 and $1/\log(\text{resistivity})$ in the lithospheric mantle.

457 Our results can be compared to similar correlation plots developed by
 458 Jones et al. [2013] for the South African cratonic lithosphere. Their correla-
 459 tion plots were built from S-wave velocity and resistivity measurements from
 460 high-resolution tomographic (surface-wave) and magnetotelluric experiments
 461 in the region, and focused at depths of 100 km. Their values ranged between
 462 4.4 and 4.8 km/s for S-velocity and between 0.1 and 0.6 for $1/\log(\text{resistivity})$,
 463 similar to our inferred values at similar depths (recall Figure 12) and, perhaps

most interestingly, broadly fell within theoretical bounds for extreme water content conditions. The theoretical bounds were developed assuming linear dependence of elastic moduli and density with pressure and temperature, and physical resistivity models allowing for small polarization conduction, Mg vacancy conduction and (to later consider the presence of water) hydrogen diffusion. A linear relationship between S-velocity and $1/\log(\text{resistivity})$ emerged by simultaneously varying modal composition, temperature and Mg number between limiting values from two dry xenolith samples, and extreme condition bounds were then obtained by varying water content from 0 ppm through 300 ppm in olivine (while keeping water content constant for the remaining modal components).

Those extreme condition bounds are superimposed to our correlation plot for 100 km depth in Figure 12, revealing that only stations within the Proterozoic Borborema Province fall within the South African extreme condition bounds. Stations within the São Francisco craton clearly fall outside the bounds and must be reconciled thorough either anomalous S-velocities and/or anomalous $1/\log(\text{resistivity})$ values. If S-velocities in the São Francisco craton are larger than expected for the given $1/\log(\text{resistivity})$ then, considering that S-velocity mainly depends on temperature (e.g. Jones et al. [2013]), that would imply that the São Francisco craton lithosphere is colder than expected. However, given the proximity of our stations to the postulated sublithospheric mantle flow under the neighbouring Araripe Basin ([Garcia et al., 2019]; [Simões Neto et al., 2019]), we find this possibility unlikely. If, alternatively, it is the $1/\log(\text{resistivity})$ values that are larger than expected for the given S-velocity values, that would imply abnormally

489 high amounts of water (above 300 ppm) in the cratonic lithosphere. Jones
490 et al. [2013] warn that such high values are to be discarded, and that other
491 conductive processes - perhaps the presence of a minor conducting phase -
492 should be considered instead.

493 *4.2. Implications for imaging an electric Moho*

494 Another important aspect that can be investigated with our structurally
495 consistent S-velocity and resistivity models is the imaging of an electric Moho
496 under our study area. Jones et al. [2013] argue that, from a petrological point
497 of view, a step-like increase in electrical resistivity of at least one order of
498 magnitude would be expected at the crust-mantle boundary if dry and ab-
499 sent of any minor conducting phases. Due to enhanced conduction from
500 sulphides, iron oxides and graphite in the lower crust and water in the lower
501 crust and upper mantle, however, the electric Moho is more likely to happen
502 at the bottom of a conductive layer sandwiched between two resistive layers.
503 Moreover, the resistivity steps at the top and bottom of the conductive layer
504 are supposed to be age-dependent. More in particular, from a compilation
505 of resistivity studies worldwide, Jones et al. [2013] proposed nominal values
506 of 40,000/10,000/50,000 Ωm for Archean terrains, 10,000/1,000/50,000 Ωm
507 for Proterozoic terrains, and 1,000/87.5/5,000 Ωm for Phanerozoic terrains.
508 They note, nonetheless, that exceptions may and do occur. An overall resis-
509 tive ($> 40,000 \Omega\text{m}$) crust below 5 km depth and a moderately resistive upper
510 mantle ($\sim 4,000 \Omega\text{m}$) in the southwestern part of the Archean Slave craton,
511 for instance, renders an electrical Moho characterized as an anomalous resis-
512 tivity decrease instead of the expected resistivity increase.

513 Our results on S-velocity structure (Figure 9) demonstrate consistency in

514 Moho depth and overall S-velocity structure with those reported by Nemocón
 515 et al. [2021] and previous regional receiver function studies [Luz et al., 2015;
 516 Almeida et al., 2015], giving us confidence that our simplified layered struc-
 517 ture succeeds in imaging the seismic Moho. A similar comparison between
 518 the resistivity profiles from the JI presented in this study and those con-
 519 structed from the 3D MT inversion of Garcia et al. [2019] is less successful,
 520 as smoothness constraints in the later prevented from imaging sharp discon-
 521 tinuities. Interestingly, however, we do see a sharp resistivity increase in
 522 our layered profiles at depths coincident with the seismic Moho under sta-
 523 tions AR01, AR03, AR04, AR05, AR07, AR08, AR09, NBPB, NBPN and,
 524 perhaps, AR06. Moreover, in all cases, the resistivity increase happens im-
 525 mediately under a portion of conductive lower crust of variable thickness, as
 526 expected from global compilations. As pointed out in Jones et al. [2013],
 527 these structures are difficult to resolve through DC resistivity experiments
 528 due to equivalency of structures with same vertically integrated conductance,
 529 but MT studies - which fare better in resolution and equivalency properties
 530 - should be better suited for resolving them. It thus seems that, in our case,
 531 the removal of restrictive smoothness constraints in the layered resistivity
 532 profiles, along with structural help from the collocated seismic data, helps
 533 identify an electrical Moho under most of the sites.

534 *4.3. Imaging of lithospheric thickness*

535 One of the main results of the 3D MT modeling of Garcia et al. [2019]
 536 was the imaging of a thin lithosphere (~ 120 km) centered under the Araripe
 537 basin. The study did not have enough resolution to image the bottom of the
 538 lithosphere under the São Francisco craton or the Borborema Province, but

539 a resistive lithosphere was imaged there down to (at least) 160 km depth.
 540 Indeed, resistivity-depth profiles from the 3D model displayed in Figure 9
 541 show a smooth decrease in resistivity from over 100 Ωm at ~ 50 km depth to
 542 just $\sim 50\Omega\text{m}$ at 150 km depth; considering expected values of just $10\Omega\text{m}$ for
 543 the asthenosphere, we can observe that our MT inversion is clear in imaging
 544 lithospheric thickness only beneath station AR03 - flanking the Araripe
 545 Basin - with values between 19 and 10 Ωm at $> 96\text{km}$ depth. The seismic
 546 study of Nemocón et al. [2021], on the other hand, did report evidence of a
 547 gradational S-velocity drop centered at around 150 km depth in those regions
 548 that was interpreted as the lithosphere-asthenosphere boundary.

549 One might thus expect that, similar to what we observed with the electric
 550 Moho, the structural coupling of collocated MT and seismic data might help
 551 image the bottom of the lithosphere in both resistivity and seismic velocity
 552 after a joint inversion. Unfortunately, the dispersion velocity utilized in our
 553 JI - which is the dataset that constrains absolute velocity in the S-velocity
 554 profiles - does not have enough penetration at upper mantle depths. The
 555 maximum period in our average phase velocity curve is 45 s, which has a
 556 peak sensitivity at about 40 km depth. Recall that Nemocón et al. [2021]
 557 utilized group velocities down to 100 s, which do have some sensitivity at the
 558 target depths. Joint imaging of an LAB would be a plausible outcome of any
 559 JI of seismic and MT data, but further analysis - using longer periods and,
 560 ideally, with a full 3D JI - would be necessary for it to succeed.

561 5. Conclusions

562 Joint inversion of collocated magnetotelluric and teleseismic data in the
563 western Borborema Province and northern São Francisco craton for com-
564 monly parameterized, 1D layered structures has successfully accounted for
565 both sets of observations in spite of significant geological complexity. Our
566 models largely agree with previous results from separate inversion of the same
567 datasets, thus providing strong evidence that the imaged subsurface varia-
568 tions in S-velocity and resistivity are related to common causative structures.
569 Although our jointly inverted S-velocity and resistivity-depth profiles have
570 failed to constrain any clear relationships between S-velocity and density at
571 any depth, they have succeeded in characterizing an electric Moho under
572 most of the sites as a marked resistivity increase right under a conductive
573 lower crust. Moreover, the depth of the electrical Moho has shown excellent
574 agreement with the depth of the seismic Moho in the corresponding S-velocity
575 profiles, demonstrating beneficial cross-feeding between magnetotelluric and
576 seismological data.

577

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589 **Data availability**

590 The data used in this paper are available at <http://rsbr.on.br/request.html>
591 and <http://www.pegbr.on.br/>, and the MT data are also available at the
592 CSIC's online repository [Garcia and Julià, 2016].

593 **References**

- 594 Abubakar, A., Gao, G., Habashy, T., Liu, J., 2012. Joint inversion approaches
595 for geophysical electromagnetic and elastic full-waveform data. *Inverse*
596 *Problems*. 28. doi:doi: 10.1088/0266-5611/28/5/055016.
- 597 Almeida, Y.B., Julià, J., Frassetto, A., 2015. Crustal architecture of the
598 Borborema Province, NE brazil, from receiver function CCP stacks: Im-
599 plications for Mesozoic stretching and cenozoic uplift. *Tectonophysics* 649,
600 68–80. doi:doi: 10.1016/j.tecto.2015.03.001.
- 601 Deb, K., Pratap, A., Agarwal, S., Meyarivan, T., 2002. A fast and eli-
602 tist multiobjective genetic algorithm: Nsga [U+2010]ii. *IEEE Trans. Evol.*
603 *Comput.* 6, 182–197. doi:doi: 10.1109/4235.996017.
- 604 Dziewonski, A., Anderson, D., 1981. Preliminary reference earth model.
605 *Phys. Earth Planet. Inter.* 25, 297–356.

- 606 Gallardo, L.A., Meju, M.A., 2003. Characterization of heterogeneous
607 near[U+2010]surface materials by joint 2d inversion of dc resistiv-
608 ity and seismic data. *Geophys. Res. Lett.* 30, 1658. doi:doi:
609 10.1029/2003GL017370.
- 610 Gao, G., Abubakar, A., Habashy, T., 2012. Joint petrophysical inversion
611 of electromagnetic and full-waveform seismic data. *Geophysics* 77, 53–.
612 doi:doi: 10.1190/geo2011-0157.1.
- 613 Garcia, X., Jones, A.G., 2002. Extended decomposition of MT data,
614 in *Three-Dimensional Electromagnetics*. Elseiver , 235–250doi:doi:
615 0.1016/S0076-6895(02)80095-6.
- 616 Garcia, X., Julià, J., 2016. Bodes (v.1.0) [dataset]. doi:doi: 10.20350/digi-
617 talCSIC/14694.
- 618 Garcia, X., Julià, J., Nemocón, A.M., Neukirch, M., 2019. Lithospheric
619 thinning under the Araripe Basin (NE Brazil) from a long-period magne-
620 totelluric survey: Constraints for tectonic inversion. *Gondwana Research*
621 68, 174–184. doi:doi: 10.1016/j.gr.2018.11.013.
- 622 Hansen, P.C., 1992. Analysis of discrete ill-posed problems by means of the
623 l-curve. *SIAM Rev.* 34, 561–580.
- 624 Herrmann, R., Ammon, C., 2002. *Surface Waves, Receiver Functions and*
625 *Crustal Structure*.
- 626 Jones, A.G., R., M.M., L., E.R., Javier, F., 2013. Velocity-conductivity rela-
627 tions for cratonic lithosphere and their application: Example of Southern

628 Africa. *Geochemistry, Geophysics and Geosystems* (G3, AGU) 14, 423–
629 436.

630 Kelbert, A., Meqbel, N., Egbert, G.D., Tandon, K., 2014. ModEM: A mod-
631 ular system for inversion of electromagnetic geophysical data. *Computers*
632 & *Geosciences* 66, 40–53. doi:doi: 10.1016/j.cageo.2014.01.010.

633 Kirkby, A., Zhang, F., Peacock, J., Hassan, R., D.J., 2019. The MTPy soft-
634 ware package for magnetotelluric data analysis and visualisation. *Journal*
635 of *Open Source Software* 4. doi:doi: 10.21105/joss.01358.

636 Krieger, L., Peacock, J., 2014. MTPy: A Python toolbox for mag-
637 netotellurics. *Computers and Geosciences* 72, 167–175. doi:doi:
638 10.1016/j.cageo.2014.07.013.

639 Ligorria, J.P., Ammon, C.J., 1999. Iterative deconvolution and receiver-
640 function estimation. *Bulletin of the Seismological Society of America* 89,
641 1395–1400.

642 Luz, R.M.N., Julià, J., do Nascimento, A.F., 2015. Bulk crustal properties of
643 the Borborema Province, NE Brazil, from P-wave receiver functions: Im-
644 plications for models of intraplate Cenozoic uplift. *Tectonophysics* 644–645,
645 81–91. doi:doi: 10.1016/j.tecto.2014.12.017.

646 McMechan, G.A., Yedlin, M.J., 1981. Analysis of dispersive waves by wave
647 field transformation. *Geophysics* 46. doi:doi: 10.1190/1.1441225.

648 Moorkamp, M., Jones, A., Fishwik, S., 2010. Joint inversion of receiver
649 functions, surface wave dispersion and magnetotelluric data. *Journal of*
650 *Geophysical Research*. 115. doi:doi: 10.1029/2009JB006369.

- 651 Moorkamp, M., Jones, A.G., Eaton, D.W., 2007. Joint inversion of teleseis-
652 mic receiver functions and magnetotelluric data using a genetic algorithm:
653 Are seismic velocities and electrical conductivities compatible?. *Geophys-*
654 *ical Research Letters*. 34. doi:doi: 10.1029/2007gl030519.
- 655 Nemocón, A.M., Julià, J., Garcia, X., 2021. Lithospheric structure of the
656 western Borborema Province from receiver functions and surface-wave dis-
657 persion: Implications for basin inversion. *Tectonophysics* .
- 658 Simões Neto, F., Julià, J., Schimmel, M., 2019. Upper-mantle structure of
659 the Borborema Province, NE Brazil, from P-wave tomography: implica-
660 tions for rheology and volcanism. *Geophys. J. Int.* 216, 231–250. doi:doi:
661 10.1093/gji/ggy421.

Number of layers	RMS MT	RMS RF	RMS SW	ID best-Pareto	Run-time
7	3.05378	5.74386	0.658468	398	13h49m
10	2.86606	5.5276	0.531185	37	17h48m
14	3.49783	6.1395	0.296851	238	19h59m

Table 1: *RMS misfit for 7, 10 and 14 layers for station AR01. Note: Computer specifications are: Processor 6a Gen Intel Core i7, Graphics NVIDIA GeForce, 16G RAM.*

<i>Layer</i>	1	2	3	4	5	6	7	8	9	10
m_1^{min}	0.1	1	1	1	5	5	10	10	20	20
δ_1	0.1	0.1	0.5	0.5	0.5	0.5	0.5	0.5	1.0	1.0
n_1	6	6	6	6	6	6	6	6	6	0
m_2^{min}	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0
δ_2	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
n_2	4	4	4	4	4	4	4	4	4	4
m_3^{min}	2.5	3.2	3.2	3.2	3.2	3.5	3.5	3.5	3.5	3.5
δ_3	0.0	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05	0.05
n_3	0	6	6	6	6	6	6	6	6	6

Table 2: *Parameter encoding for station AR01. Index i values correspond to: $i = 1$ for thickness, $i = 2$ for resistivity, $i = 3$ for S -velocity.*

STATION	LAT	LON	ALT(m a.s.l.)	#RF	#Freq	mode-MT	Run-time	ID Par.	RMS MT	RMS RF	RMS SW
NBPB	-5.5432	-39.5836	260	24	22	YX	20h 58m 58s	223	2.5616	4.9397	0.225983
AR01	-6.1682	-39.4098	305	21	15	XY	17h 48m 1s	37	2.8660	5.5276	0.531185
AR02	-6.6419	-39.3402	341	24	19	XY	19h 8m 53s	47	4.21214	5.25789	0.774657
AR03	-7.0466	-39.4958	479	18	20	XY	19h 11m 44s	33	1.9859	10.5744	0.766649
AR04	-7.5758	-39.6746	493	19	20	XY	14h 31m 44s	214	2.647	6.099	0.2966
AR05	-8.0932	-39.8983	441	19	20	XY	16h 18m 46s	183	2.7815	5.34494	0.324653
AR06	-8.6305	-40.2249	454	19	19	XY	13h 51m 44s	249	1.62643	5.5302	0.232836
AR08	-9.1450	-40.3727	394	8	20	YX	9h 32m 42s	208	3.23448	7.14573	0.64079
AR07	-9.6154	-40.3837	405	20	20	XY	17h 16m 42s	138	3.88699	5.24251	0.392208
AR09	-10.2136	-40.1954	678	12	20	YX	10h 38m 1s	78	1.72699	5.25566	1.00266
NBPN	-10.8468	-40.1988	387	18	18	XY	13h 40m 56s	317	1.4719	5.33278	0.211906

Table 3: Localization, fit mode parameter for MT data and RMS value for all BODES stations.

Figure Captions

Fig. 1. Geologic map of the Boroborema Province with the magnetotelluric and seismic stations of the BODES Experiment (red triangles) and the Precambrian network of shear zones (solid lines) superimposed.

Fig. 2. Rayleigh wave dispersion. Phase velocities were obtained through the array-processing approach of McMechan and Yedlin [1981], as implemented in the Computer Programs in Seismology package of Herrmann and Ammon [2002]. The red color corresponding to the largest stack value with maximum amplitude and the ridge defines the location of the dispersion curve.

Fig. 3. Example of preprocessing MT data. The exclusion of high frequency and outliers in MT data is necessary to remove sections more affected by the complex 3D geology of the study area.

Fig. 4. Roughness-misfit trade-off curves (top left), for inversion of MT data alone at station AR01, performed for inversions with 7, 8, 9, 10, 11, 12 and 14 layers and their optimum models MT (top right). Minimum RMS as function of number of layers for MT data (below).

Fig. 5. Trade-off plots for the data station AR01. Each plot show two objective function values and color the points by third objective function value. The red star is the best model selected.

Fig. 6. The seismic (left) and electric (right) parameters for station AR01. Plots for all models in Pareto front (green). The best model selected in trade-off, Figure 5, (black); comparisons in left panels with the PREM model (blue) and joint inversion with a previous methodology (red).

Fig. 7. The observed (black) and predicted (red) data for site AR01.

687 Top panel for dispersion curves and bottom panels for resistivity and phase
688 from magnetotelluric data.

689 **Fig. 8.** The observed (black) and predicted (red) data for Receiver
690 functions in site AR01.

691 **Fig. 9.** The seismic (top) and electric (bottom) models by JI of MT an
692 seimological data for all sites. The best model selected in trade-off (black);
693 comparisons with a previous methodologies (red), for RF with Nemocón et al.
694 [2021] and for MT with [Garcia et al., 2019].

695 **Fig. 10.** Resistivity-Velocity relationships between seismic S-velocity
696 and $1/\log(\rho)$ from our JI results. Correlation plots are color-coded by terrain
697 age.

698 **Fig. 11.** Similar correlation plots between seismic S-velocity and $1/\log(\rho)$
699 from our previous results Nemocón et al. [2021] and [Garcia et al., 2019].

700 **Fig. 12.** Resistivity-Velocity relationships for 100 km depth from our JI
701 results compared with Jones et al. [2013] (blue stars).

702 **Figures**

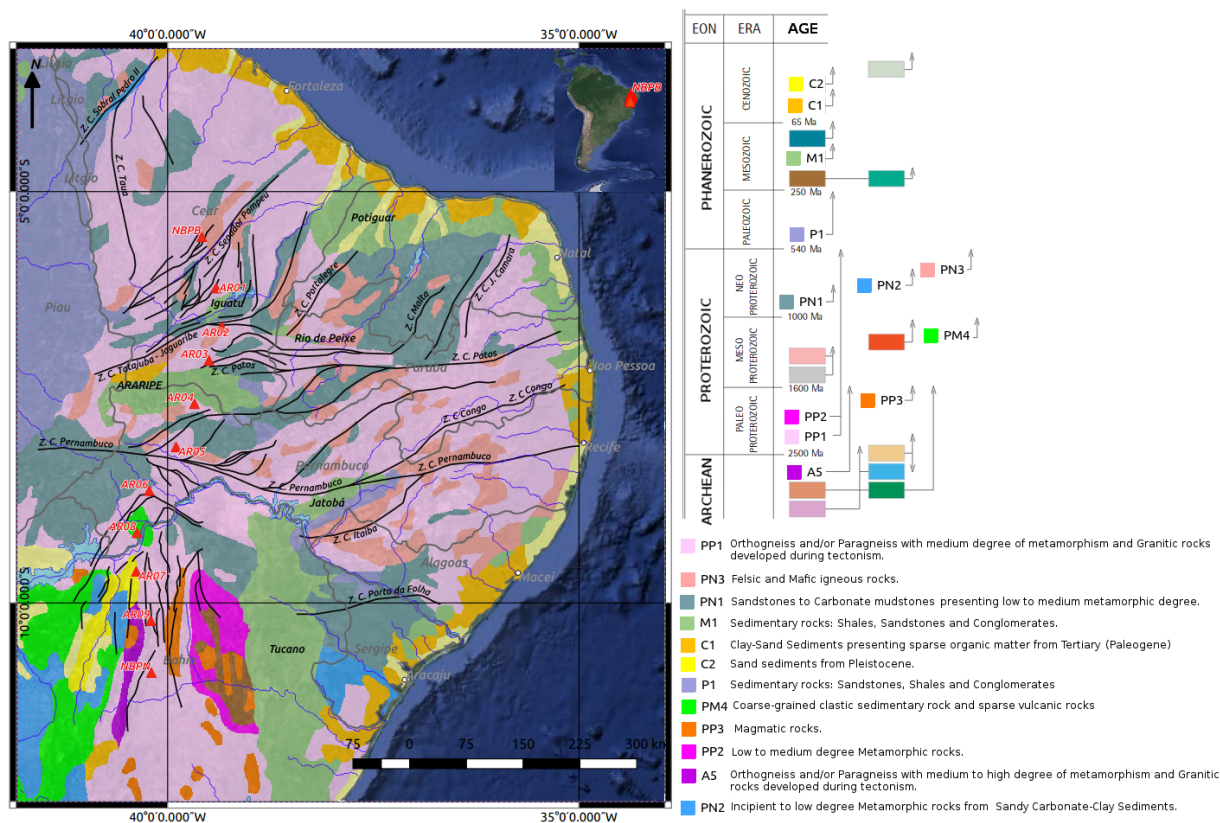


Figure 1: Geologic map of the Borborema Province with the magnetotelluric and seismic stations of the BODES Experiment (red triangles) and the Precambrian network of shear zones (solid lines) superimposed.

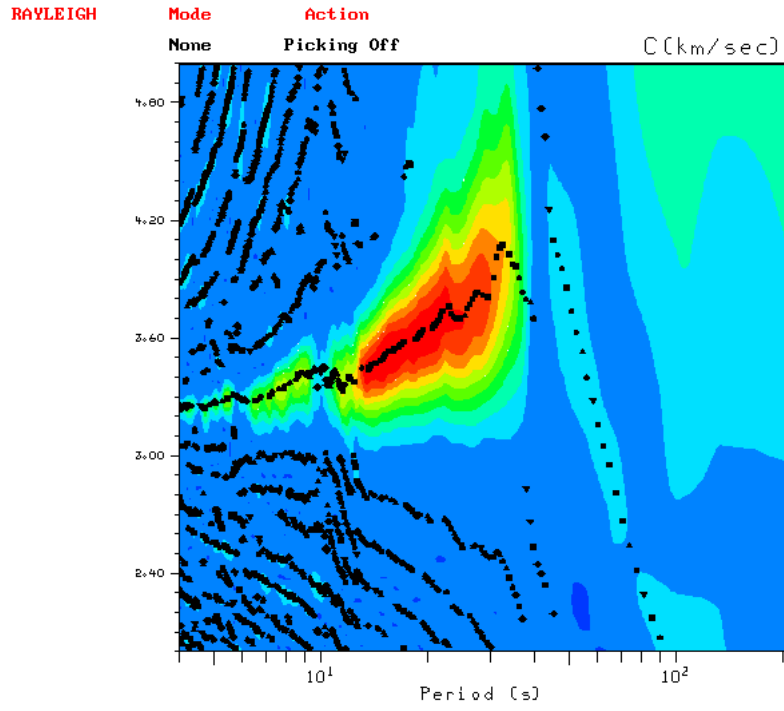


Figure 2: *Rayleigh wave dispersion. Phase velocities were obtained through the array-processing approach of Mcmechan and Yedlin [1981], as implemented in the Computer Programs in Seismology package of Herrmann and Ammon [2002]. The red color corresponding to the largest stack value with maximum amplitude and the ridge defines the location of the dispersion curve.*

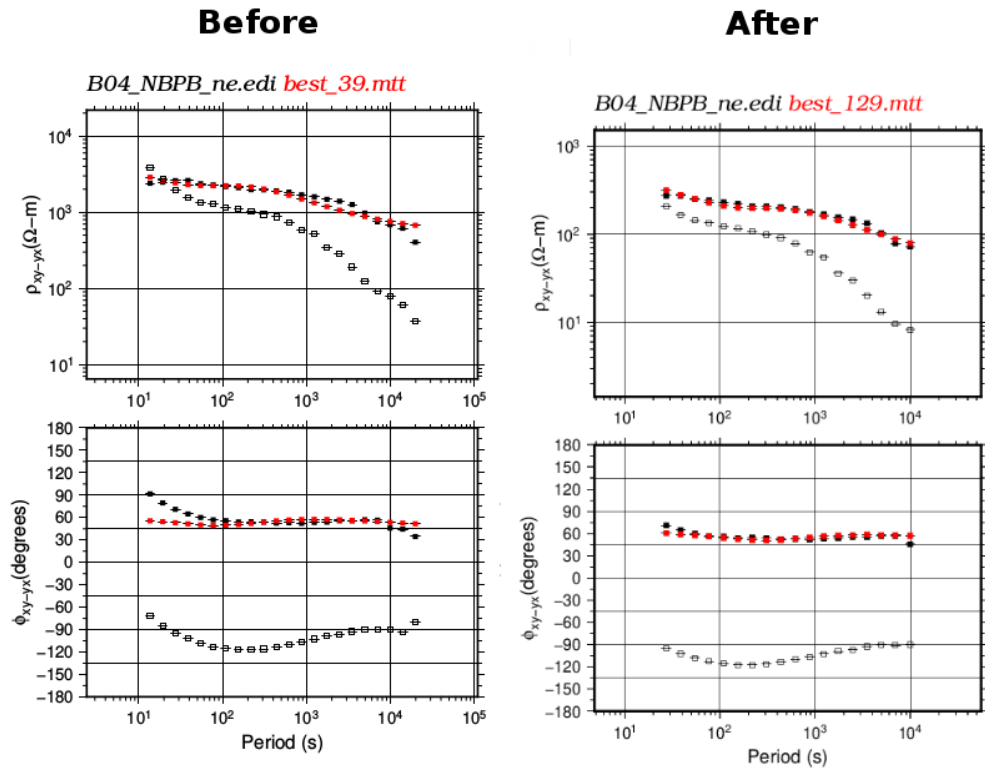


Figure 3: *Example of preprocessing MT data. The exclusion of high frequency and outliers in MT data is necessary to remove sections more affected by the complex 3D geology of the study area.*

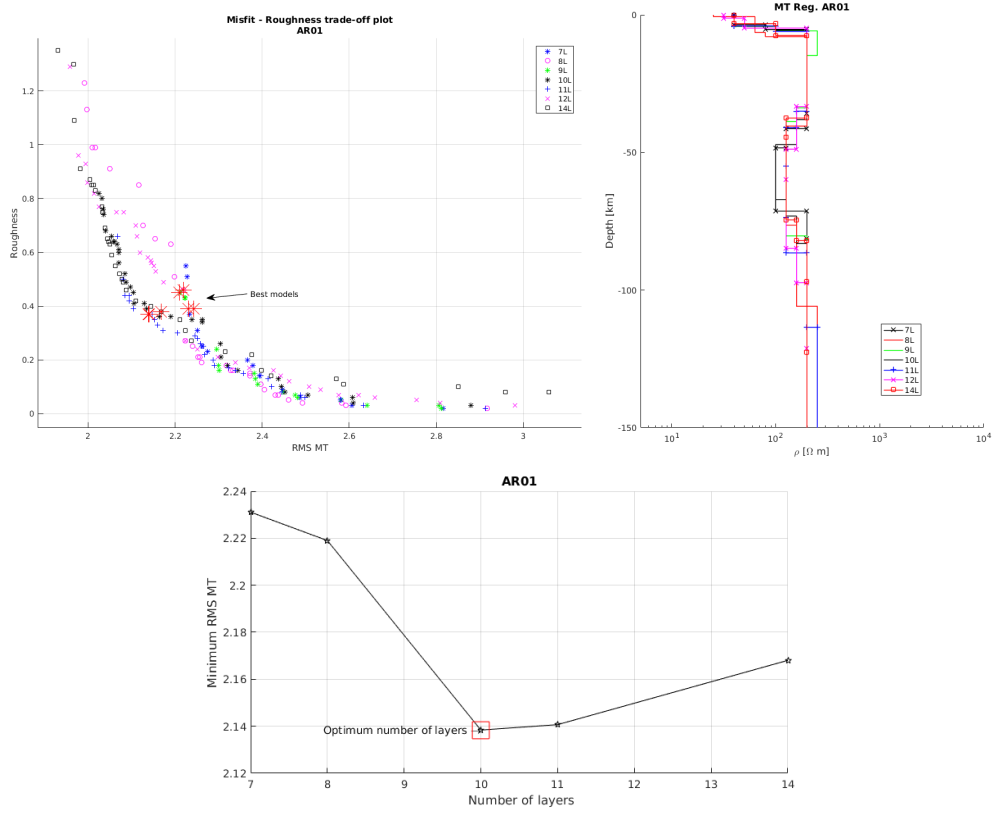


Figure 4: Roughness-misfit trade-off curves (top left), for inversion of MT data alone at station AR01, performed for inversions with 7, 8, 9, 10, 11, 12 and 14 layers and their optimum models MT (top right). Minimum RMS as function of number of layers for MT data (below).

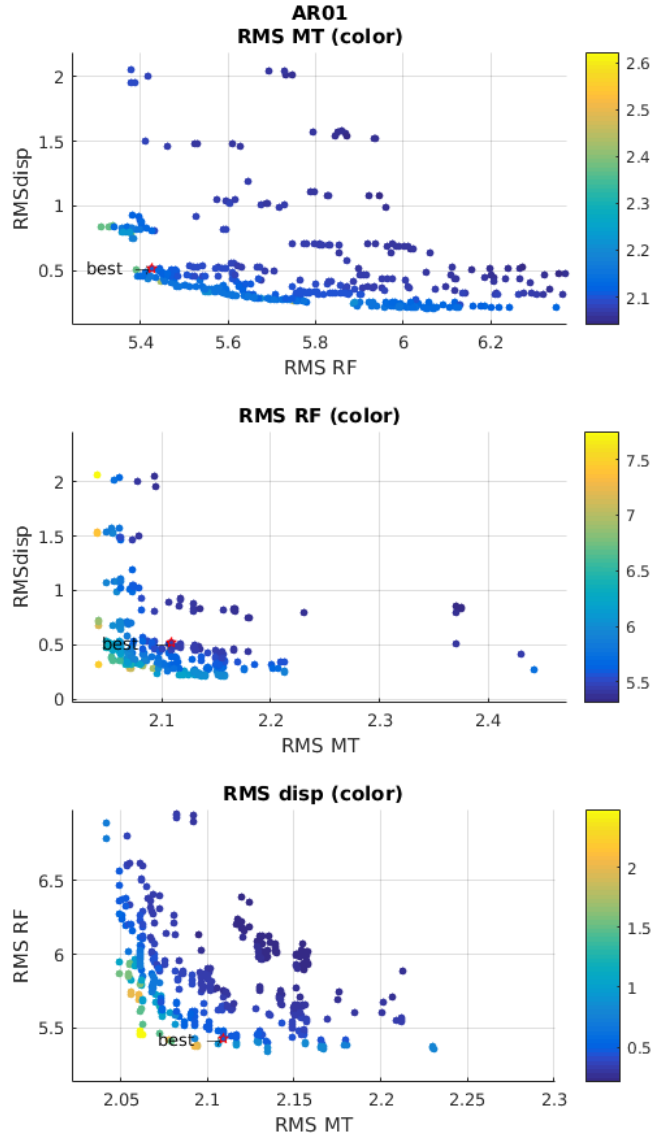


Figure 5: Trade-off plots for the data station AR01. Each plot show two objective function values and color the points by third objective function value. The red star is the best model selected.

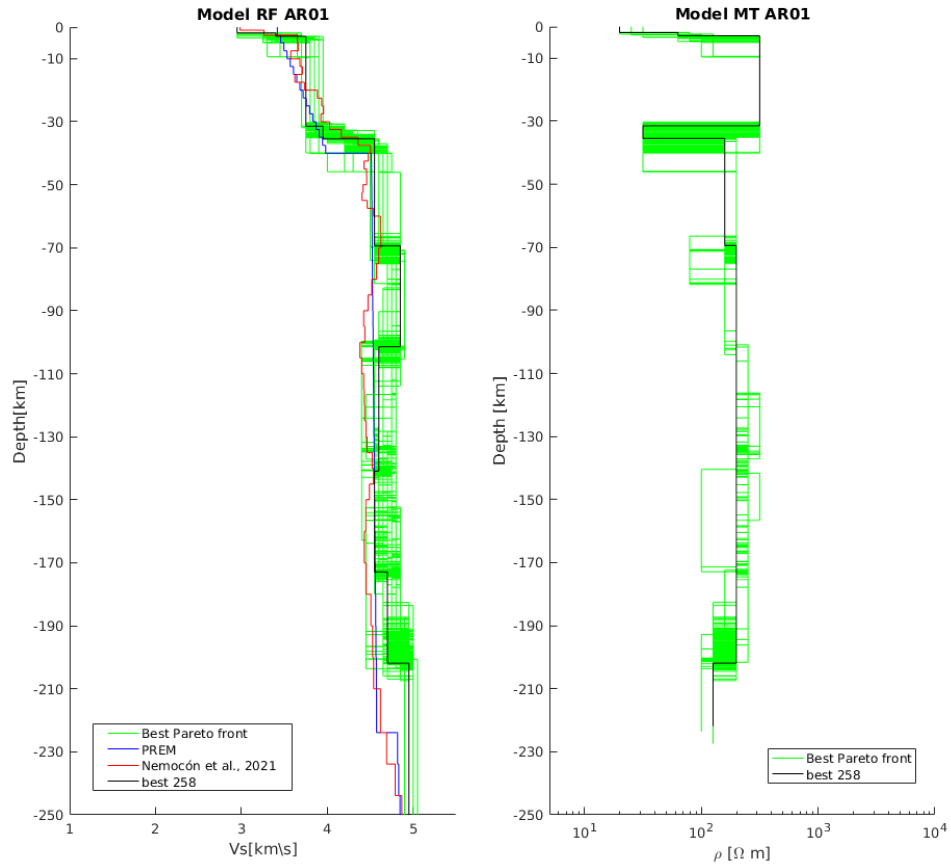


Figure 6: *The seismic (left) and electric (right) parameters for station AR01. Plots for all models in Pareto front (green). The best model selected in trade-off, Figure 5, (black); comparisons in left panels with the PREM model (blue) and joint inversion with a previous methodology (red).*

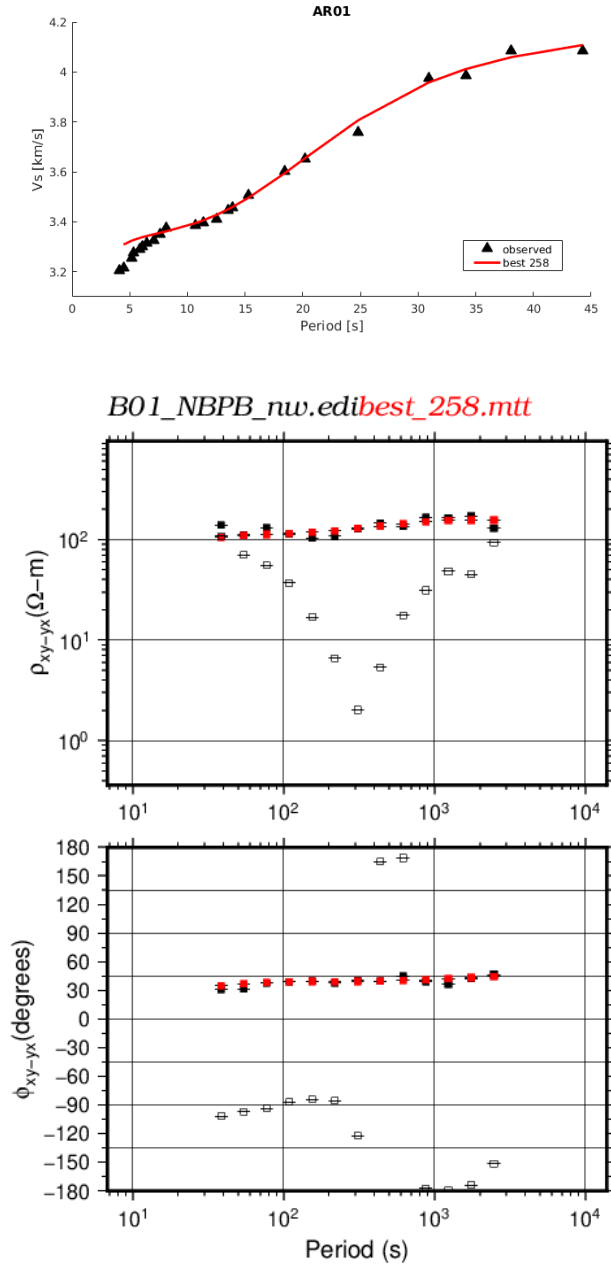


Figure 7: The observed (black) and predicted (red) data for site AR01. Top panel for dispersion curves and bottom panels for resistivity and phase from magnetotelluric data.

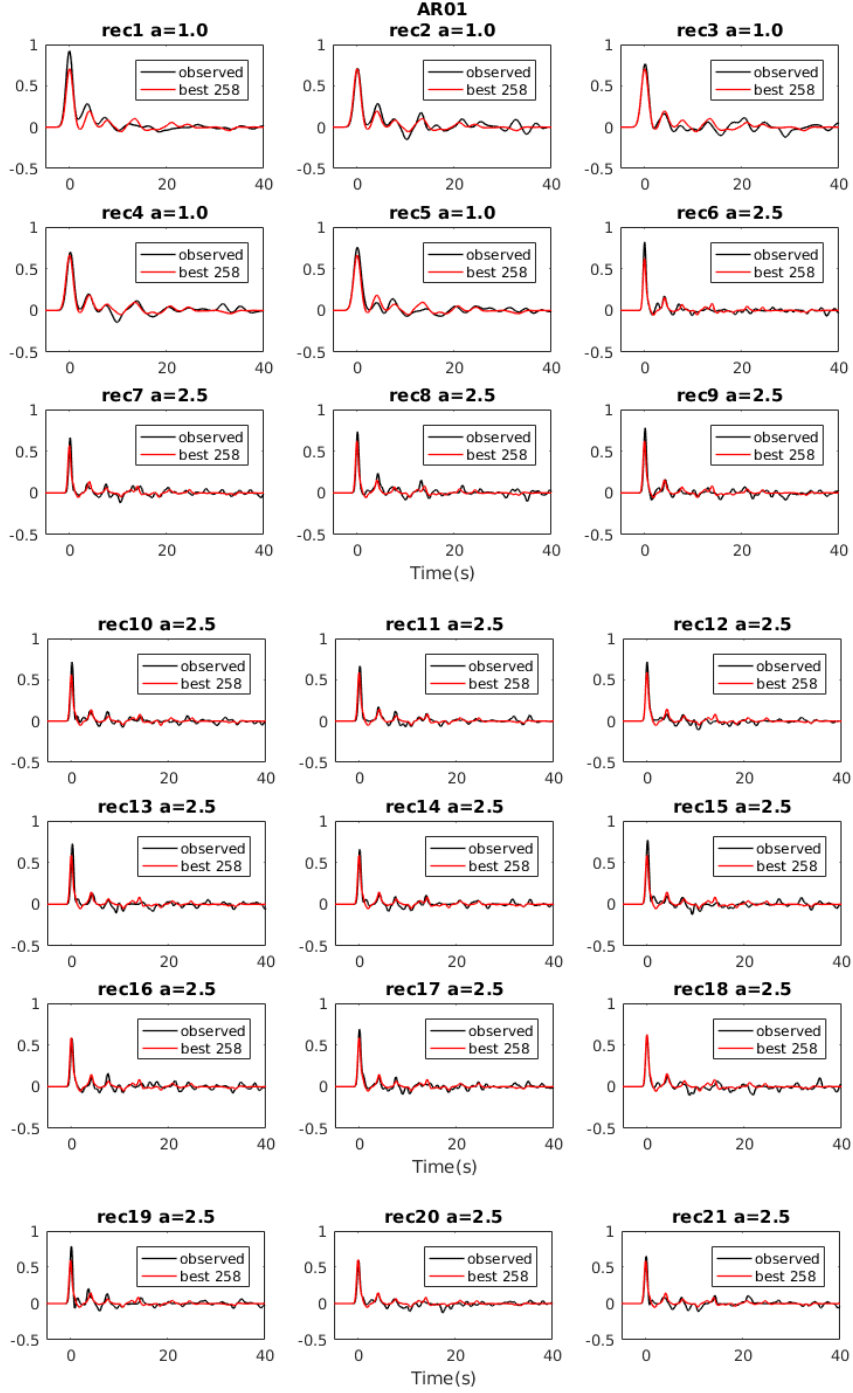


Figure 8: The observed (black) and predicted (red) data for Receiver functions in site AR01.

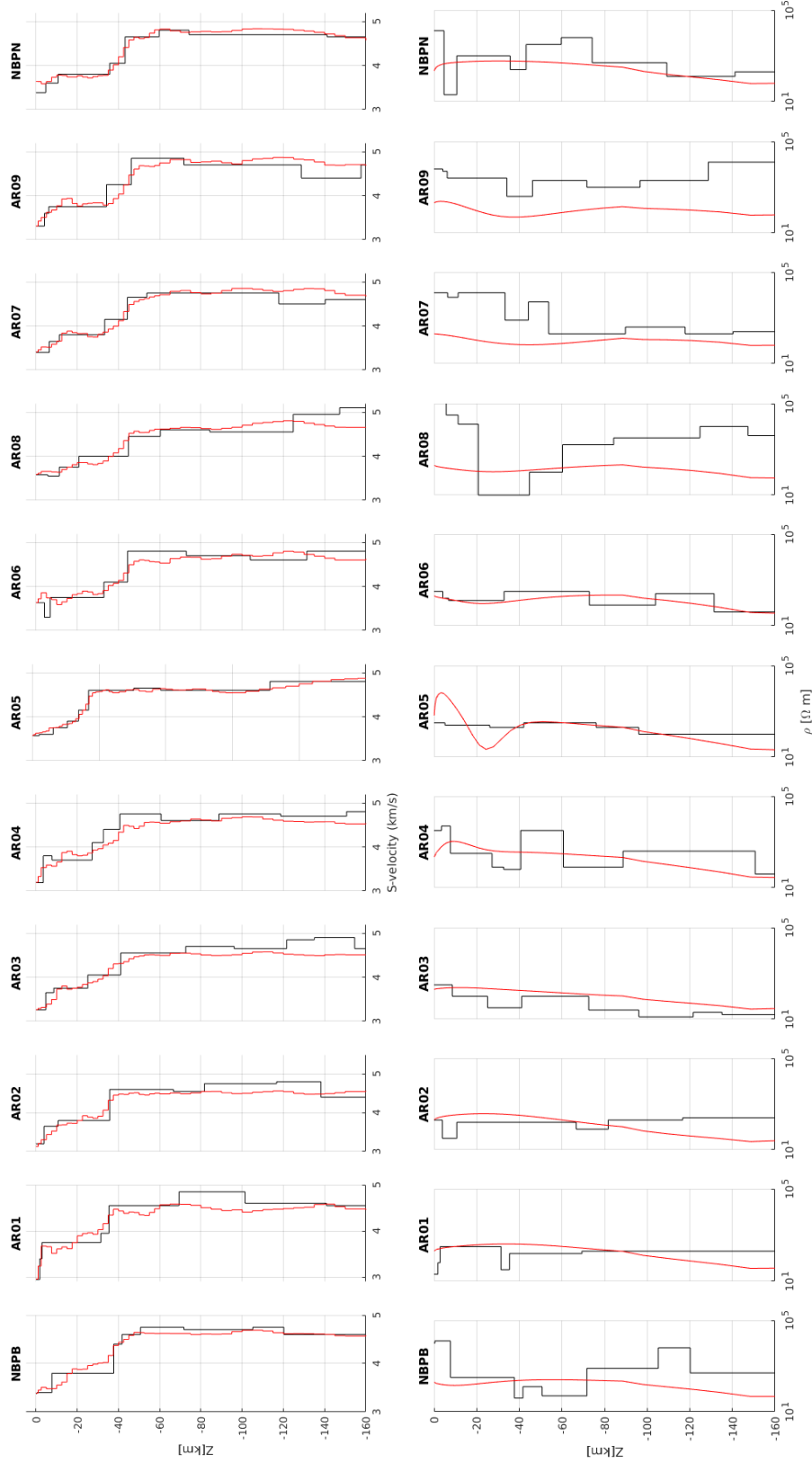


Figure 9: The seismic (top) and electric (bottom) models by *JI* of *MT* an seimological data for all sites. The best model selected in trade-off (black); comparisons with a previous methodologies (red), for *RF* with Nemocón et al. [2021] and for *MT* with Garcia et al. [2019].

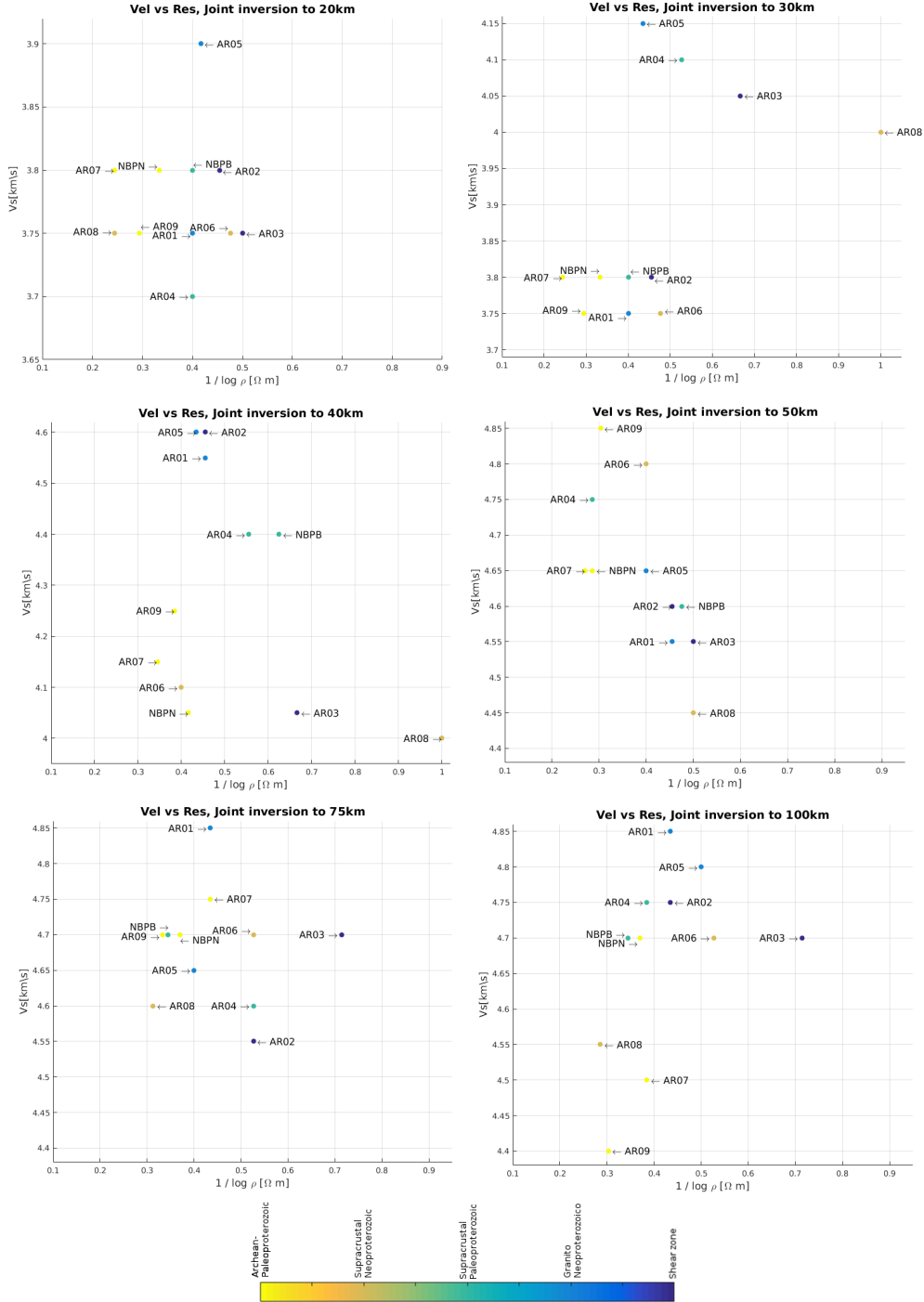
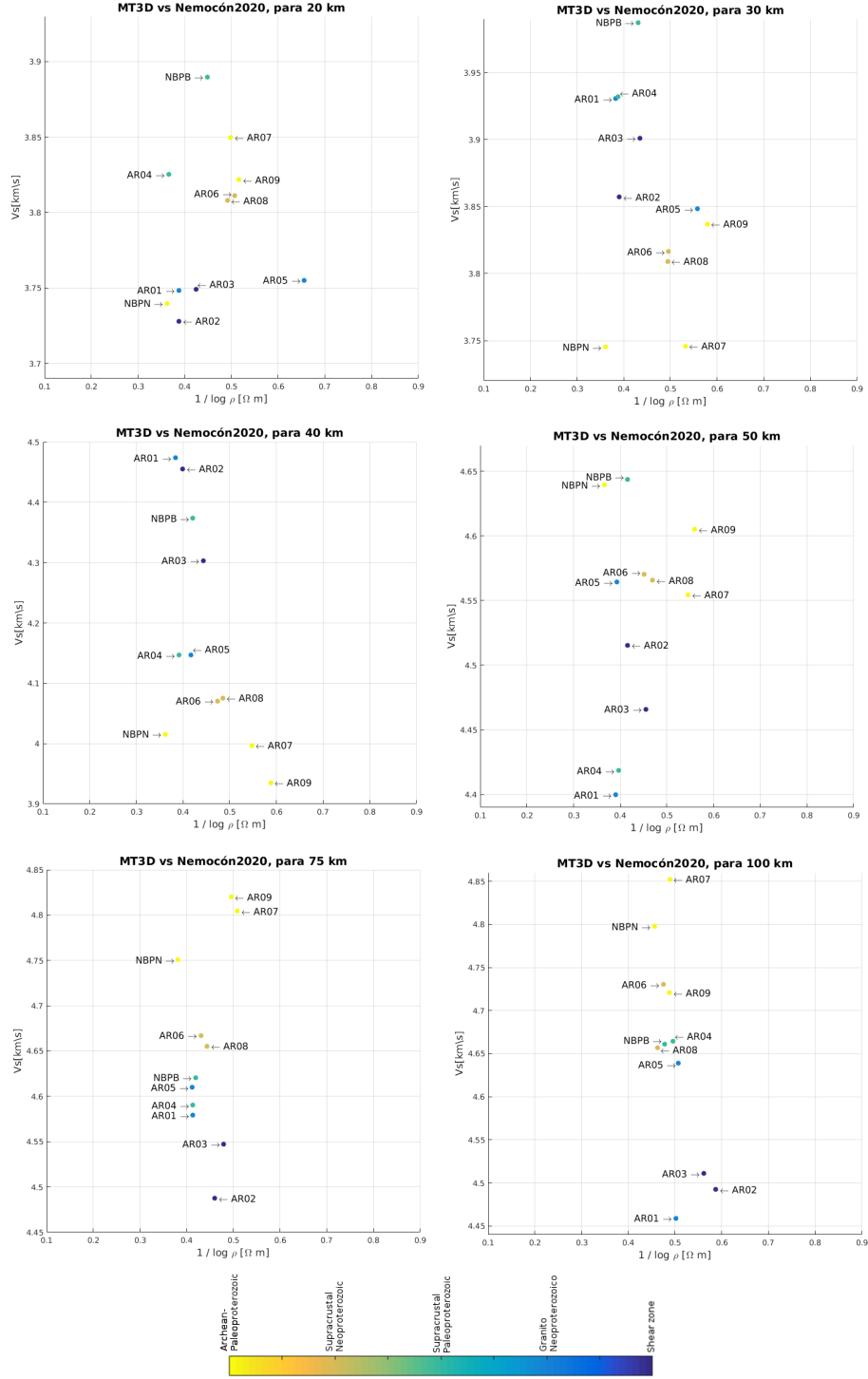


Figure 10: Resistivity-Velocity relationships between seismic S -velocity and $1/\log(\rho)$ from our JI results. Correlation plots are color-coded by terrain age.



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Figure 11: Similar correlation plots between seismic S -velocity and $1/\log(\rho)$ from our previous results Nemocón et al. [2021] and [Garcia et al., 2019].

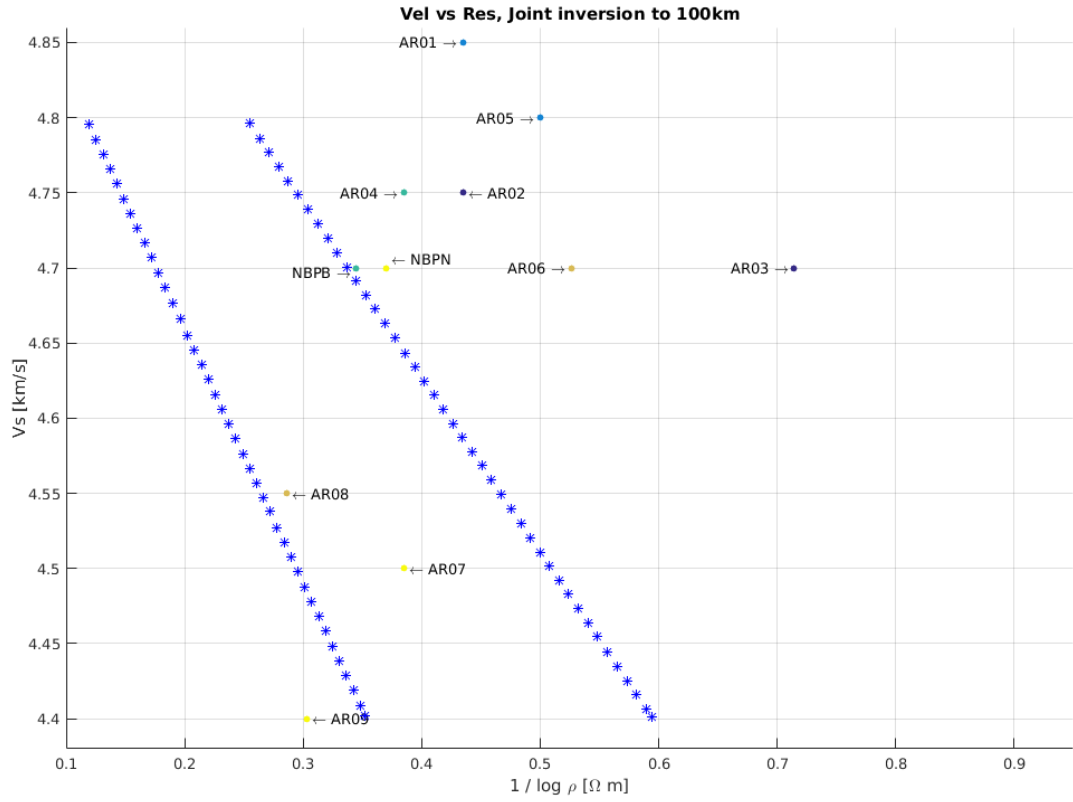


Figure 12: Resistivity-Velocity relationships for 100 km depth from our JI results compared with Jones et al. [2013] (blue stars).

Chapter 6

Conclusions and future work

The manuscripts presented in the previous two chapters have reported on results produced through two different approaches involving seismological and electromegnetic datasets, in order to develop constraints on Earth structure in NE Brazil. The first paper contributed to further develop the tectonic framework that led to the present-day elevated topography of the Araripe Basin. The main results included: (i) the crust is typically 34-38 km thick under the western Province, whereas it is 43-46 km thick under the northern craton; (ii) no apparent crustal thinning is observed under the Araripe Basin, due to the presence of mafic underplate, inferred from a 10 – 14 km thick, high-velocity (4.1 - 4.3 km/s) in the lowermost crust; and (iii) the lithosphere under the basin is thin (~ 120 km) when compared to the regions flanking the basin. It must be stressed that such conclusions could only be reached because seismological interpretation was aided by the 3D resistivity model developed from the inversion of collocated magnetotelluric data (Garcia et al., 2019).

The second paper focuses on the imaging of the electric Moho under NE Brazil, using an actual joint inversion of collocated magnetotelluric and teleseismic data. In that study, we were able to demonstrate the structural consistency between S-velocity and resistivity subsurface structures; it is worth mentioning that such consistency was achieved through

common parameterization of simple (1D) layered structures in a complex (3D) structural setting. As a result of such common parameterization, we successfully characterized a resistivity jump at the bottom of a conductive lower crust at depths coincident with the seismic Moho. Such resistivity jump is identified as the electric Moho.

The main drawback from the joint inversion of teleseismic and magnetotelluric data was the lack of long periods in the phase-velocity curves, which prevented constraining lithospheric and sub-lithospheric structure with both datasets. Future research that could be performed in the short-run would thus include the development of station-dependent dispersion curves at longer periods, as such curves might potentially provide tighter constraints on the depth to the LAB and help refine existing inversion models for the Araripe basin. Additional, longer term improvements would include: (i) performing a 3D joint inversion of receiver functions, dispersion curves and magnetotelluric data; (ii) explore different types of coupling between seismic and magnetotelluric data during joint inversion; and (iii) obtain a collocated gravimetric profile in the study region and carry an expanded joint inversion of magnetotelluric, gravity and seismic data.

Bibliography

- A. Abubakar, G. Gao, T. Habashy, and J. Liu. Joint inversion approaches for geophysical electromagnetic and elastic full-waveform data. Inverse Problems., 28(5), 2012. doi: 10.1088/0266-5611/28/5/055016.
- F. Alkmim, S. Marshak, and M. Fonseca. Assembling west gondwana in the neoproterozoic: clues from the são francisco craton region, brazil. Geology, 29:319–322, 2001. doi: 10.1130/0091-7613(2001)029<0319:AWGITN>2.0.CO;2.
- F. F. Alkmim and M. A. Martins-Neto. Proterozoic first-order sedimentary sequences of the são francisco craton, eastern brazil. Marine and Petroleum Geology, 33:127–139, 2012. doi: 10.1016/j.marpetgeo.2011.08.011.
- F. F. M. Almeida, Y. Hasui, B. B. Brito Neves, and R. A. Fuck. Brazilian structural provinces: An introduction. Earth Science Reviews, 17:1–29, 1981. ISSN 0012-8252. doi: 10.1016/0012-8252(81)90003-9.
- Y. B. Almeida, J. Julià, and A. Frassetto. Crustal architecture of the borborema province, NE brazil, from receiver function CCP stacks: Implications for mesozoic stretching and cenozoic uplift. Tectonophysics, 649:68–80, 2015. ISSN 0040-1951. doi: 10.1016/j.tecto.2015.03.001.
- C. J. Ammon. The isolation of receiver effects from teleseismic p waveforms. Bulletin

- of the Seismological Society of America, 81(6):2504–2510, 1991. ISSN 0037-1106, 1943-3573.
- C. J. Ammon, G. E. Randall, and G. Zandt. On the nonuniqueness of receiver function inversions. Journal of Geophysical Research, 95:15303–15318, 1990. ISSN 2156-2202. doi: 10.1029/JB095iB10p15303.
- M. Assine. Araripe basin [bacia do araripe]. Boletim de Geociencias da Petrobras, 15(2): 371–389, 2007. ISSN 01029304. cited By 81.
- I. Bianchi, J. Park, N. Agostinetti, and V. Levin. Mapping seismic anisotropy using harmonic decomposition of receiver functions: An application to northern apennines, italy. Journal of Geophysical Research, 115:1–14, 2010. doi: 10.1029/2009JB00706.
- B. Brito Neves, R. Fuck, U. Cordani, and F. Thomaz. Influence of basement structures on the evolution of the major sedimentary basins of brazil: A case of tectonic heritage. Journal of Geodynamics, 1(3):495–510, 1984. ISSN 02643707. doi: 10.1016/0264-3707(84)90021-8.
- L. Cagniard. Basic theory of the magnetotelluric method pf geophysical prospecting. Geophysics, 18:605–635, 1953.
- J. M. Carcione, B. Ursin, and J. I. Nordskag. Cross-property relations between electrical conductivity and the seismic velocity of rocks. Geophysical Research Letters, 72, September 2007. doi: 10.1190/1.2762224.
- T. Cavaliere and A. Jones. On the identification of a transition zone in electrical conductivity between the lithosphere and asthenosphere: a plea for more precise phase data. Journal of Geophysics - Zeitschrift fur Geophysik, 55(1):23–30, 1984.
- N. I. Christensen and W. D. Mooney. Seismic velocity structure and composition of the

- continental crust: A global view. Journal of Geophysical Research: Solid Earth, 100: 9761–9788, 1995. ISSN 2156-2202. doi: 10.1029/95JB00259.
- J. Coimbra, M. Arai, and A. Carreño. Biostratigraphy of lower cretaceous microfossils from the araripe basin, northeastern brazil. Geobios., 35:687–698, 2002.
- D. A. Coley. An Introduction to Genetic Algorithms for Scientists and Engineers. World Scientific Publishing Co. Pte. Ltd., P O Box 128, Farrer Road, Singapore 912805., 1999. ISBN 98 1-02-3602-6.
- S. Cruz and F. F. Alkmim. The tectonic interaction between the Paramirim aulacogen and the Araçuaí belt, São Francisco craton region, Eastern Brazil. Anais da Academia Brasileira de Ciências, 78:151–173, 2006.
- A. Danderfer, B. De Waele, A. Pedreira, and H. Nalini. New geochronological constraints on the geological evolution of Espinhaço basin within the São Francisco craton e Brazil. Precambrian Research, 170:116–128, 2009.
- D. L. de Castro and R. M. G. Branco. Caracterização da arquitetura interna das bacias do vale do cariri (ne do brasil) com base em modelagem gravimétrica 3-d. Rev. Bras. Geof., 17(2–3), November 1999. doi: 10.1590/S0102-261X1999000200003.
- R. M. D. De Matos. The Northeast Brazilian rift system. Tectonics, 11(4):766–791, 1992. ISSN XXXX-XXXX. doi: 10.1029/91TC03092.
- R. M. D. De Matos. History of the northeast Brazilian rift system: kinematic implications for the break-up between Brazil and West Africa. Geological Society, London, Special Publications, 153(1):55–73, 1999. ISSN 0305-8719. doi: 10.1144/GSL.SP.1999.153.01.04.
- K. Deb, A. Pratap, S. Agarwal, and T. Meyarivan. A fast and elitist multiobjective genetic

- algorithm: Nsga-ii. IEEE Trans. Evol. Comput., 6(2):182–197, 2002. doi: 10.1109/4235.996017.
- R. C. Dias, J. Julià, and M. Schimmel. Rayleigh-wave, group-velocity tomography of the borborema province, ne brazil, from ambient seismic noise. Pure Appl. Geophys., 172: 1429–1449, June 2015. doi: 10.1007/s00024-014-0982-9.
- A. Dziewonski and D. Anderson. Preliminary reference earth model. Phys. Earth Planet. Inter., 25(1):297–356, 1981.
- A. Dziewonski, S. Bloch, and M. Landisman. A technique for analysis of transient seismic signals. Bull. Seismol. Soc., 59(1):427–444, 1969.
- B. Efron and R. Tibshirani. Statistical data analysis in the computer age. Science, 253 (5018):390–395, 1991. ISSN 0036-8075, 1095-9203. doi: 10.1126/science.253.5018.390.
- M. E. Everett and A. Shultz. Two-dimensional nonlinear magnetotelluric inversion using a genetic algorithm. J. Geomag. Geoelectr., 45:1013–1026, 1993.
- M. Feng, M. Assumpção, and S. Van der Lee. Group-velocity tomography and lithospheric s-velocity structure of the south american continent. Physics of the Earth and Planetary Interiors, 147(4):315–331, 2004. ISSN 0031-9201. doi: 10.1016/j.pepi.2004.07.008.
- M. Feng, S. van der Lee, and M. Assumpção. Upper mantle structure of south america from joint inversion of waveforms and fundamental mode group velocities of rayleigh waves. Journal of Geophysical Research, 112:B04312, 2007. ISSN 2156-2202. doi: 10.1029/2006JB004449.
- A. Frassetto, G. Zandt, H. Gilbert, T. J. Owens, and C. H. Jones. Improved imaging with phase-weighted common conversion point stacks of receiver functions. Geophysical

- Journal International, 182(1):368–374, 2010. ISSN 1365-246X. doi: 10.1111/j.1365-246X.2010.04617.x.
- L. Gallardo, M. Meju, and M. Pérez-Flores. A quadratic programming approach for joint image reconstruction: mathematical and geophysical examples. Inverse Problems., 21: 435–452, 2005.
- L. A. Gallardo and M. A. Meju. Characterization of heterogeneous near-surface materials by joint 2d inversion of dc resistivity and seismic data. Geophys. Res. Lett., 30(13): 1658, 2003. doi: 10.1029/2003GL017370.
- C. E. Ganade de Araujo, R. Weinberg, and U. Cordani. Extruding the borborema province (ne-brazil): a two-stage neoproterozoic collision process. Terra Nova, pages 1–12, 2013. doi: 10.1111/ter.12084.
- X. Garcia, J. Julià, A. M. Nemocón, and M. Neukirchd. Lithospheric thinning under the araripe basin (ne brazil) from a long-period magnetotelluric survey: Constraints for tectonic inversion. Gondwana Research, 68:174–184, April 2019. doi: 10.1016/j.gr.2018.11.013.
- A. Gatzemeier and M. M. 3D modelling of electrical anisotropy from electromagnetic array data: Hypothesis testing for different upper mantle conduction mechanisms. Physics of The Earth and Planetary Interiors., 149:225–242, April 2005. doi: 10.1016/j.pepi.2004.10.004.
- H. J. Gilbert and A. F. Sheehan. Images of crustal variations in the intermountain west. Journal of Geophysical Research, 109:B03306, 2004. ISSN 2156-2202. doi: 10.1029/2003JB002730.
- D. E. Goldberg. Genetic Algorithms in Search, Optimization and Machine Learning. Addison-Wesley, Boston, Mass., 1989.

- J. Gomes. Projeto radambrasil, geologia, geomorfologia, pedologia, vegetação e uso potencial do solo. MME, Rio de Janeiro., 24-27:27–300, 1981.
- K. Hegarty, J. Morais Neto, and G. Karner. The enigma of the araripe plateau: new constraints on its uplift and tectonic history using afta. SBG, Congresso Brasileiro de Geologia, 41, João Pessoa, Anais., page 495, 2002.
- E. Herrin and T. Goforth. Phase-matched filters: Application to the study of rayleigh waves. Bull. Seismol. Soc., 67(1):1259–1275, 1977.
- R. Herrmann and C. Ammon. Surface Waves, Receiver Functions and Crustal Structure, 2002.
- W. Hu, A. Abubakar, and T. Habashy. Joint electromagnetic and seismic inversion using structural constraints. Geophysics., 74:99–109, 2009.
- E. F. Jardim de Sá, R. I. F. Da Trindade, M. H. B. M. de Hollanda, J. M. M. Araújo, A. C. Galindo, V. E. Amaro, Z. S. de Souza, J.-L. Vignerresse, and J.-M. Lardeaux. Brasiliano syntectonic alkaline granites emplaced in a strike slip/extensional setting (Eastern Seridó Belt, NE Brazil). Anais da Academia Brasileira de Ciencias, 71(1): x2–27, 1999. ISSN 00013765.
- M. Jegen, R. Hobbs, P. Tarits, and A. Chave. Joint inversion of marine magnetotelluric and gravity data incorporating seismic constraints. preliminary results of sub-basalt imaging off the faroe shelf. Earth and Plan. Science Lett., 282:47–55, 2009.
- M. Jervis, P. L. Stoffa, and M. K. Sen. 2-d migration velocity estimation using a genetic algorithm. Geophysical Research Letters., 20:1495–1498, 1993.
- A. G. Jones. Imaging the Continental Upper Mantle Using Electromagnetic Method. Lithos., 48(1), December 1999. doi: 10.1016/S0024-4937(99)00022-5.

- A. G. Jones and I. J. Ferguson. The electric Moho. Nature., 409:331–333, February 2001. doi: 10.1038/35053053.
- A. G. Jones, R. L. Evans, and E. D. W. Velocity–conductivity relationships for mantle mineral assemblages in archean cratonic lithosphere based on a review of laboratory data and hashin–shtrikman extremal bounds. Lithos., 109:131–143, April 2009. doi: 10.1016/j.lithos.2008.10.014.
- J. Julià and J. Mejia. Thickness and vp/vs ratio variation in the iberian crust. Geophysical Journal International., 156(1):59–72, December 2003. doi: 10.1111/j.1365-246X.2004.02127.x.
- J. Julià and J. Mejia. Thickness and vp/vs ratio variation in the iberian crust. Geophysical Journal International., 156(1):59–72, 2004. doi: 10.1111/j.1365-246X.2004.02127.x.
- J. Julià, C. J. Ammon, R. B. Herrmann, and A. M. Correig. Joint inversion of receiver function and surface wave dispersion observations. Geophysical Journal International, 143(1):99–112, 2000. ISSN 1365-246X. doi: 10.1046/j.1365-246x.2000.00217.x.
- J. Julià, C. J. Ammon, and R. B. Herrmann. Lithospheric structure of the arabian shield from the joint inversion of receiver functions and surface-wave group velocities. Tectonophysics, 371(1):1–21, 2003. ISSN 0040-1951. doi: 10.1016/S0040-1951(03)00196-3.
- R. Kind and X. Li. Deep Earth Structure - Transition Zone and Mantle Discontinuities. Elsevier, GFZ Potsdam, Potsdam, Germany, 2007. doi: 10.1016/B978-0-444-53802-4.00017-8.
- K. M. Knesel, Z. S. Souza, P. M. Vasconcelos, B. E. Cohen, , and F. V. Silveira. Young volcanism in the borborema province, ne brazil, shows no evidence for a trace of the fernando de noronha plume on the continent. Earth Planet. Sci. Lett., 302:38–50, 2011.

- I. Lahti, T. Korja, P. Kaikkonen, and K. Vaittinen. Decomposition analysis of the bear magnetotelluric data: implications for the upper mantle conductivity in the fennoscandian shield. Geophys. Journal International., 163:900–914, 2005. doi: 10.1111/j.1365-246X.1995.tb05915.x.
- C. A. Langston. Structure under mount rainier, washington, inferred from teleseismic body waves. Journal of Geophysical Research, 84:4749–4762, 1979. ISSN 2156-2202. doi: 10.1029/JB084iB09p04749.
- V. Levin and J. Park. P-sh conversions in a flat-layered medium with anisotropy of arbitrary orientation. Geophys. J. Int., 131:253–266, 1997. doi: 10.1111/j.1365-246X.1997.tb01220.x.
- J. P. Ligorria and C. J. Ammon. Iterative deconvolution and receiver-function estimation. Bulletin of the Seismological Society of America, 89(5):1395–1400, 1999. ISSN 0037-1106, 1943-3573.
- N. Linde, A. Binley, A. Tryggvason, L. B. Pedersen, and A. Revil. Improved hydrogeophysical characterization using joint inversion of cross-hole electrical resistance and ground-penetrating radar traveltime data. Water Resour. Res., 42, 2006. doi: 10.1029/2006WR005131.
- R. M. N. Luz, J. Julià, and A. F. do Nascimento. Bulk crustal properties of the borborema province, NE brazil, from p-wave receiver functions: Implications for models of intraplate cenozoic uplift. Tectonophysics, 644–645:81–91, 2015. ISSN 0040-1951. doi: 10.1016/j.tecto.2014.12.017.
- F. O. Marques, F. C. C. Nogueira, F. H. R. Bezerra, and D. L. De Castro. The Araripe Basin in NE Brazil: An intracontinental graben inverted to a high-standing horst. Tectonophysics, 630:251–264, 2014. doi: 10.1016/j.tecto.2014.05.029.

- G. Marquis, A. G. Jones, and R. D. Hyndman. Coincident conductive and reflective middle and lower crust in southern british columbia. Geophys. Journal International., 120:111–131, 1995. doi: 10.1111/j.1365-246X.1995.tb05915.x.
- R. M. D. d. Matos. Magmatism and hotspot trails during and after continental break-up in the south atlantic. Marine and Petroleum Geology, 129:105077, Jul 2021. ISSN 0264-8172. doi: 10/gk49s6.
- G. A. Mcmechan and M. J. Yedlin. Analysis of dispersive waves by wave field transformation. Geophysics, 46, 1981. doi: 10.1190/1.1441225.
- A. M. P. Mizusaki, A. Thomaz-Filho, E. J. Milani, and P. Césero. Mesozoic and cenozoic igneous activity and its tectonic control in northeastern brazil. J. South Am. Earth Sci., 15:183–198, 2002.
- M. Moorkamp. Joint inversion of magnetotelluric and receiver function data. PhD thesis, University of Ireland, 2007.
- M. Moorkamp, A. G. Jones, and D. W. Eaton. Joint inversion of teleseismic receiver functions and magnetotelluric data using a genetic algorithm: Are seismic velocities and electrical conductivities compatible?. Geophysical Research Letters., 34, August 2007. doi: 10.1029/2007gl030519.
- M. Moorkamp, A. Jones, and S. Fishwik. Joint inversion of receiver functions, surface wave dispersion and magnetotelluric data. Journal of Geophysical Research., 115, April 2010. doi: 10.1029/2009JB006369.
- M. Moorkamp, B. Heincke, M. Jegen, A. Roberts, and R. Hobbs. A framework for 3-d joint inversion of mt, gravity and seismic refraction data. Geophys. Journal International., 184:477–493, 2011. doi: 10.1111/j.1365-246X.2010.04856.x.

- J. Morais Neto, K. Hegarty, G. Karner, and F. Alkmim. Timing and mechanisms for the generation and modification of the anomalous topography of the Borborema Province, northeastern Brazil. Marine and Petroleum Geology, 26(7):1070–1086, aug 2009. ISSN 02648172. doi: 10.1016/j.marpetgeo.2008.07.002.
- M. Neukirch and X. Garcia. Nonstationary magnetotelluric data processing with instantaneous parameter. Journal of Geophysical Research: Solid Earth, 119(3):1634–1654, mar 2014. ISSN 21699313. doi: 10.1002/2013JB010494.
- R. G. Oliveira and W. E. Medeiros. Deep crustal framework of the borborema province, ne brazil, derived from gravity and magnetic data. Precambrian Research, 315:45–65, 2018.
- R. G. d. Oliveira and W. E. de Medeiros. Evidences of buried loads in the base of the crust of Borborema Plateau (NE Brazil) from Bouguer admittance estimates. Journal of South American Earth Sciences, 37:60–76, aug 2012. ISSN 08959811. doi: 10.1016/j.jsames.2012.02.004.
- T. J. Owens, G. Zandt, and S. R. Taylor. Seismic evidence for an ancient rift beneath the cumberland plateau, tennessee: A detailed analysis of broadband teleseismic p wave-forms. Journal of Geophysical Research, 89:7783–7795, 1984. ISSN 2156-2202. doi: 10.1029/JB089iB09p07783.
- J.-P. Peulvast and F. Bétard. A history of basin inversion, scarp retreat and shallow denudation: The araripe basin as a keystone for understanding long-term landscape evolution in ne brazil. Geomorphology, 233:20–40, 2015. ISSN 0169-555X. doi: 10.1016/j.geomorph.2014.10.009.
- F. Ponte, F.C.and Ponte Filho. Estrutura geológica e evolução tectônica da bacia do araripe. Recife: DNPM., 17(2–3):68, November 1996.

M. A. Pérez-Flores and A. Schultz. Application of 2-d inversion with genetic algorithms to magnetotelluric data from geothermal areas. Earth, Planets, and Space., 54:607–616, 2002.

N. Rawlinson and M. Sambridge. The fast marching method: an effective tool for tomographic imaging and tracking multiple phases in complex layered media. Exploration Geophysics, 36:341–350, 2005.

M. Schimmel and H. Paulssen. Noise reduction and detection of weak, coherent signals through phase-weighted stacks. Geophysical Journal International, 130(2):497–505, 1997. ISSN 1365-246X. doi: 10.1111/j.1365-246X.1997.tb05664.x.

F. Simões Neto, J. Julià, and M. Schimmel. Upper-mantle structure of the borborema province, ne brazil, from p-wave tomography: implications for rheology and volcanism. Geophysical Journal International, 216:231–250, 2019. doi: 10.1093/gji/ggy421.

B. Tournerie and M. Chouteau. Three-dimensional magnetotelluric survey to image structure and stratigraphy of a sedimentary basin in hungary. Physics of The Earth and Planetary Interiors., 150:197–212, 2005. doi: 10.1016/j.pepi.2004.08.028.

N. Ussami, E. Molina, and W. Medeiros. Novos Vínculos sobre a Evolução Térmica da Margem Continental Leste do Brasil. In VII Simpósio Nacional de Estudos Tectônicos, Lençóis – BA. Resumos Expandidos, Sociedade Brasileira de Geologia, number 7, pages 20 – 23, Salvador – BA, 1999.

L. Valença, V. Neumann, and J. Mabeoone. An overview on callovian-cenomanian intracratonic basins of northeast brazil: onshore stratigraphic record of the opening of the southern atlantic. Geologica Acta, RCUB., pages 261–275, 2003. ISSN 1695-6133. doi: 0.1344/105.000001614.

M. S. Zhdanov. Geophysical electromagnetic theory and methods. ELSEIVER, Salt Lake City, UT 84112, U.S.A., 2009.

- L. Zhu and H. Kanamori. Moho depth variation in southern california from teleseismic receiver functions. Journal of Geophysical Research, 105:2969–2980, 2000. ISSN 2156-2202. doi: 10.1029/1999JB900322.

Appendix A

Supplementary information chapter 4

This appendix presents the supplementary material of the article entitled: *Lithospheric structure of the western Borborema Province from receiver functions and surface-wave dispersion: Implications for basin inversion*. This supporting information that complement the text and figures provided in the manuscript.

Supplementary material

Lithospheric structure of the western Borborema Province from receiver functions and surface-wave dispersion: Implications for basin inversion.

Ana M. Nemocón et al.

April 20, 2021

1 Receiver function stacks

Receiver function stacks for the BODES Experiment stations (model shown in Figure 2). Each panel displays two traces corresponding to the radial (top) and transverse (bottom) receiver functions, respectively. The stacks are presented as a black line within a grey shade representing 1-confidence bounds for the stacked amplitudes. The number of waveforms, average back-azimuth, and average ray parameter (in s/km) is indicated on top of each panel.

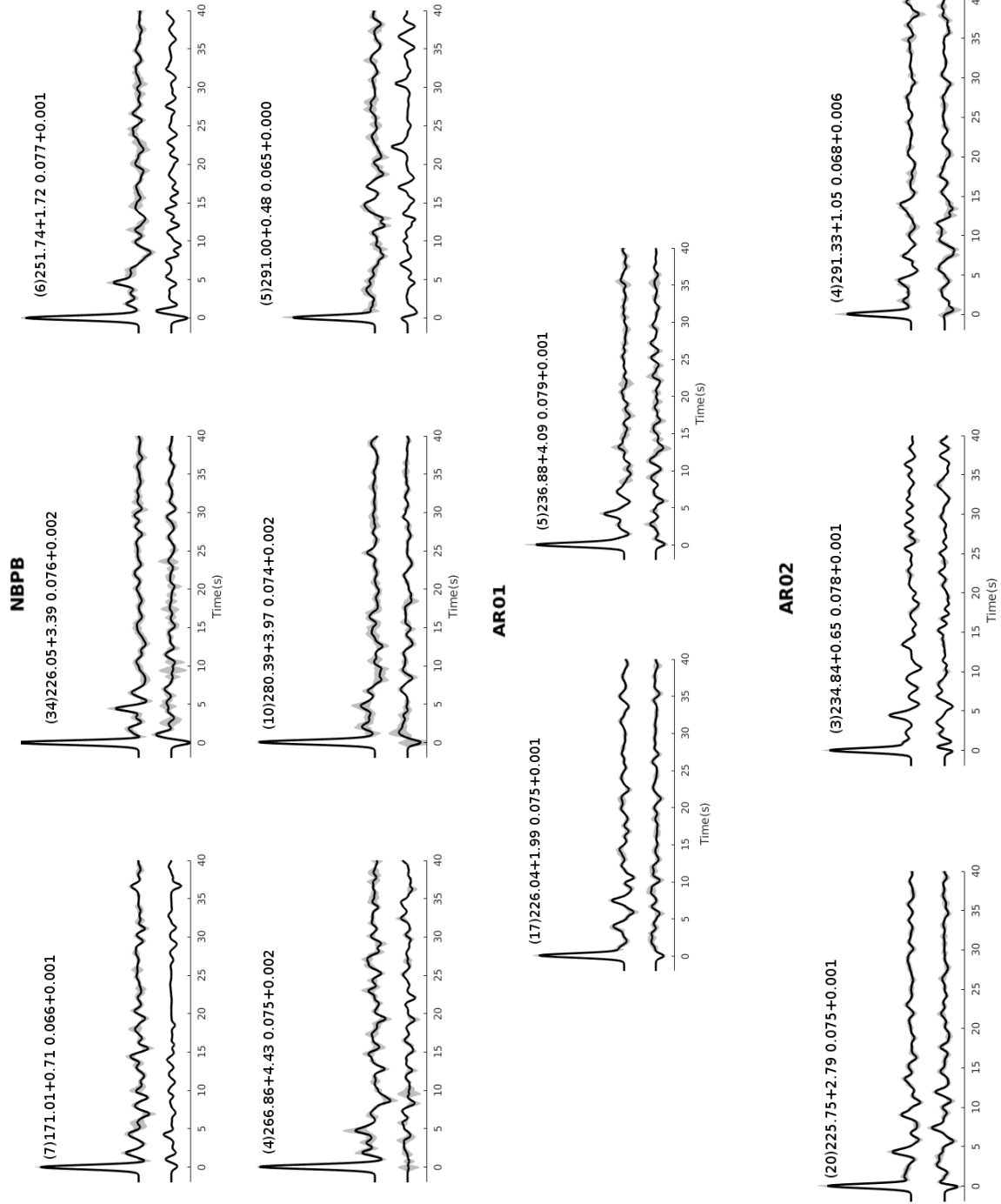
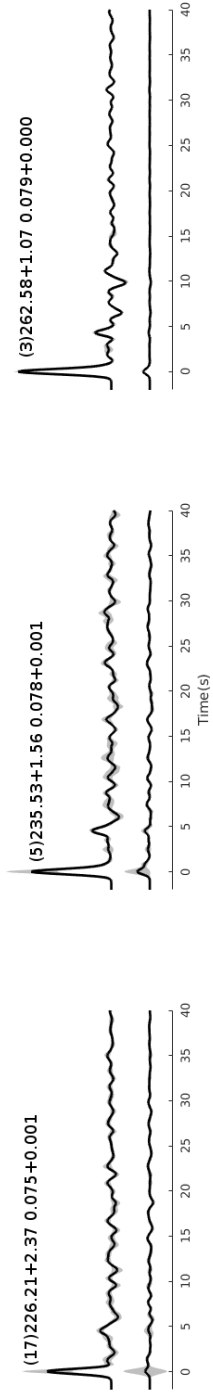
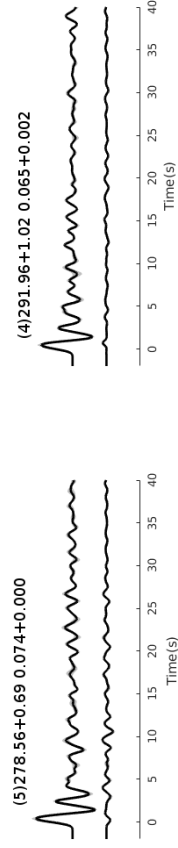
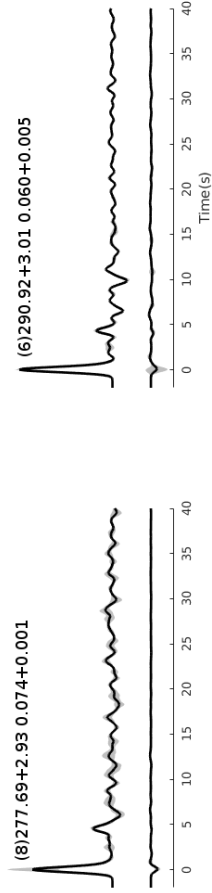


Figure S1. Sample receiver function stacks. Each panel displays two traces corresponding to the radial (top) and transverse (bottom) receiver functions, respectively. The stacks are presented as a grey line within a grey shade representing 1 σ confidence bounds for the stacked amplitudes. The number of waveforms, average back azimuth, and average ray parameter (in s/km) is indicated on top of each panel.

AR03



AR50



AR51

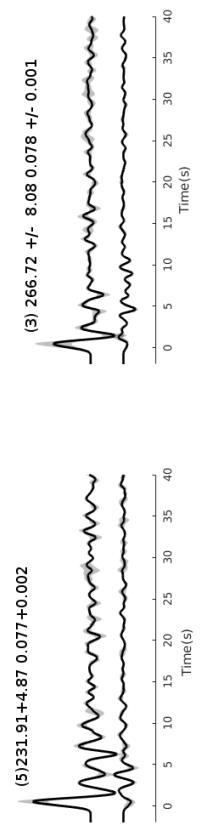


Figure S2. Sample receiver function stacks. Same legend as in Figure S1.

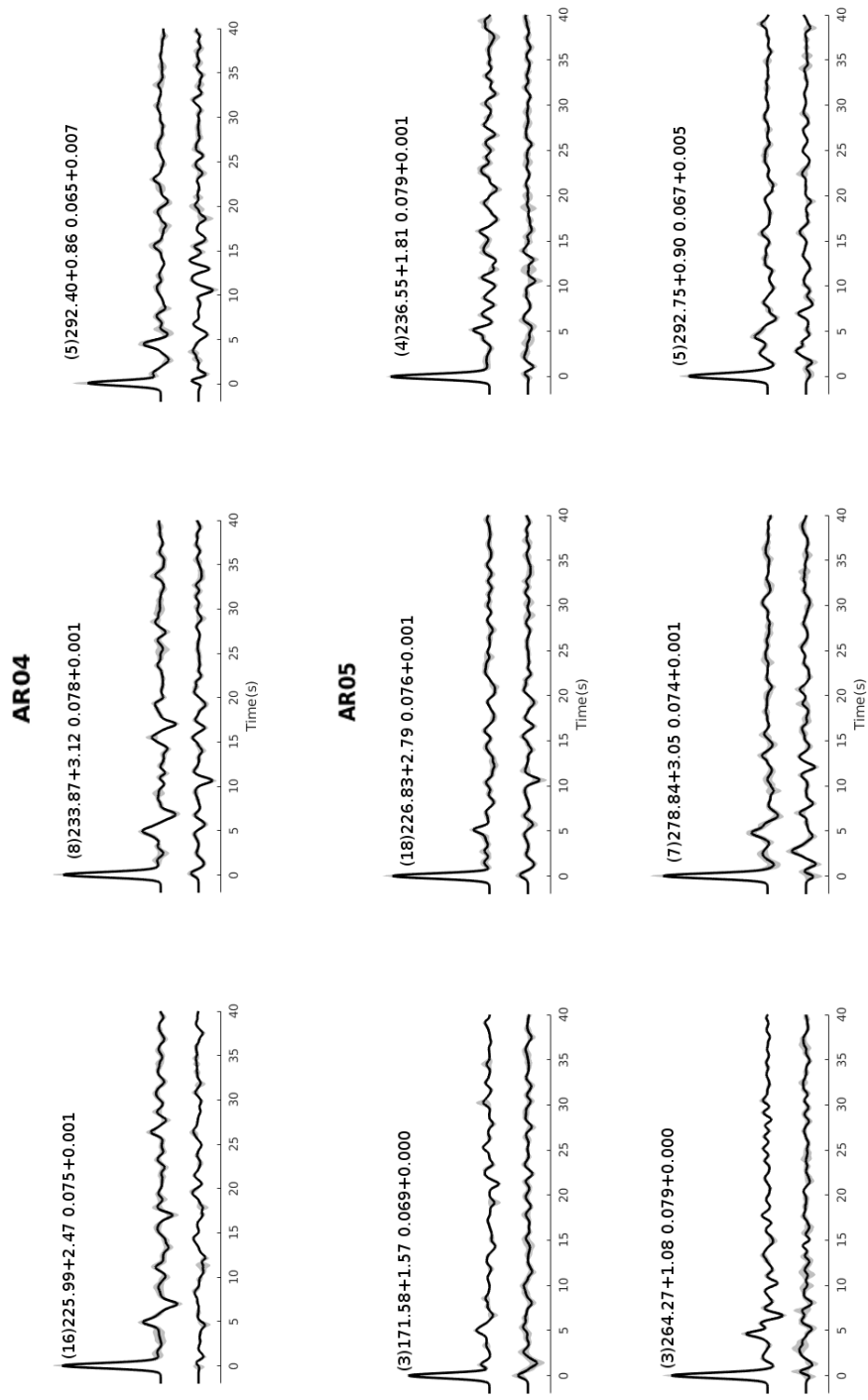


Figure S3. Sample receiver function stacks. Same legend as in Figure S1.

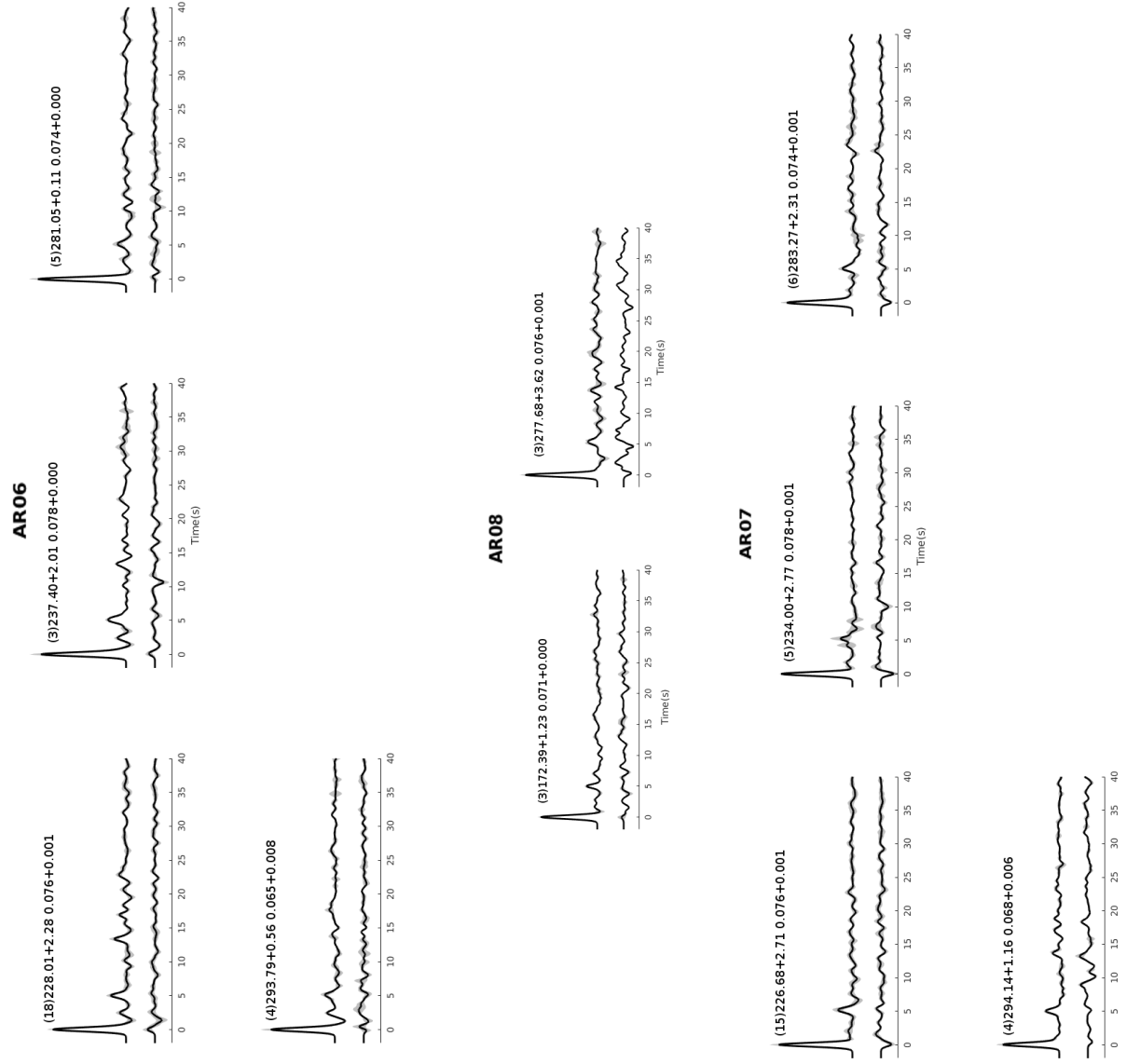


Figure S4. Sample receiver function stacks. Same legend as in Figure S1.

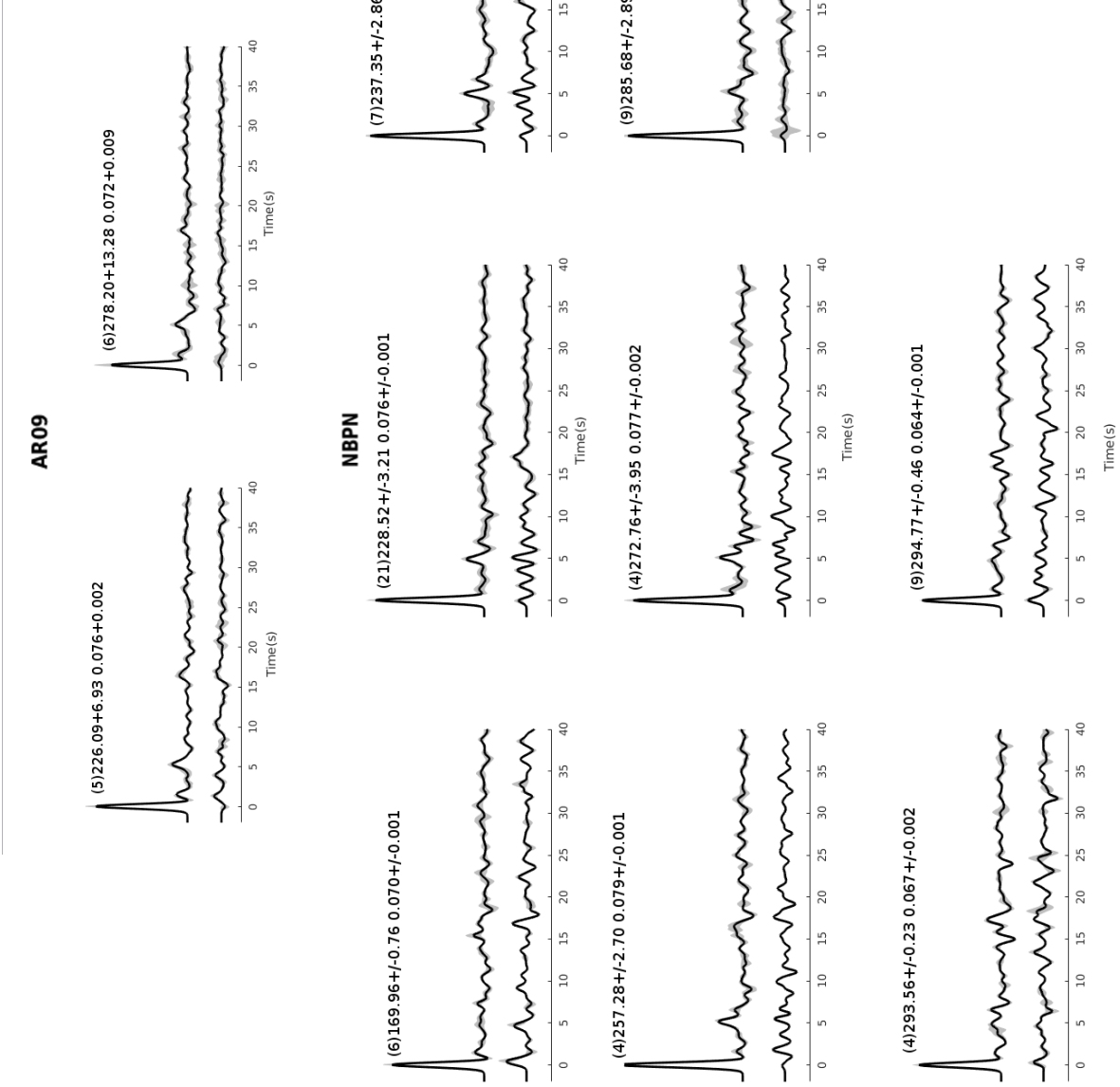


Figure S5. Sample receiver function stacks. Same legend as in Figure S1.

2 H- κ stacking

H- κ stacking results for the BODES Experiment stations (model shown in Figure 3).

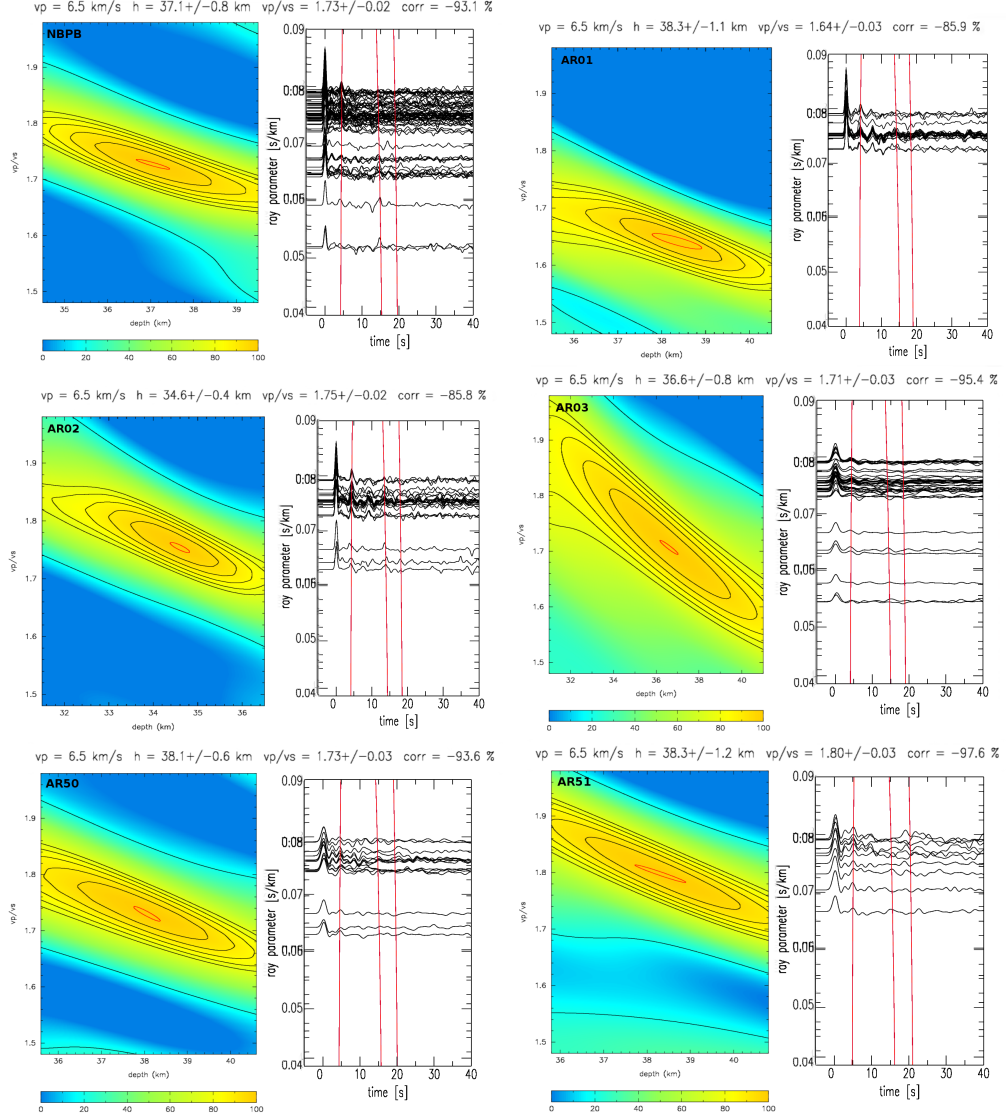


Figure S6. Bodes stations N S, which are identified in the upper left. The right panels show the corresponding radial receiver functions used during the stacking procedure, sorted by ray parameter, while the left panels show the H κ stacking surfaces and the identification of each station. The thin red lines in the right panels are the phase moveout curves for the Ps, PpPs and PpSs + PsPs phases associated with the maximum in the H κ stacking surface. The thick red line in the left panels is the 1 σ confidence ellipse from bootstrapping.

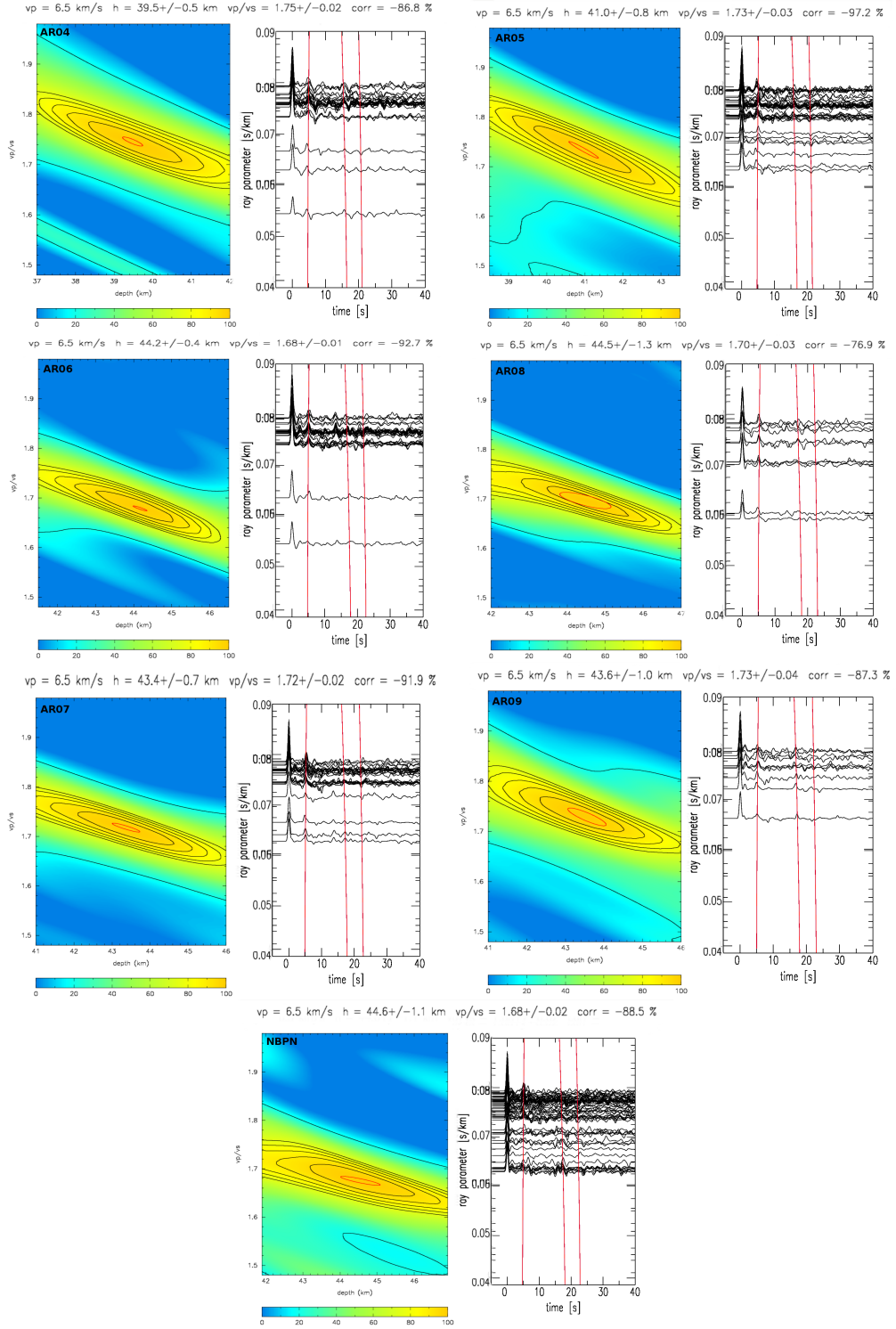


Figure S7. Same legend as in Figure S6.

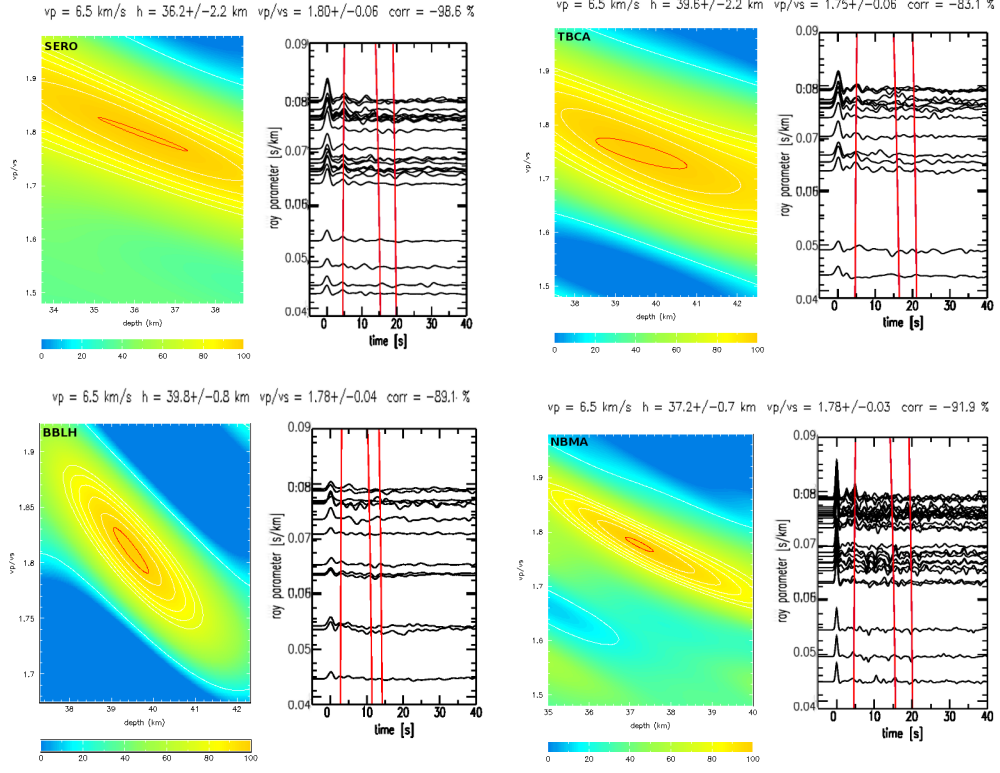


Figure S8. Same legend as in Figure S6.

2.1 H- κ stacking varying the initial parameters

H- κ stacking results for the BODES Experiment stations varying V_p 6.3 to 6.8 range. We found variations of ~ 0.02 for V_p/V_s ratio and $\sim 4 \text{ km}$ for H. Also, we realized H- κ stacking varying weights of the initial parameters.

Table S1.

STATION	Vp (km/s)	Vp/Vs	eps Vp/Vs	H	epsilon H	Corr %
NBPB	6,30	1,743	0,010	35,380	0,374	-90,53
	6,40	1,741	0,012	35,960	0,426	-87,07
	6,50	1,73	0,02	37,1	0,8	93,1
	6,60	1,731	0,010	37,370	0,429	-88,93
	6,70	1,726	0,011	38,100	0,437	-84,56
	6,80	1,722	0,012	38,790	0,520	-87,27
AR01	6,30	1,653	0,051	36,570	2,678	-94,30
	6,40	1,648	0,051	37,310	2,724	-95,49
	6,50	1,637	0,060	38,290	3,250	-95,54
	6,60	1,632	0,053	39,070	3,080	-95,55
	6,70	1,624	0,052	40,010	3,078	-96,55
	6,80	1,618	0,058	40,870	3,563	-95,33
AR02	6,30	1,765	0,019	33,200	0,540	-85,97
	6,40	1,759	0,018	33,890	0,523	-86,21
	6,50	1,75	0,02	34,6	0,4	85,8
	6,60	1,749	0,019	35,260	0,611	-88,76
	6,70	1,744	0,017	35,940	0,566	-88,50
	6,80	1,738	0,017	36,660	0,570	-87,88
AR03	6,30	1,722	0,037	35,010	1,195	-97,36
	6,40	1,718	0,033	35,680	1,070	-96,41
	6,50	1,71	0,03	36,6	0,8	95,4
	6,60	1,710	0,039	37,060	1,338	-97,54
	6,70	1,706	0,042	37,770	1,507	-97,70
	6,80	1,701	0,040	38,510	1,470	-98,09
AR50	6,30	1,754	0,022	36,580	0,516	-81,32
	6,40	1,748	0,024	37,350	0,512	-83,79
	6,50	1,73	0,03	38,1	0,6	93,6
	6,60	1,737	0,021	38,880	0,504	-76,69
	6,70	1,731	0,021	39,670	0,545	-74,85
	6,80	1,726	0,022	40,420	0,512	-76,34
AR51	6,30	1,817	0,027	36,450	1,082	-96,91
	6,40	1,812	0,025	37,180	1,022	-97,39
	6,50	1,80	0,03	38,3	1,2	97,6
	6,60	1,799	0,027	38,730	1,132	-98,44
	6,70	1,793	0,027	39,510	1,192	-97,96
	6,80	1,787	0,028	40,300	1,261	-98,39
AR04	6,30	1,743	0,020	38,100	0,440	-84,05
	6,40	1,739	0,020	38,830	0,452	-82,92
	6,50	1,75	0,02	39,5	0,5	86,8
	6,60	1,728	0,020	40,430	0,500	-87,89
	6,70	1,723	0,019	41,230	0,473	-86,50
	6,80	1,717	0,019	42,090	0,453	-84,84
AR05	6,30	1,746	0,034	39,300	0,781	-98,00
	6,40	1,741	0,035	40,100	0,807	-97,67
	6,50	1,73	0,03	41,0	0,8	97,2
	6,60	1,731	0,040	41,680	1,001	-98,15
	6,70	1,725	0,035	42,530	0,899	-97,89
	6,80	1,719	0,034	43,400	0,900	-97,17
AR06	6,30	1,695	0,013	42,080	0,516	-93,27
	6,40	1,690	0,014	42,950	0,573	-95,16
	6,50	1,68	0,01	44,2	0,4	92,7
	6,60	1,679	0,015	44,730	0,670	-96,09
	6,70	1,673	0,017	45,670	0,734	-95,83
	6,80	1,667	0,017	46,620	0,780	-96,70
AR08	6,30	1,691	0,032	42,840	1,547	-95,62
	6,40	1,689	0,032	43,560	1,537	-94,70
	6,50	1,70	0,03	44,5	1,3	76,9
	6,60	1,681	0,032	45,270	1,634	-95,53
	6,70	1,676	0,035	46,150	1,847	-96,03
	6,80	1,671	0,032	47,070	1,717	-95,37
AR07	6,30	1,722	0,018	41,990	0,592	-95,85
	6,40	1,717	0,016	42,820	0,585	-92,94
	6,50	1,72	0,02	43,4	0,7	91,9
	6,60	1,707	0,013	44,520	0,488	-94,99
	6,70	1,702	0,014	45,380	0,499	-94,72
	6,80	1,696	0,013	46,270	0,470	-93,66
AR09	6,30	1,735	0,031	41,880	0,746	-86,15
	6,40	1,730	0,027	42,750	0,706	-81,78
	6,50	1,73	0,04	43,6	1,0	87,3
	6,60	1,721	0,026	44,500	0,728	-79,31
	6,70	1,714	0,026	45,380	0,725	-83,73
	6,80	1,710	0,028	46,200	0,820	-83,65

*Number of RFs, Gaussian widths and weights as in Table 1.

Table S1.

NBMA	6,30	1,782	0,033	35,880	0,755	-95,37
	6,40	1,778	0,035	36,569	0,833	-96,01
	6,50	1,78	0,03	37,2	0,7	-91,9
	6,60	1,769	0,033	37,959	0,847	-94,39
	6,70	1,764	0,028	38,729	0,762	-93,58
	6,80	1,756	0,027	39,540	0,710	-92,14
SERO	6,30	1,810	0,070	34,800	2,300	-99,00
	6,40	1,800	0,070	35,500	2,300	-98,80
	6,50	1,80	0,06	36,2	2,2	-98,6
	6,60	1,790	0,060	36,900	2,200	-98,60
	6,70	1,780	0,060	37,700	2,200	-98,60
	6,80	1,780	0,050	38,500	2,100	-98,30
TBCA	6,30	1,750	0,070	38,200	2,300	-89,00
	6,40	1,750	0,060	38,900	2,300	-87,50
	6,50	1,75	0,06	39,6	2,2	-83,1
	6,60	1,740	0,060	40,400	2,300	-84,90
	6,70	1,740	0,060	41,200	2,100	-85,40
	6,80	1,730	0,060	42,100	2,200	-86,00
BBLH	6,30	1,82	0,04	38	0,8	-93,3
	6,40	1,81	0,04	38,8	0,8	-93,4
	6,50	1,78	0,04	39,8	0,8	-89,1
	6,60	1,78	0,04	40,2	0,8	-95,2
	6,70	1,78	0,04	40,9	0,8	-94
	6,80	1,77	0,04	41,8	0,9	-94,5

*Number of RFs, Gaussian widths and weights as in Table 1.

3 L-curves

To determine values for the influence (p) and smoothness (θ) parameters, we initially fixed the influence factor to $p = 0.5$ (equal importance to each data set) and varied the smoothness factor between $0.1 \leq \theta \leq 1.9$. The best value was then chosen through L-curve analysis of the trade-off between RMS misfit and model roughness. Posteriorly, we fixed the smoothness factor and varied the influence factor between $0.1 \leq p \leq 0.9$. In general, influence factors chosen to be around 0.5, while smoothness factors were in the 0.7-0.8 range.

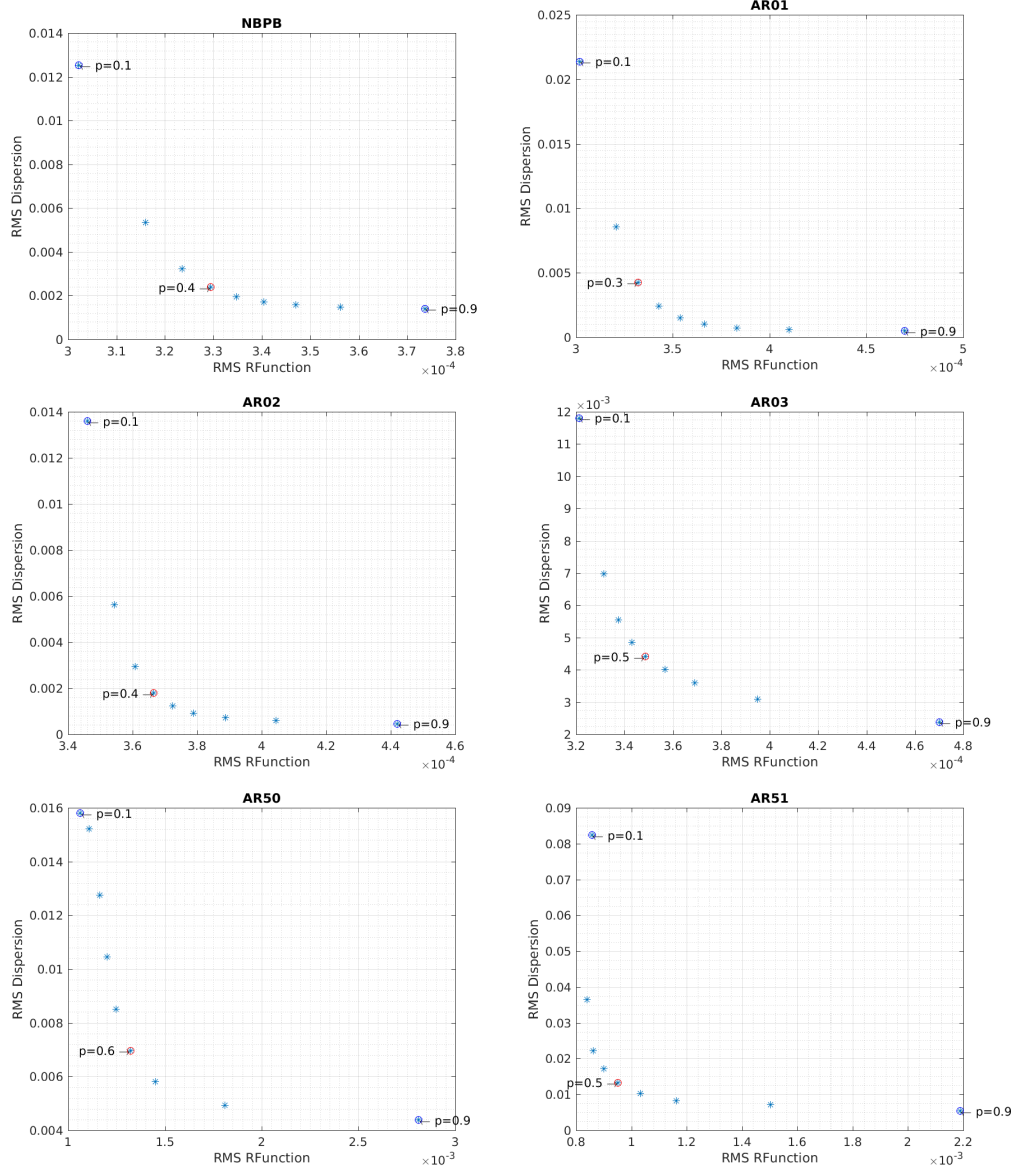


Figure S9. Details L curves, influence (p) parameter calculated for the BODES Experiment stations. Each panel display the trade off curve between RMS receiver function misfit versus RMS dispersion velocity misfit, with the optimal value in the trade off curves indicated.

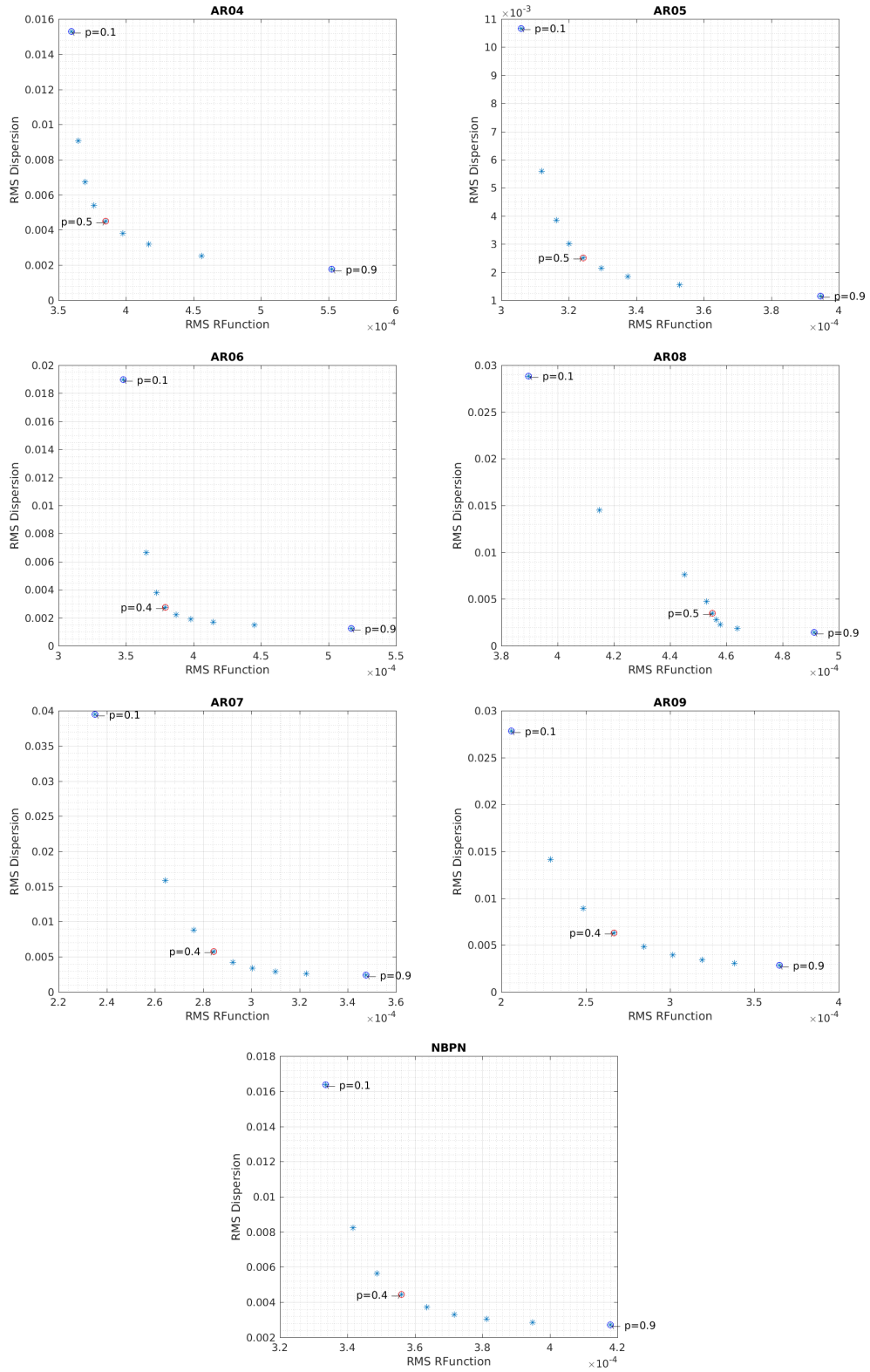


Figure S10. Details L curves, influence (p) parameter calculated for the BODES Experiment stations. Each panel display the trade off curve between RMS receiver function misfit versus RMS dispersion velocity misfit, with the optimal value in the trade off curves indicated.

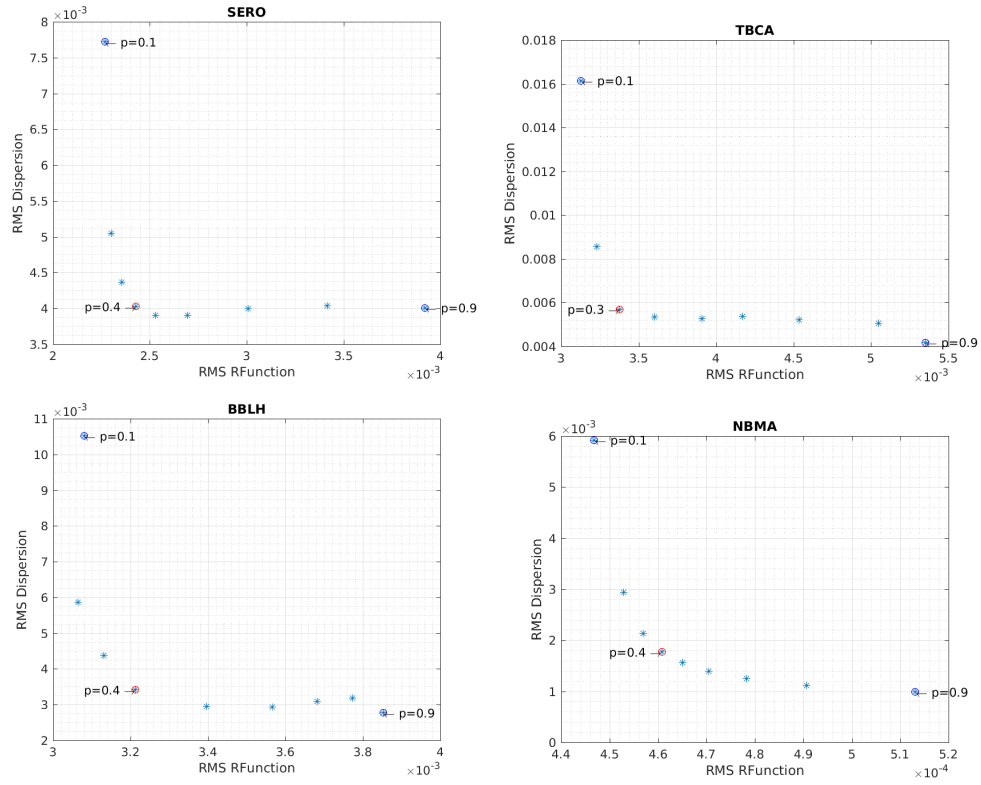


Figure S11. Details L curves, influence (p) parameter calculated for the BODES Experiment stations. Each panel display the trade off curve between RMS receiver function misfit versus RMS dispersion velocity misfit, with the optimal value in the trade off curves indicated.

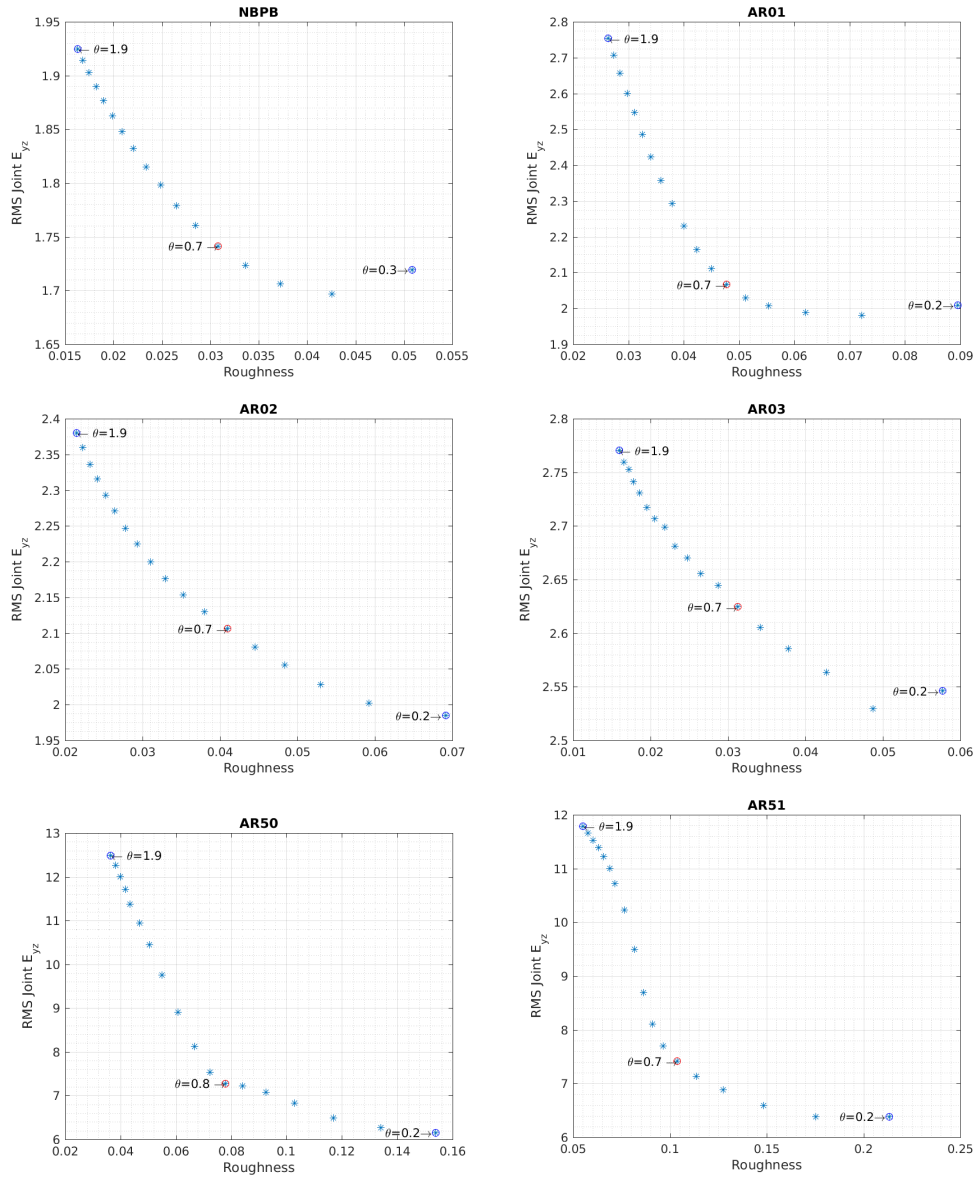


Figure S12. Details L curves, smoothness (θ) parameter calculated for the BODES Experiment stations. Each panel display the trade off curve between Roughness vs RMS misfit, with the optimal value in the trade off curves indicated.

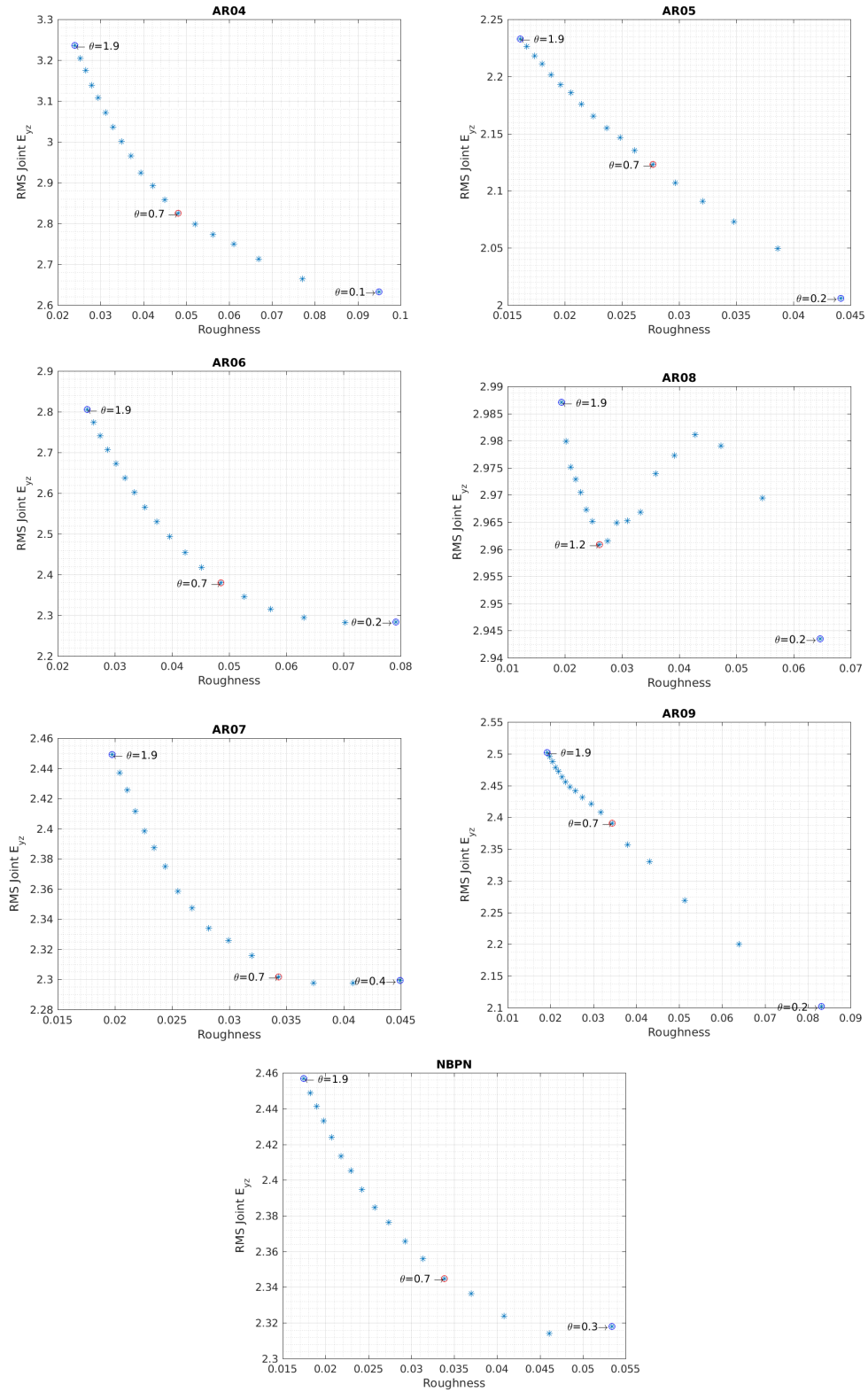


Figure S13. Details L curves, smoothness (θ) parameter calculated for the BODES Experiment stations. Each panel display the trade off curve between Roughness vs RMS misfit, with the optimal value in the trade off curves indicated.

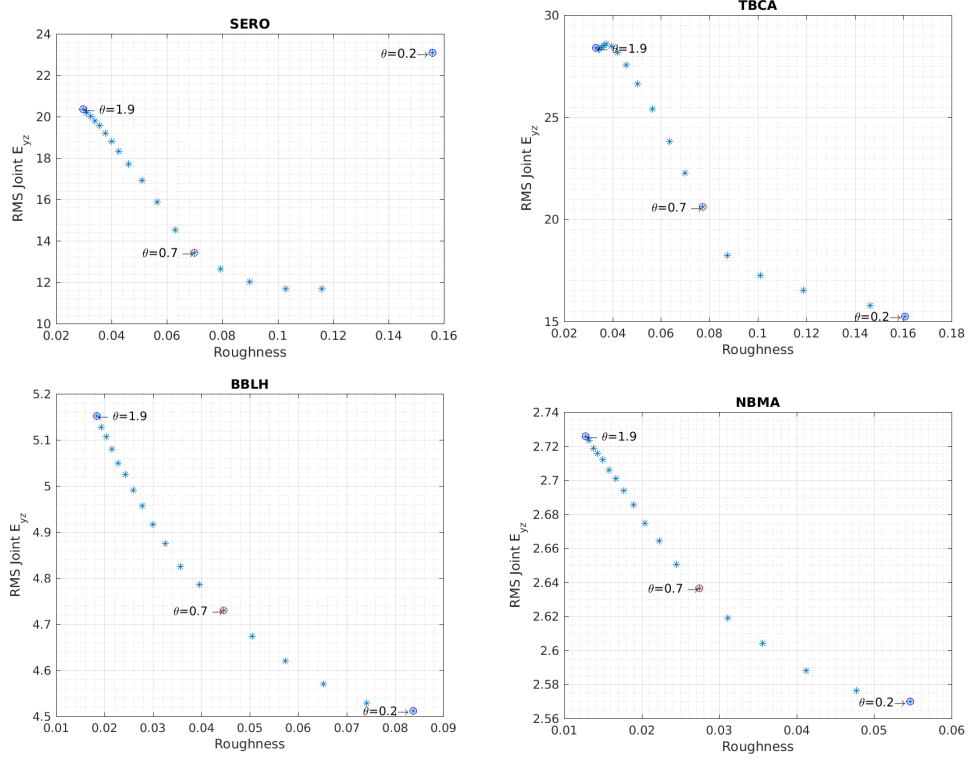


Figure S14. Details L curves, smoothness (θ) parameter calculated for the BODES Experiment stations. Each panel display the trade off curve between Roughness vs RMS misfit, with the optimal value in the trade off curves indicated.

4 Joint inversion results

Detailed joint inversion results for the BODES Experiment stations (model shown in Figure 6).

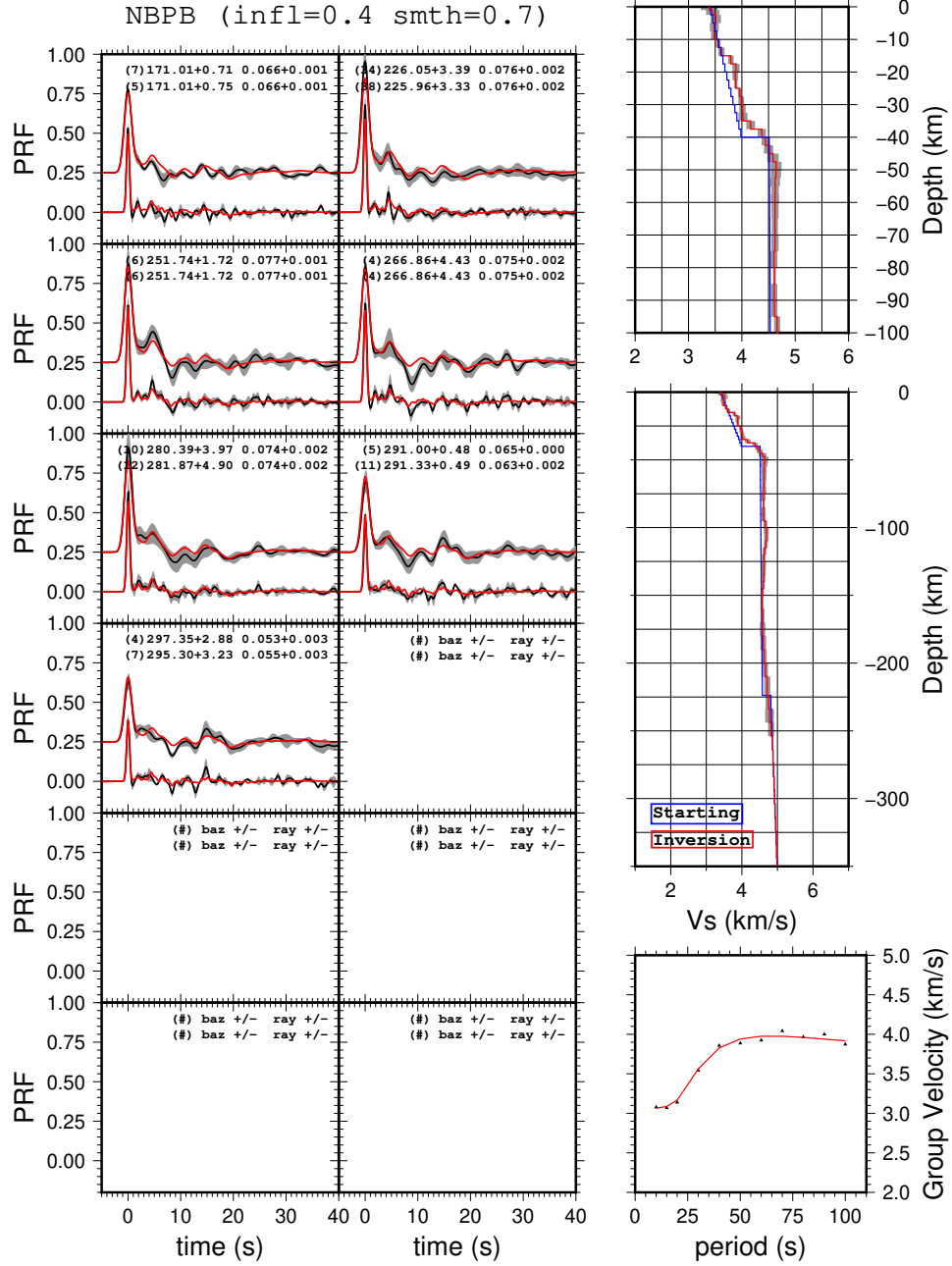


Figure S15. Detailed joint inversion results for station NBPB. Observed (black) and predicted (red) receiver functions at high and low frequency contents are displayed on the left hand panels. Observed (red) and predicted (triangles) dispersion velocities are displayed in the upper right panel. The inverted (red) and starting (blue) velocity models are displayed in the lower right corner, along with the 2σ confidence bounds (grey shade).

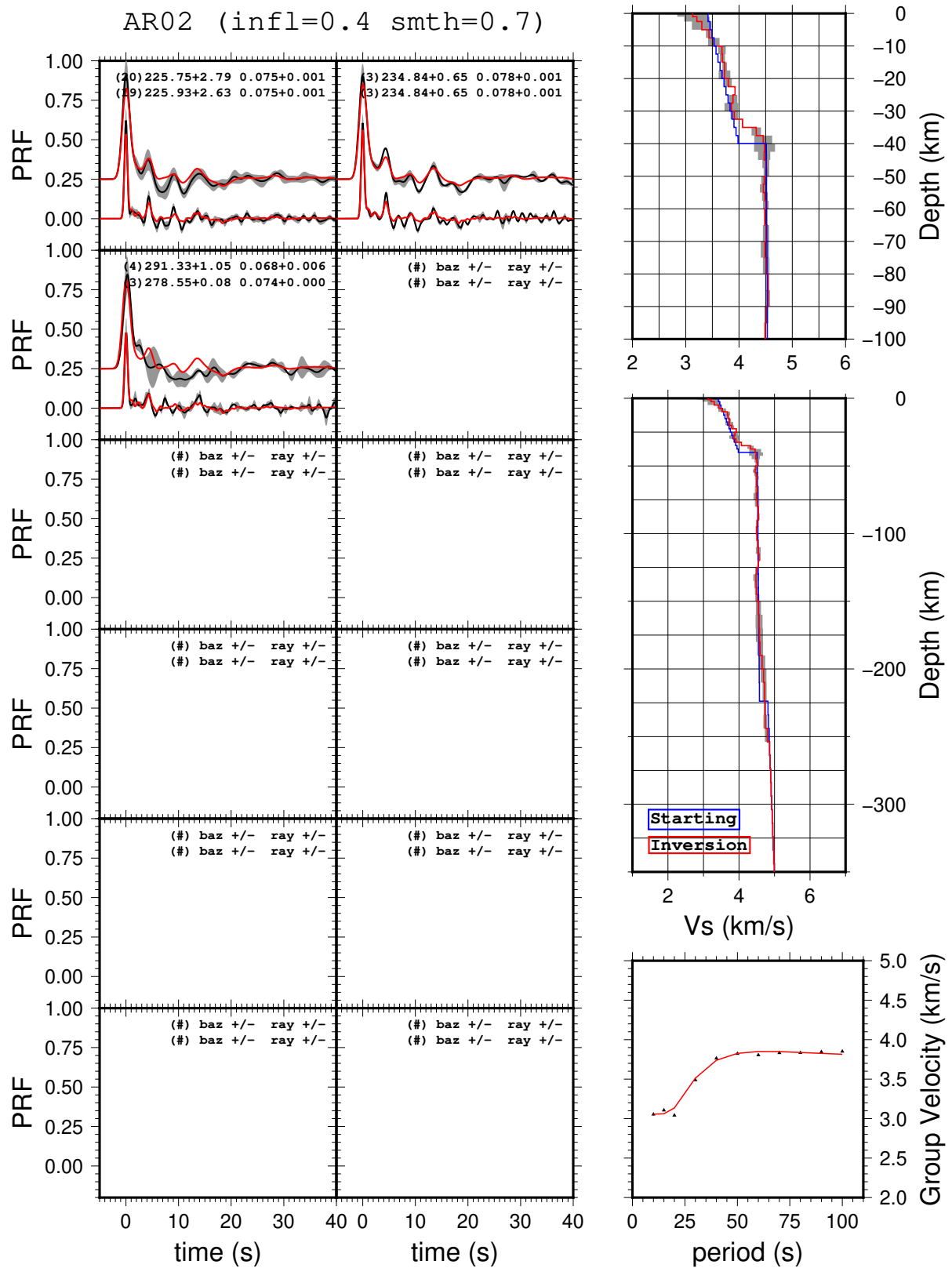


Figure S17. Detailed joint inversion results for station AR02. Same legend as in Figure S15.

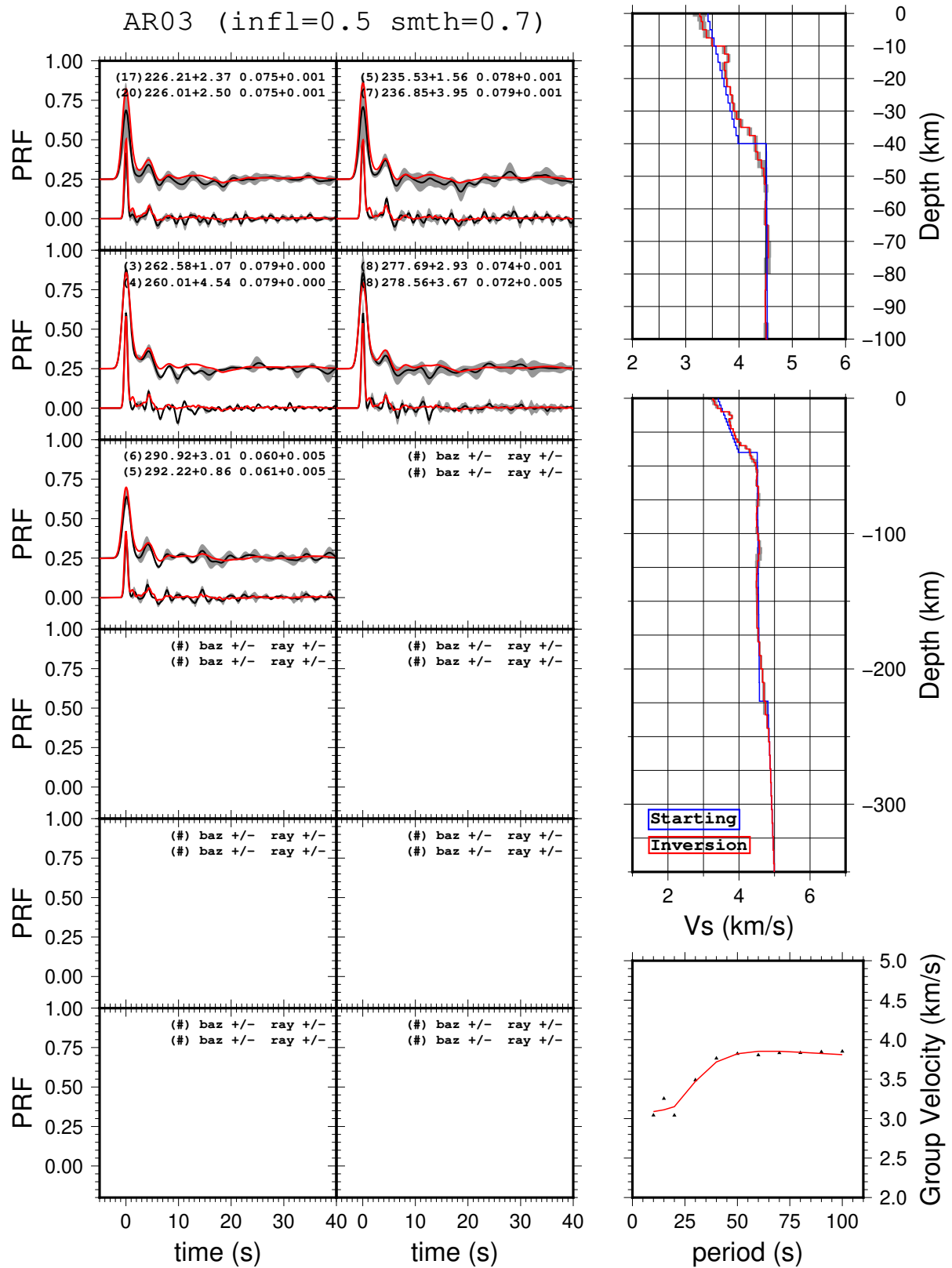


Figure S18. Detailed joint inversion results for station AR03. Same legend as in Figure S15.

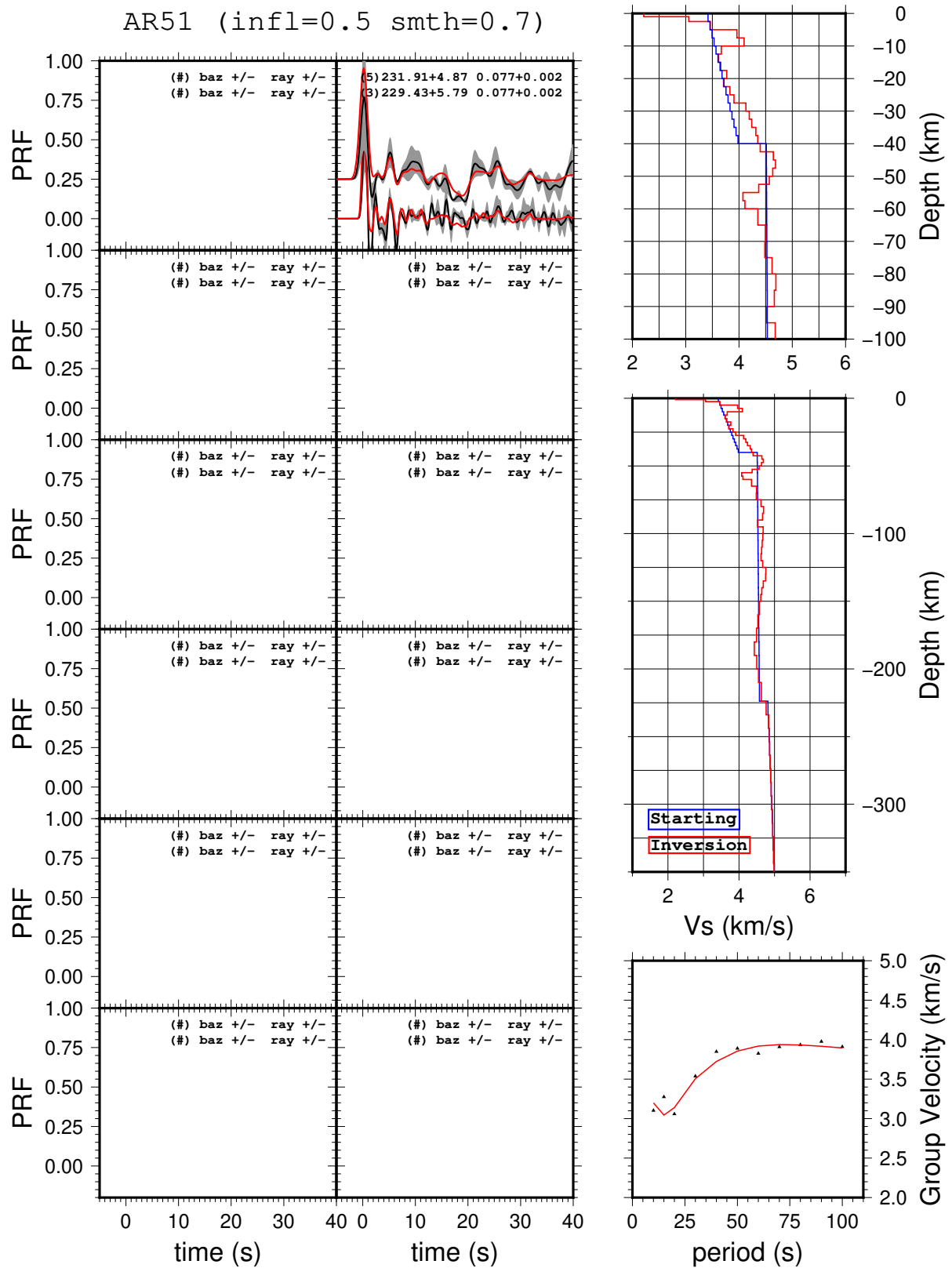


Figure S20. Detailed joint inversion results for station AR51. Same legend as in Figure S15.

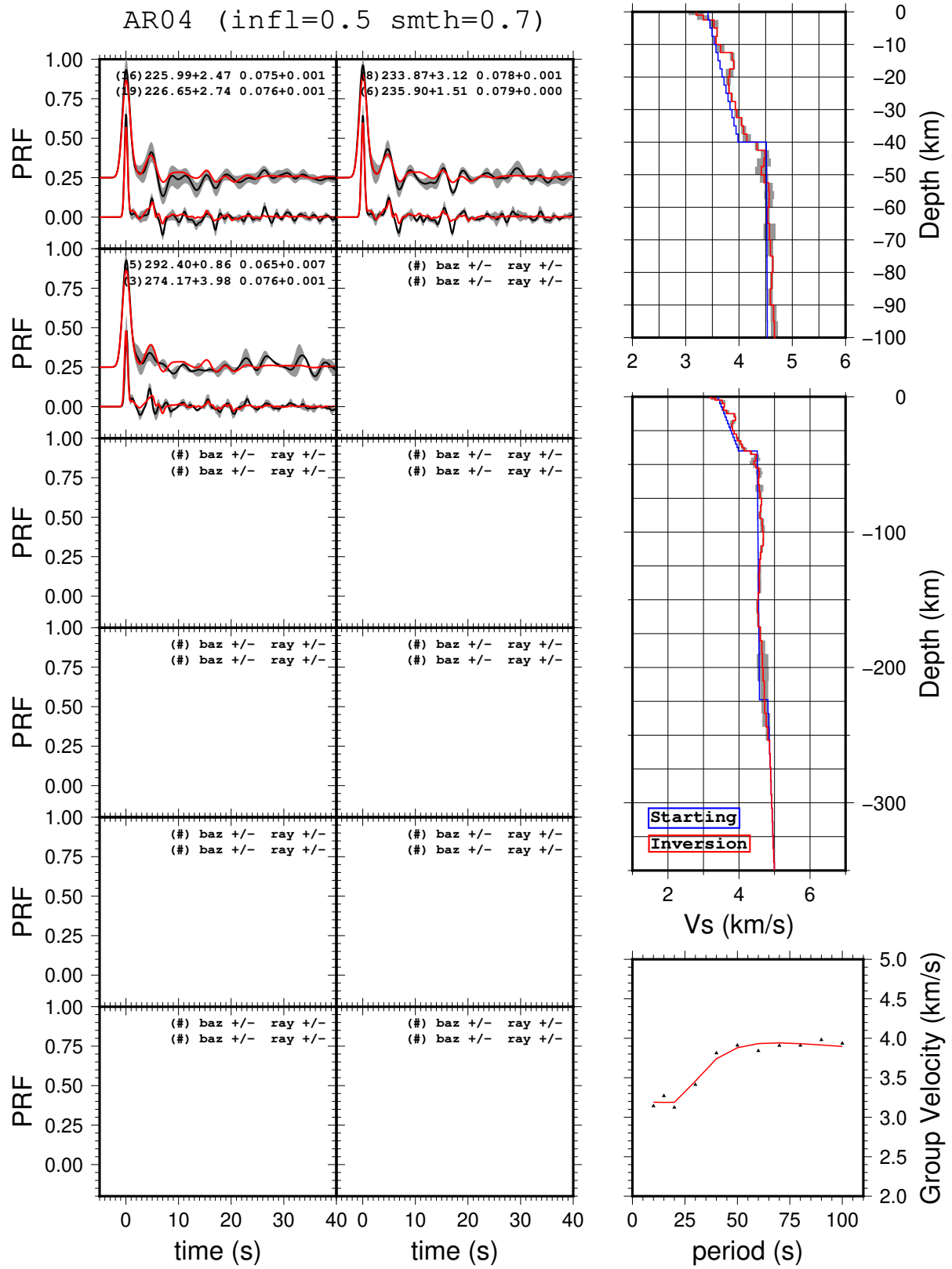


Figure S21. Detailed joint inversion results for station AR04. Same legend as in Figure S15.

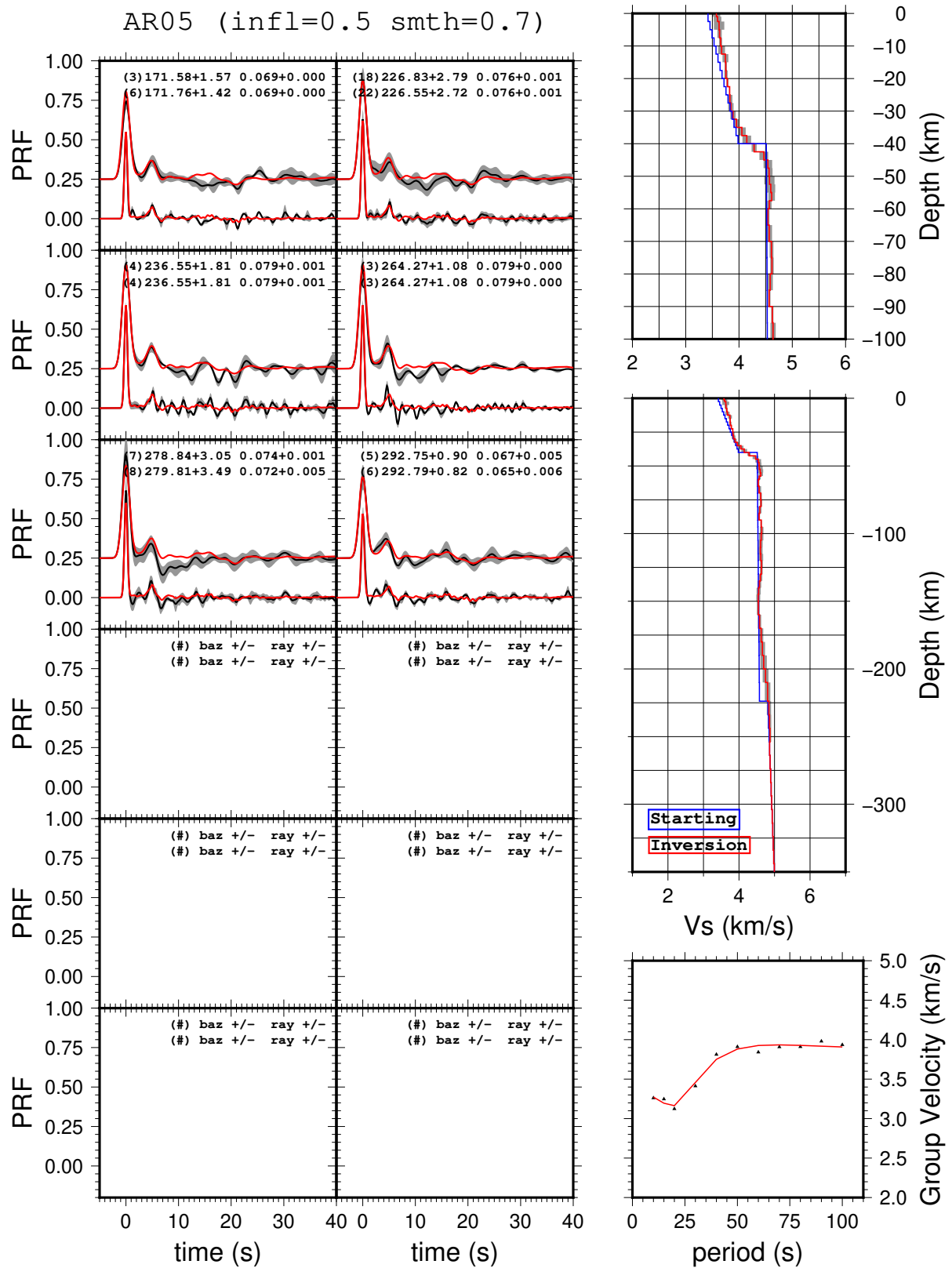


Figure S22. Detailed joint inversion results for station AR05. Same legend as in Figure S15.

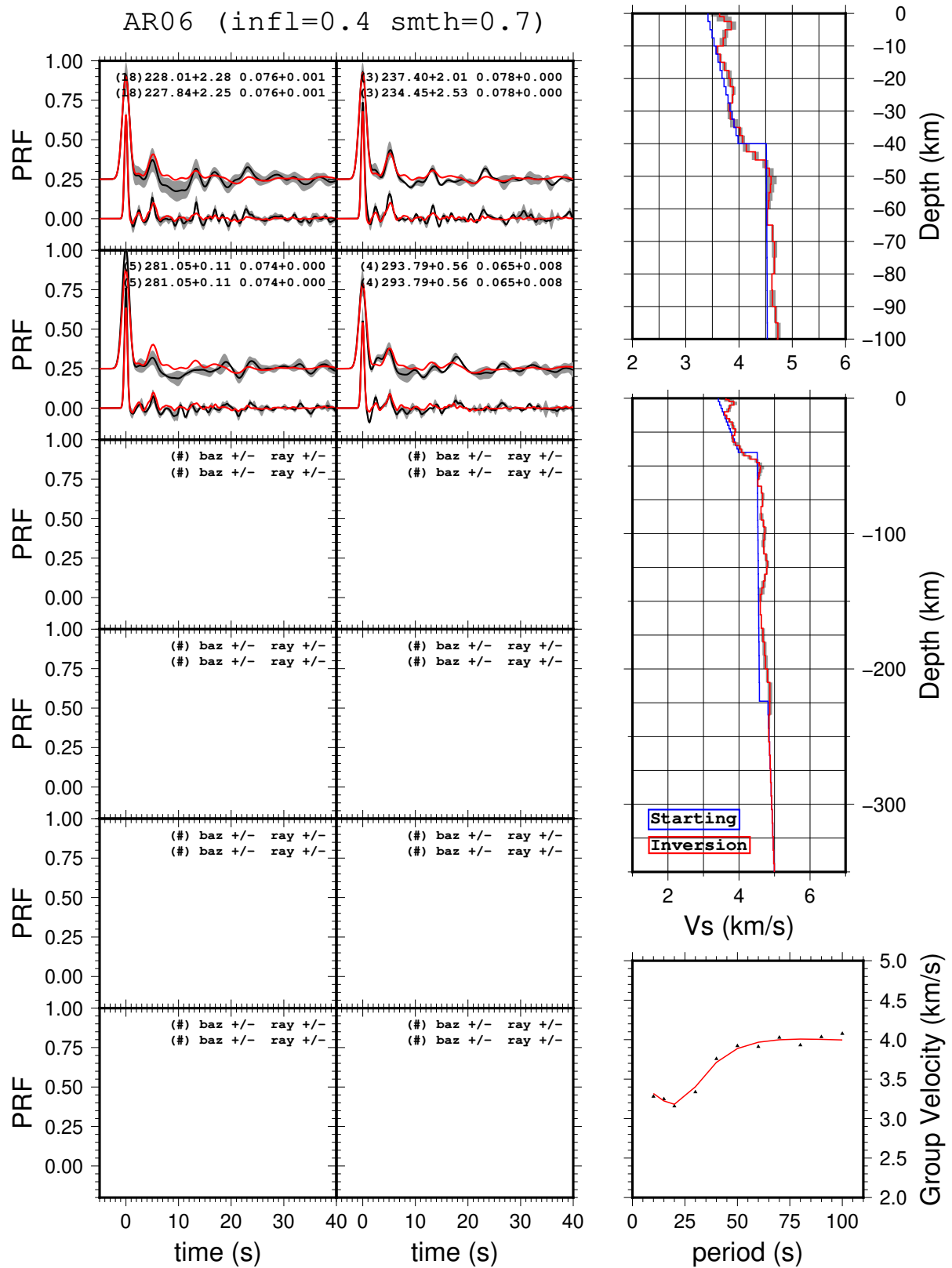


Figure S23. Detailed joint inversion results for station AR06. Same legend as in Figure S15.

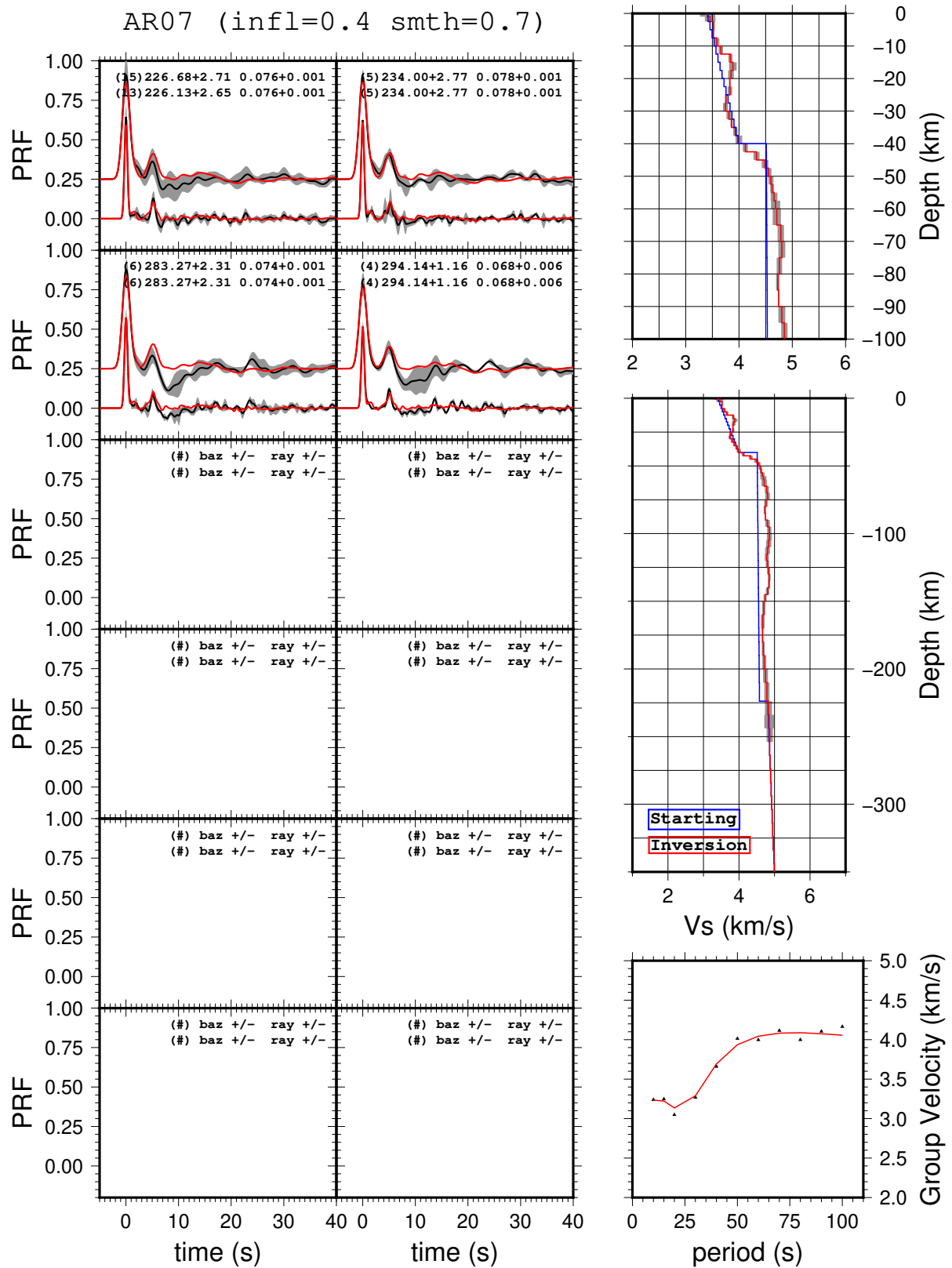


Figure S25. Detailed joint inversion results for station AR07. Same legend as in Figure S15.

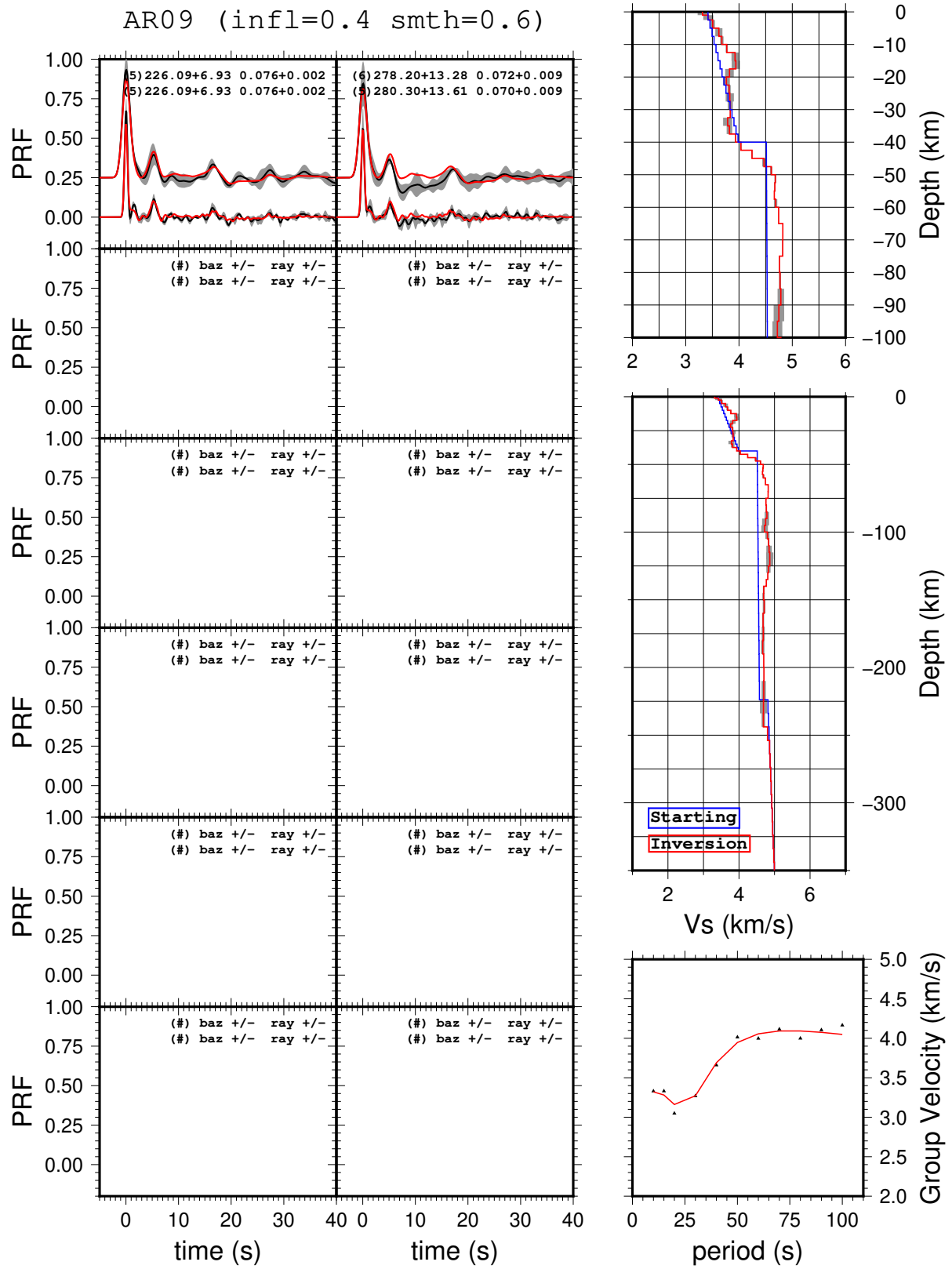


Figure S26. Detailed joint inversion results for station AR09. Same legend as in Figure S15.

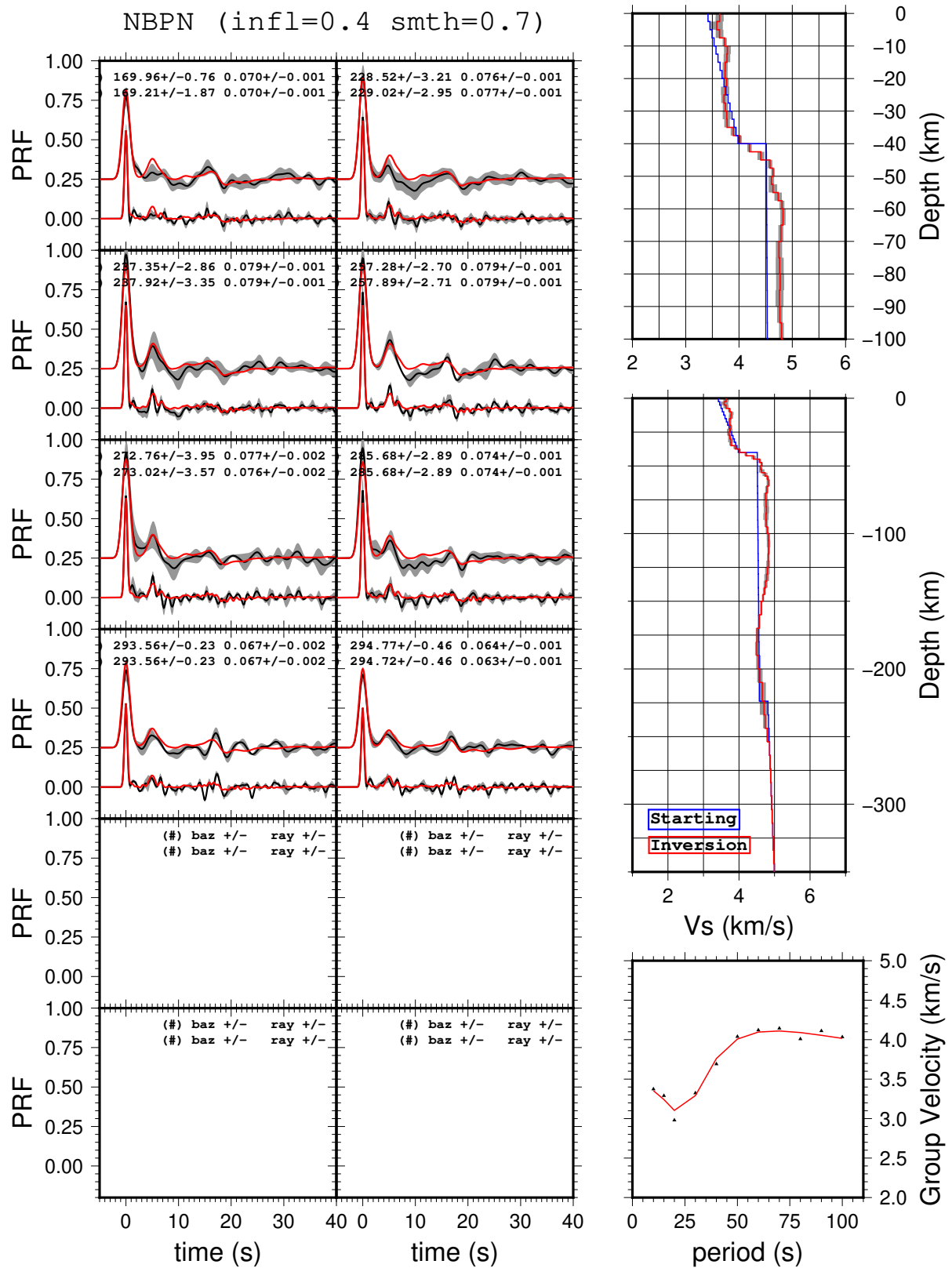


Figure S27. Detailed joint inversion results for station NBPN. Same legend as in Figure S15.

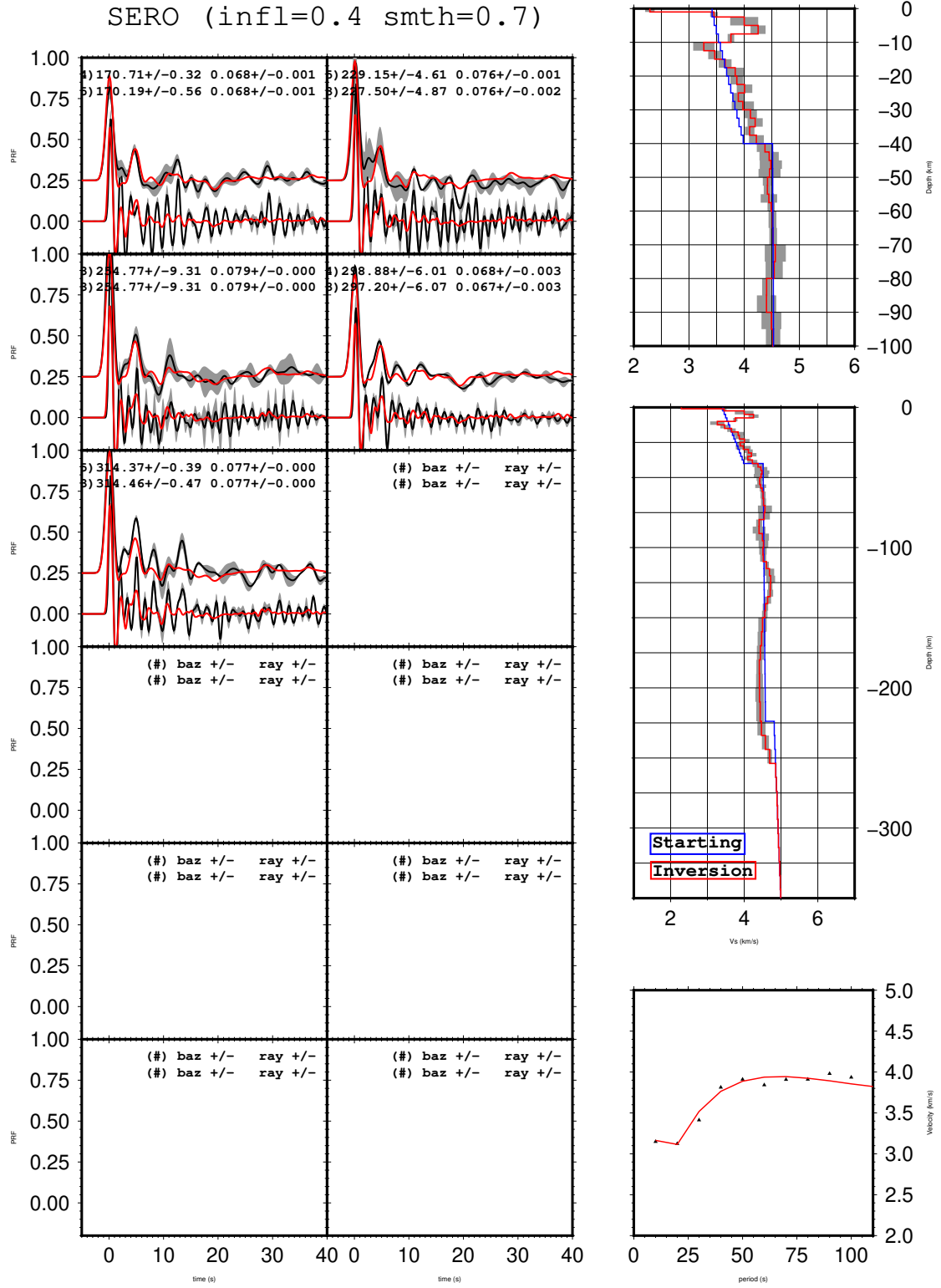


Figure S28. Detailed joint inversion results for station SERO. Same legend as in Figure S15.

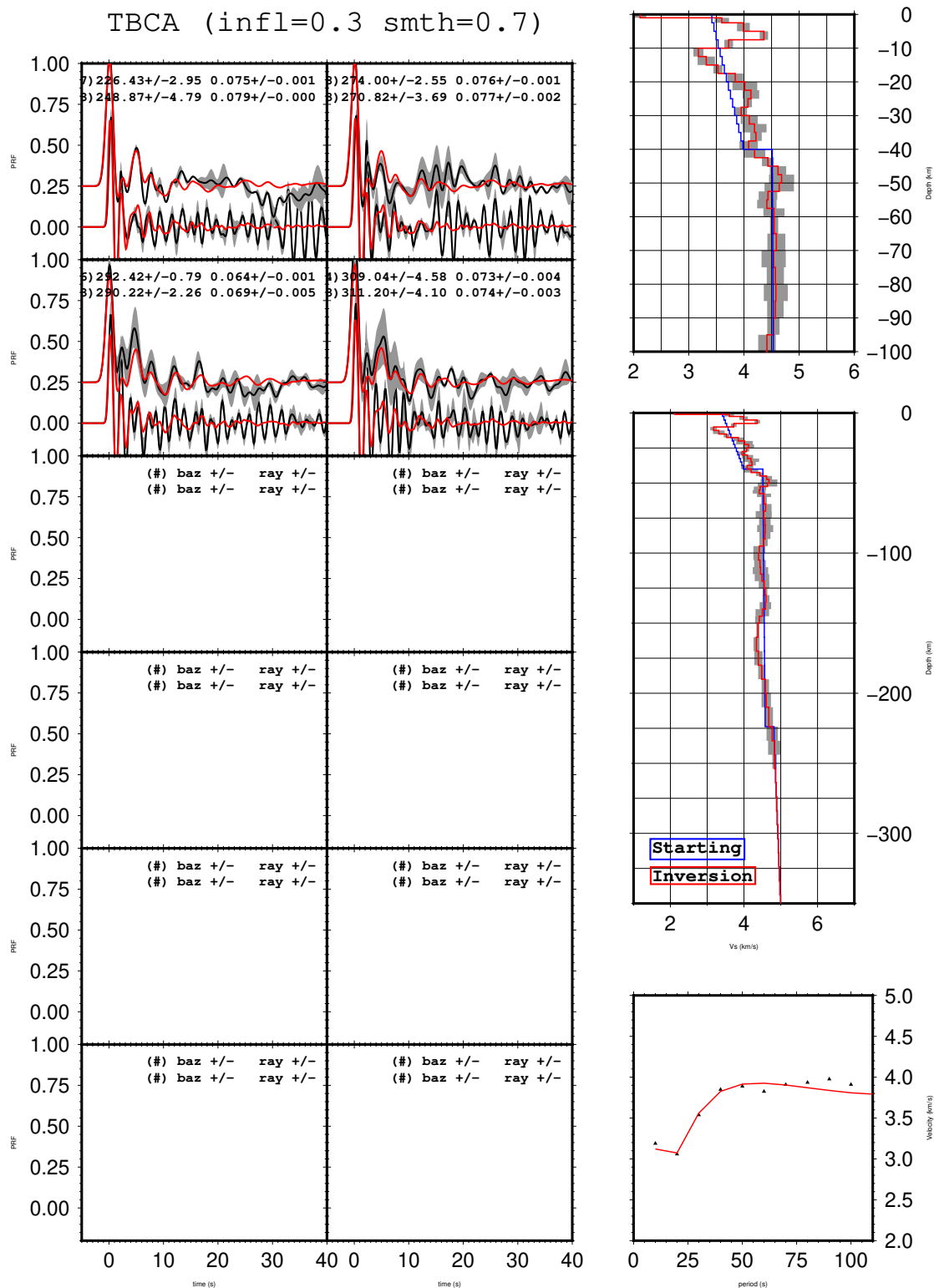


Figure S29. Detailed joint inversion results for station TBCA. Same legend as in Figure S15.

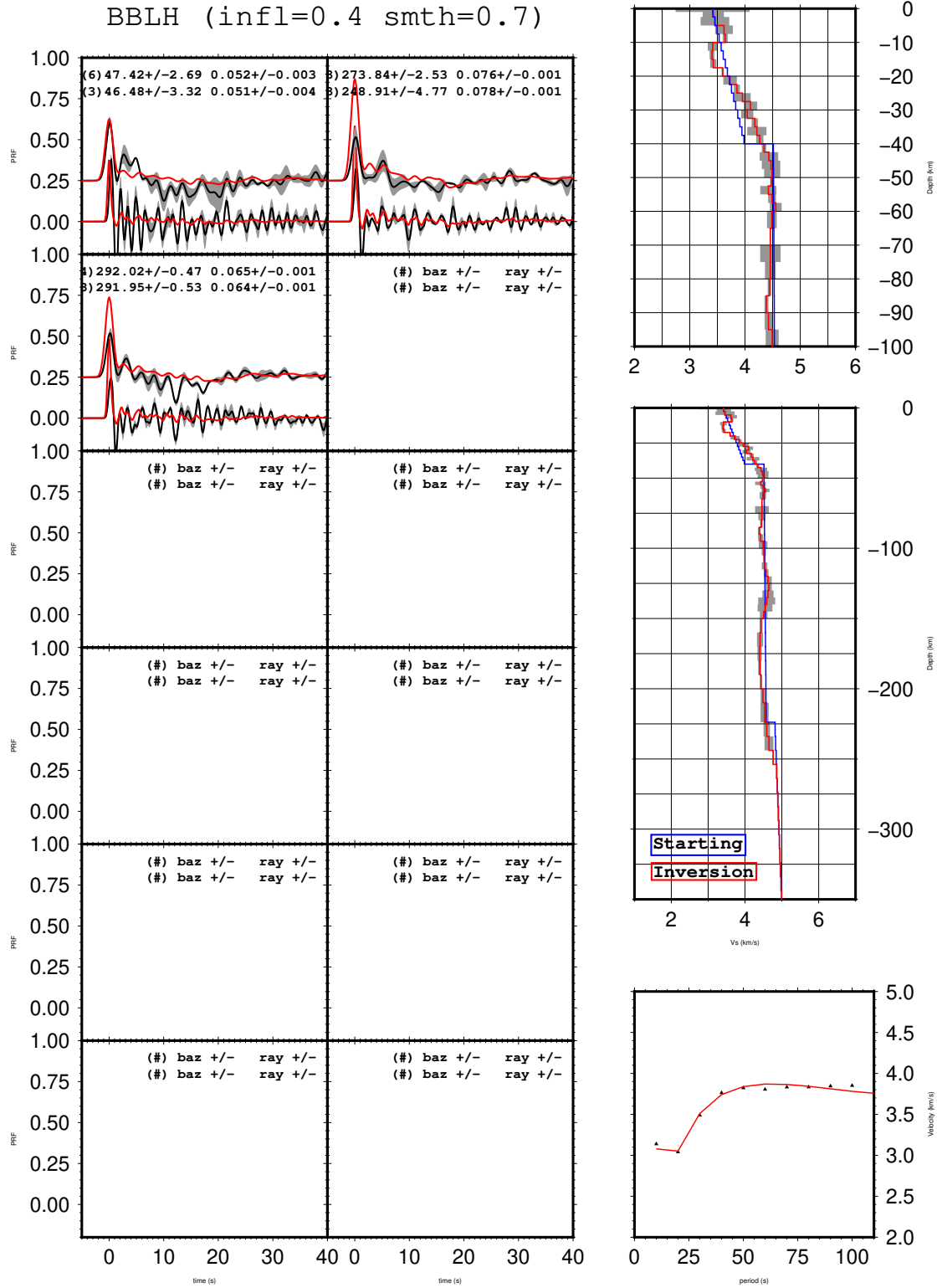


Figure S30. Detailed joint inversion results for station BBLH. Same legend as in Figure S15.

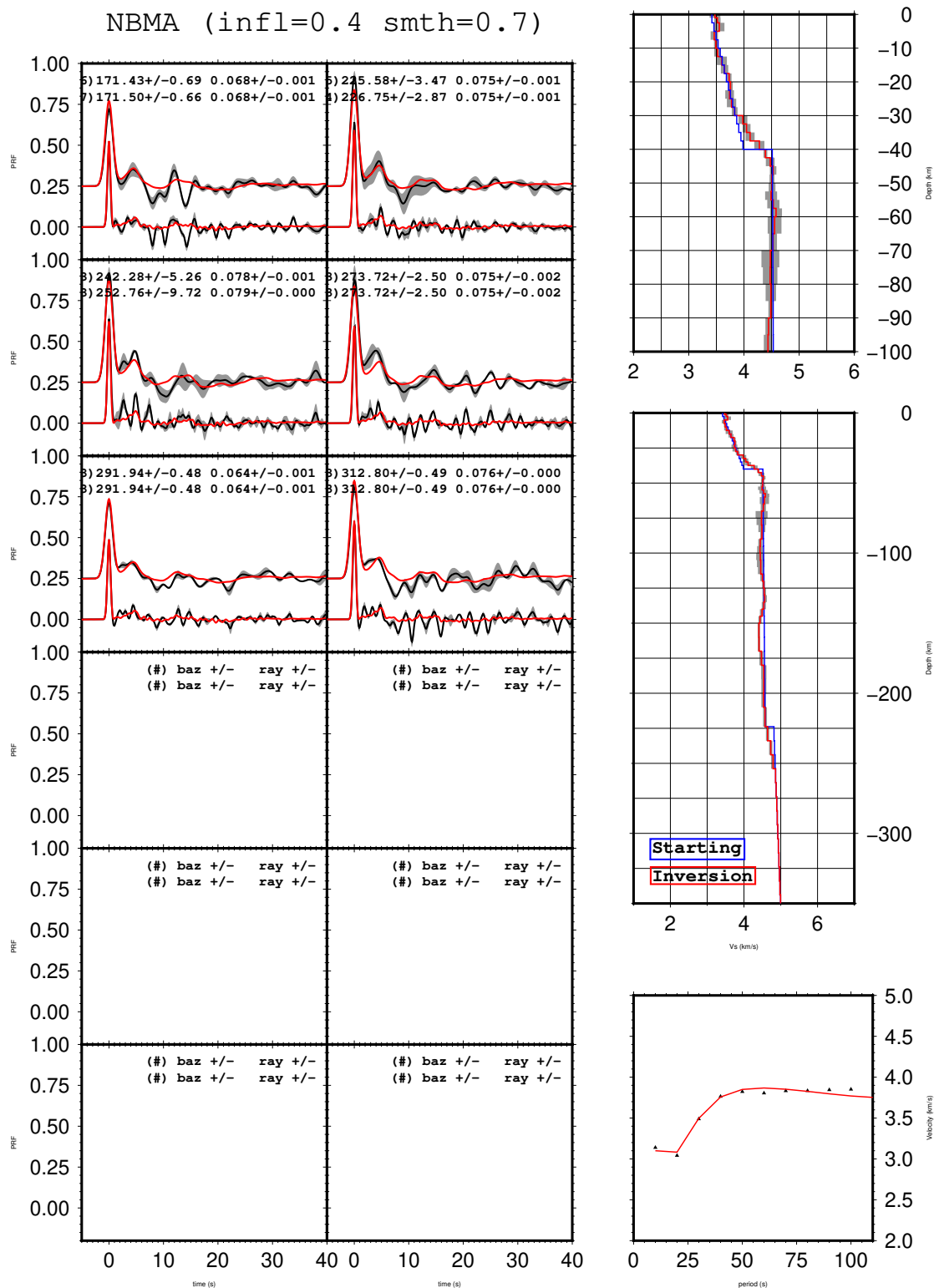


Figure S31. Detailed joint inversion results for station NBMA. Same legend as in Figure S15.

Appendix B

Supplementary information chapter 5

This appendix presents the supplementary material of the article entitled: *Joint inversion of magnetotelluric and seismological observations: Imaging the electric Moho in NE Brazil*. This supporting information that complement the text and figures provided in the manuscript.

Supplementary material

Joint inversion of magnetotelluric and seismological observations:
Imaging the electric Moho in NE Brazil
Ana M. Nemocón et al.

15 de Junho de 2023

Results of joint inversion of magnetotelluric, receiver function and dispersion curves performed for 10 inversion layers for the BODES stations.

The results shows for each station:

1. Trade-off plots or Pareto front. Each plot shows two objective function values on axis X and Y, and the third objective function value in color. The optimum model is marked by red star.
2. The best electric and seismic models from Pareto (green) and optimum model selected (black). Comparison on seismic model with S-velocity model by Nemocon:inpress-a (red) and PREM model (blue).
3. Comparison between predict (red) and observed data (black) for magnetotelluric and teleseismic data.

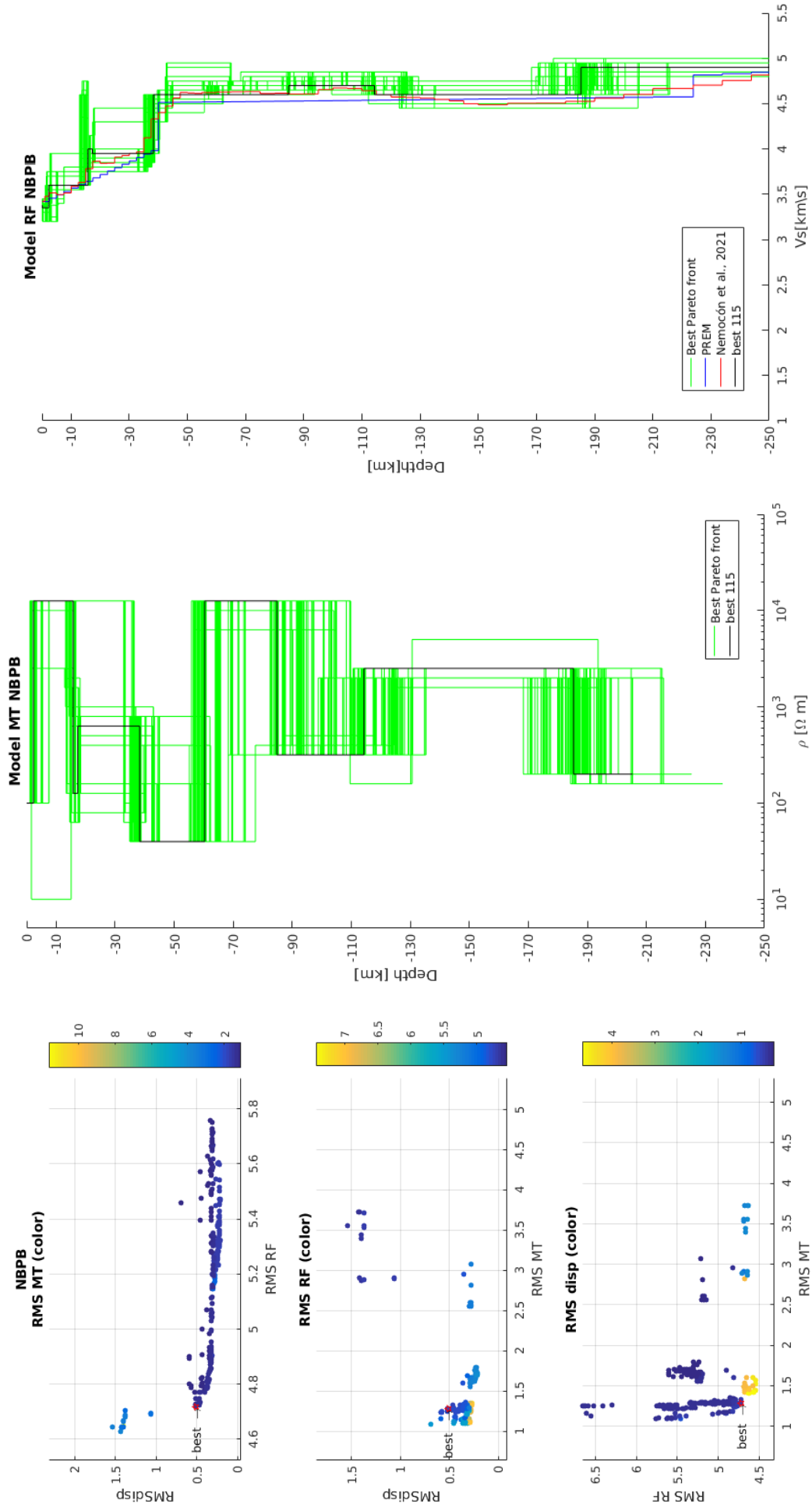


Figure 1: Station NBPB.

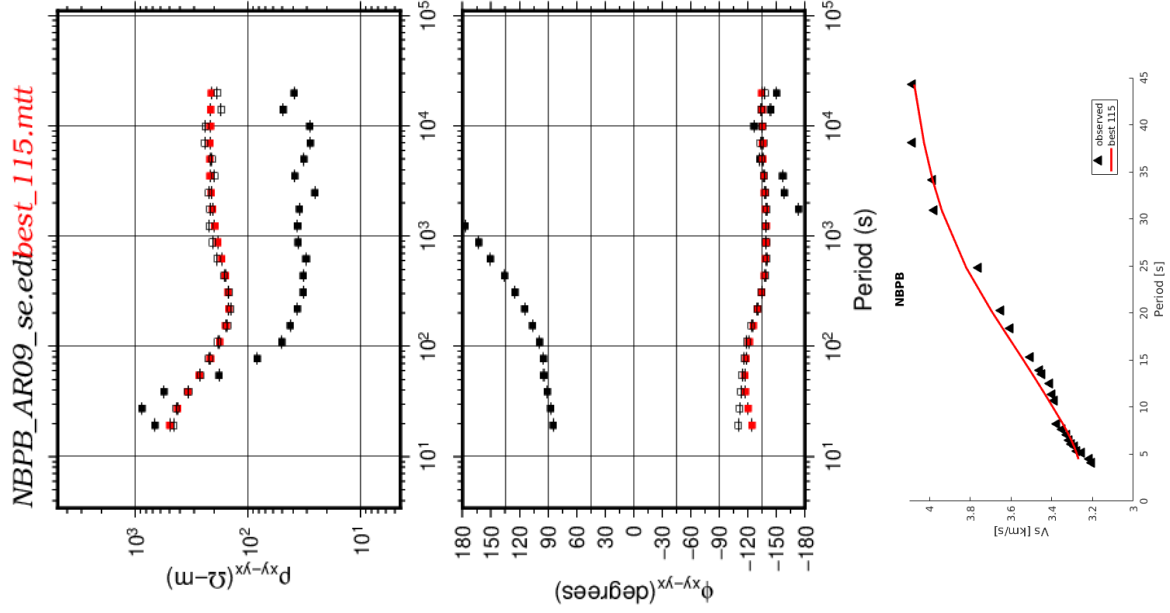
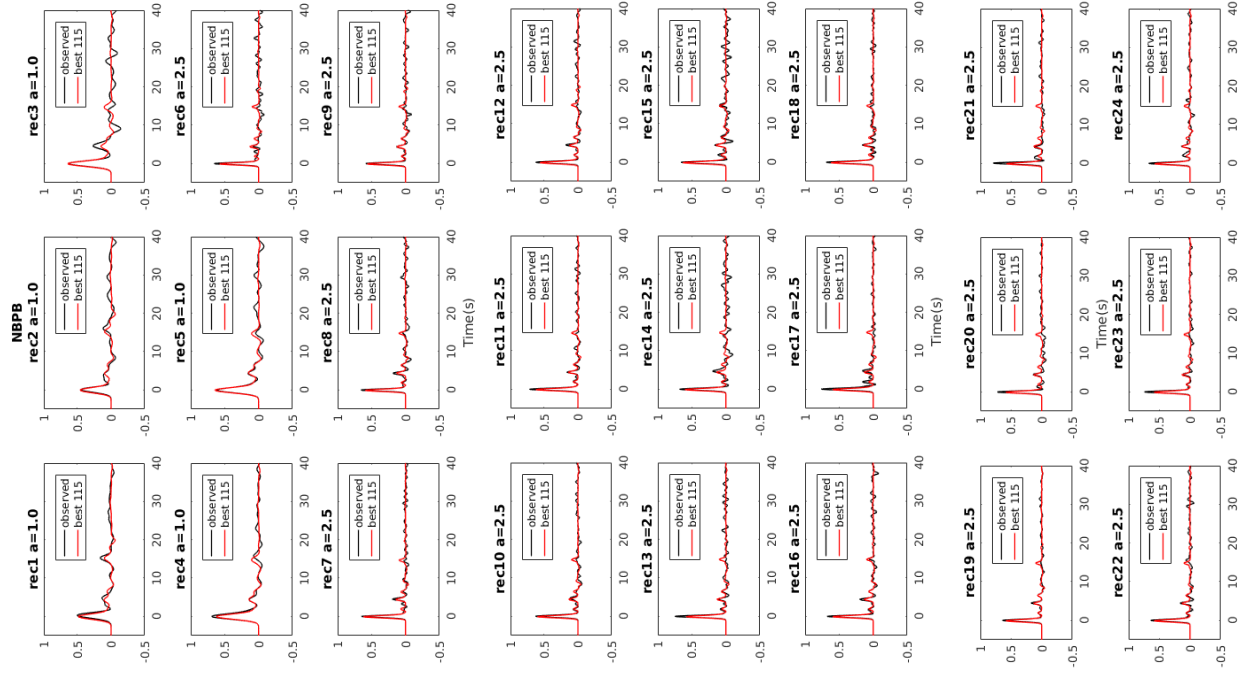


Figure 2: Station NBPB.

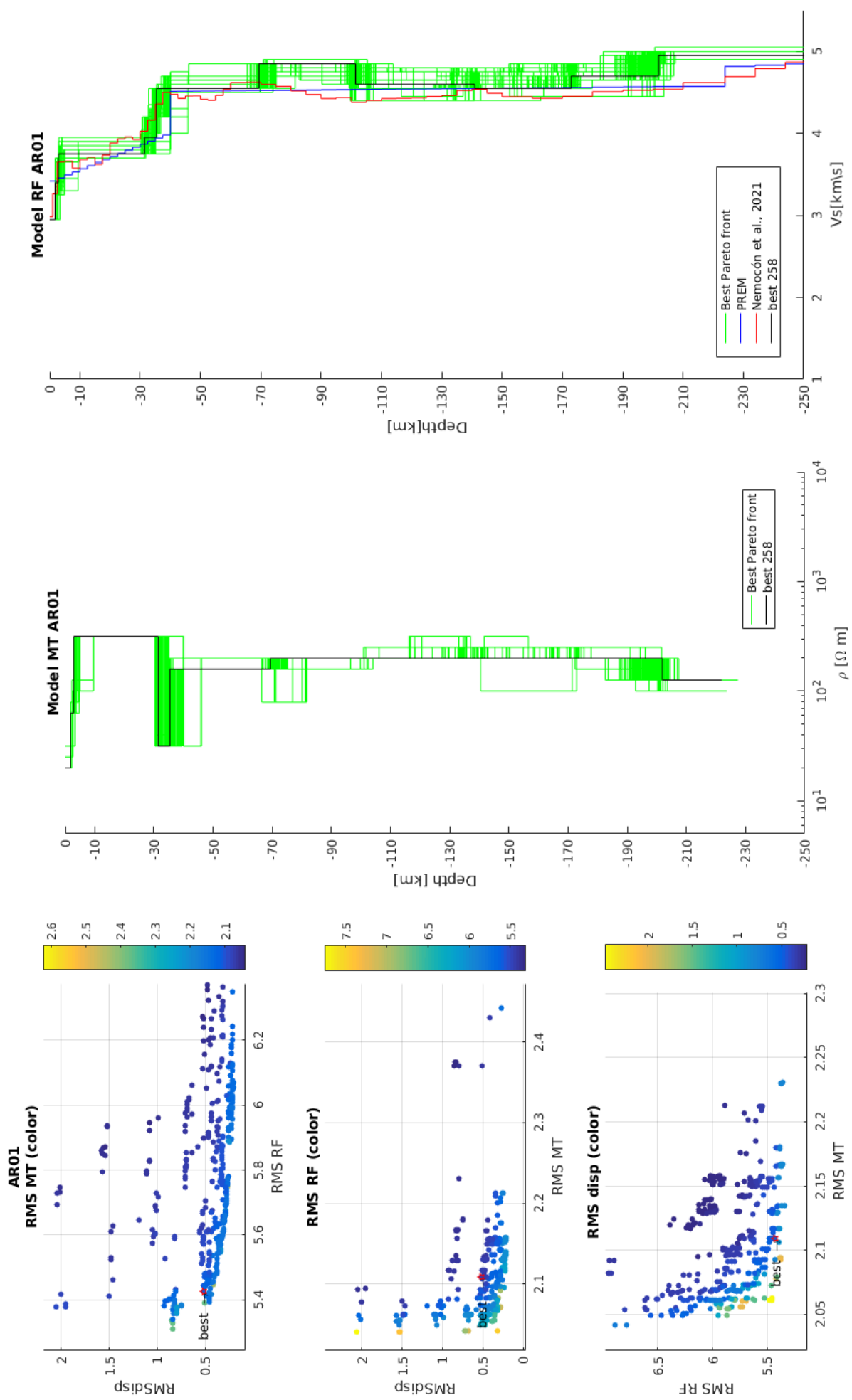


Figure 3: Station AR01.

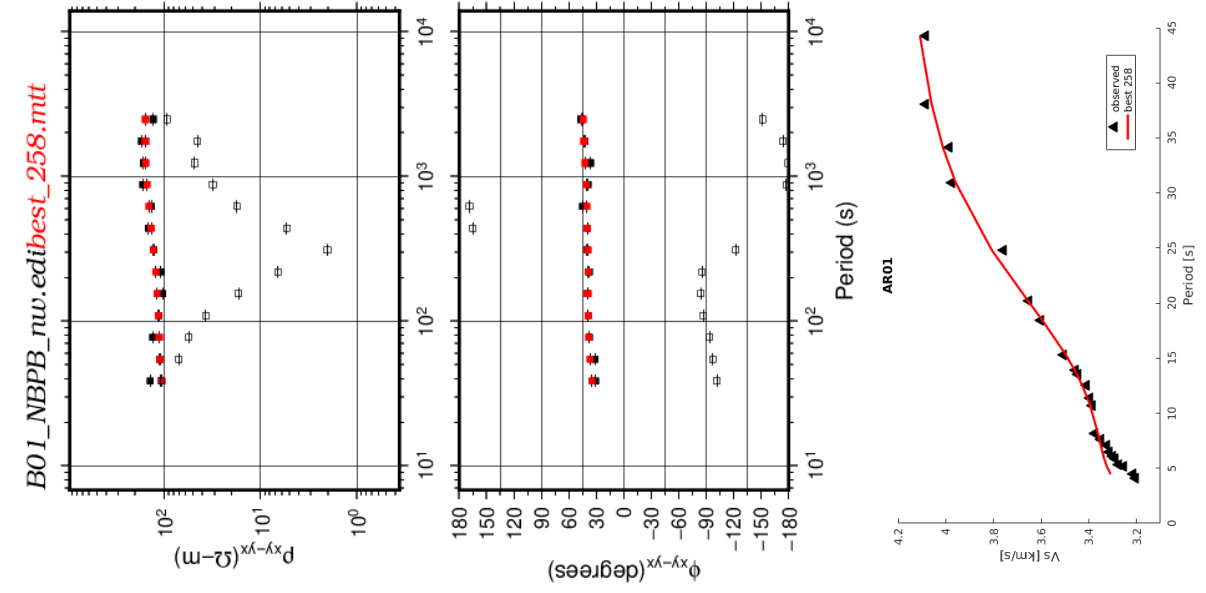


Figure 4: Station AR01.

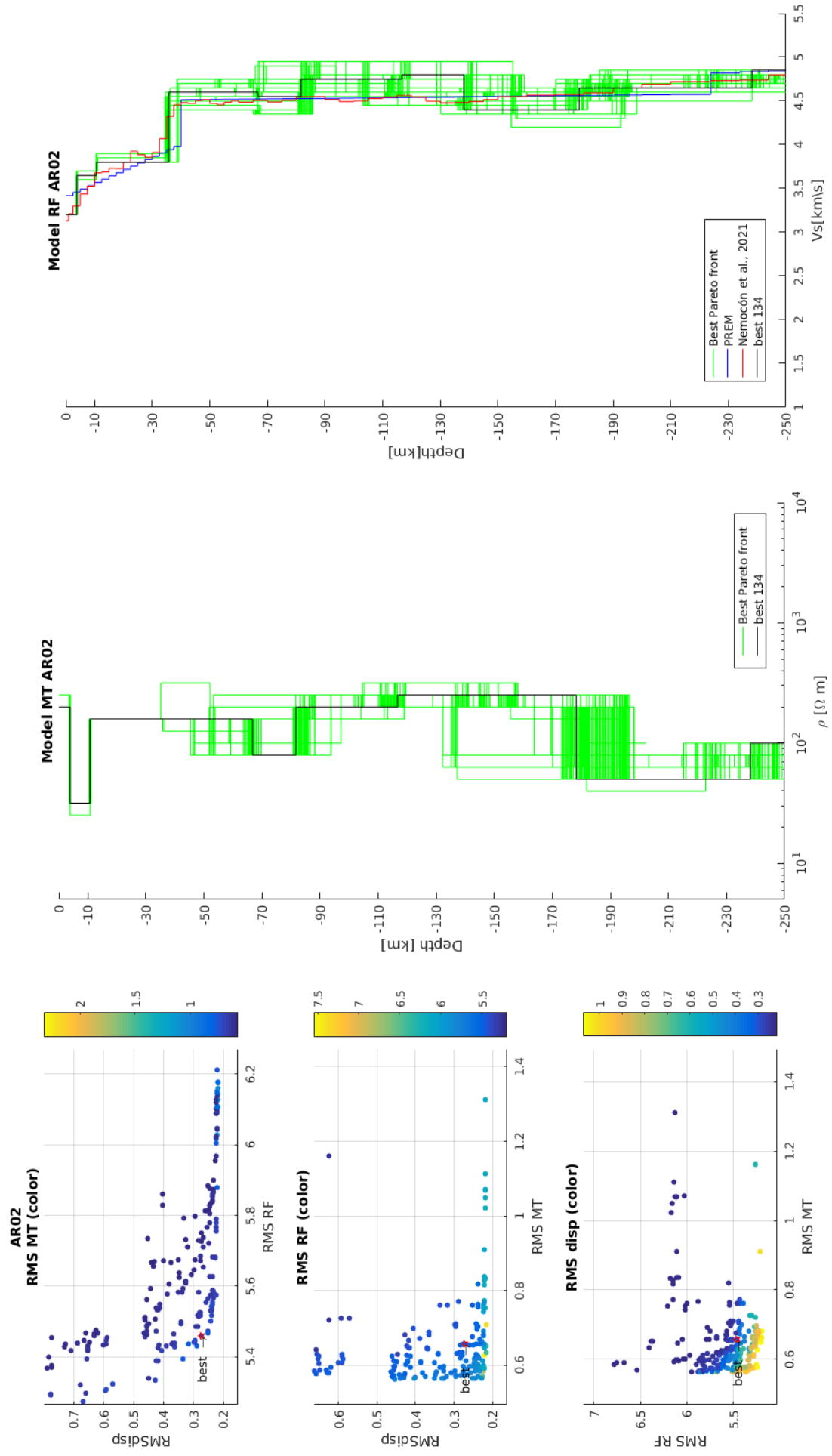


Figure 5: *Station AR02.*

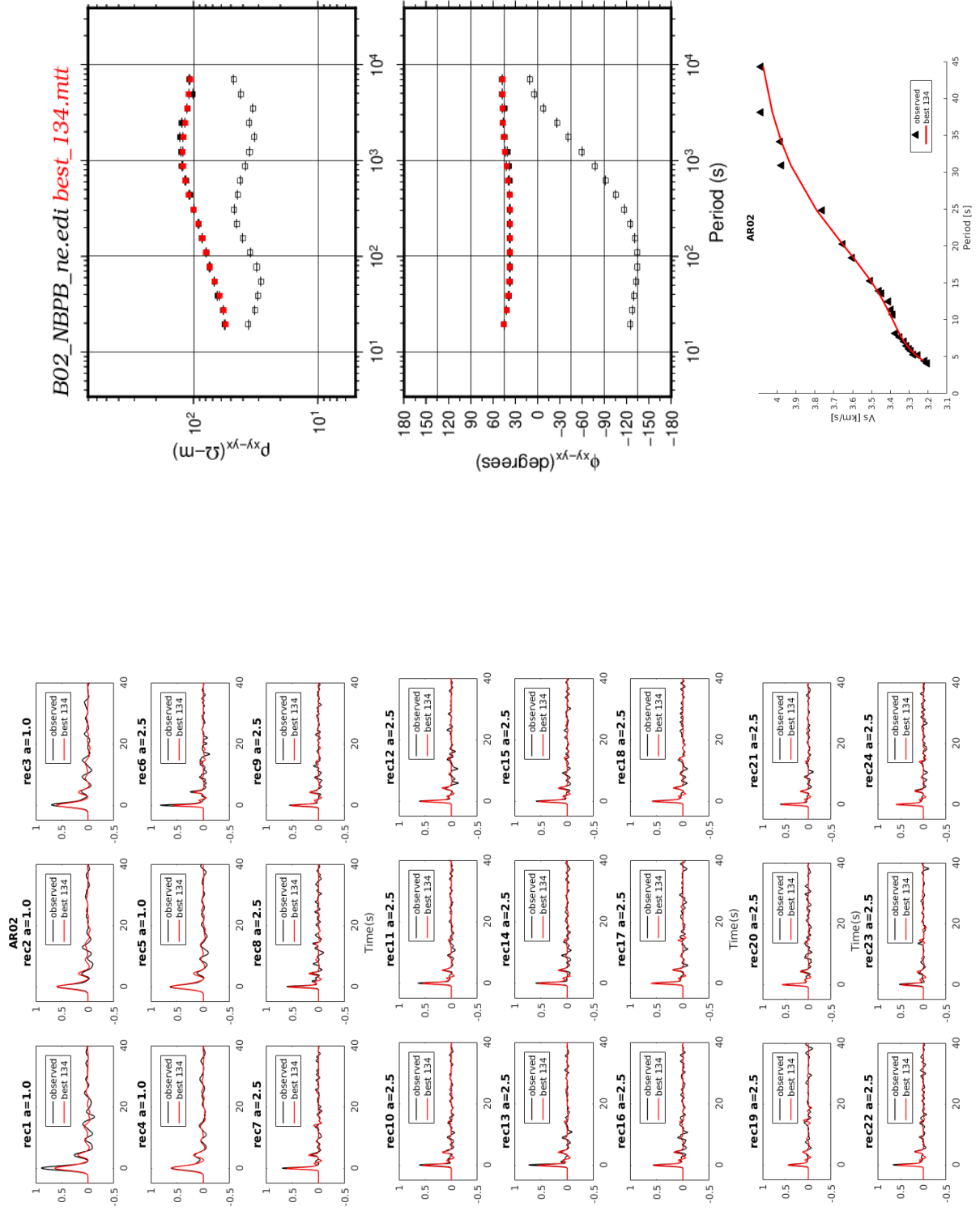


Figure 6: Station AR02.

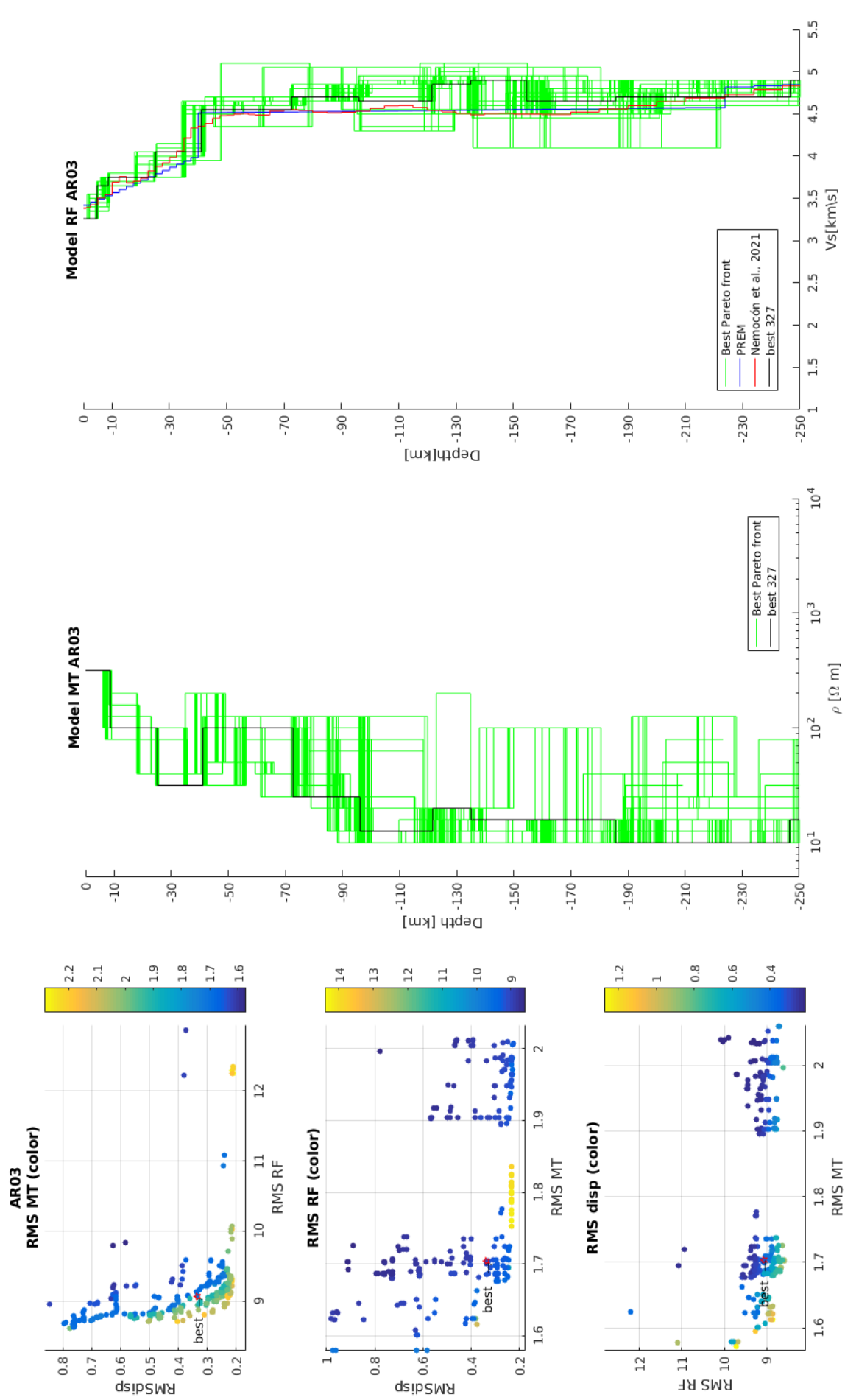


Figure 7: Station AR03.

B03_NBPB_sw.edi *best_327.mtt*

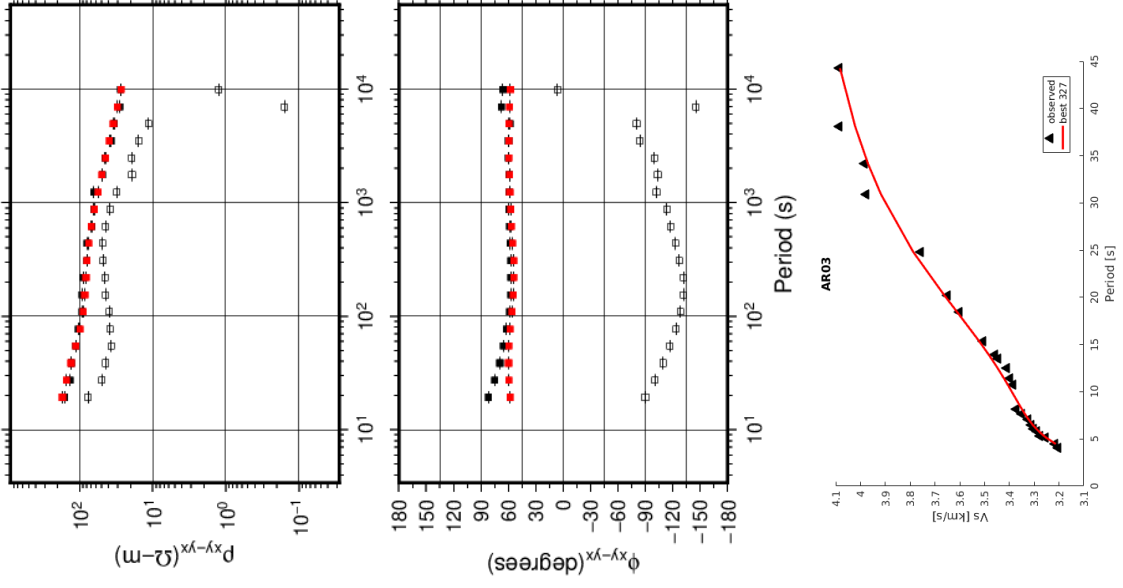
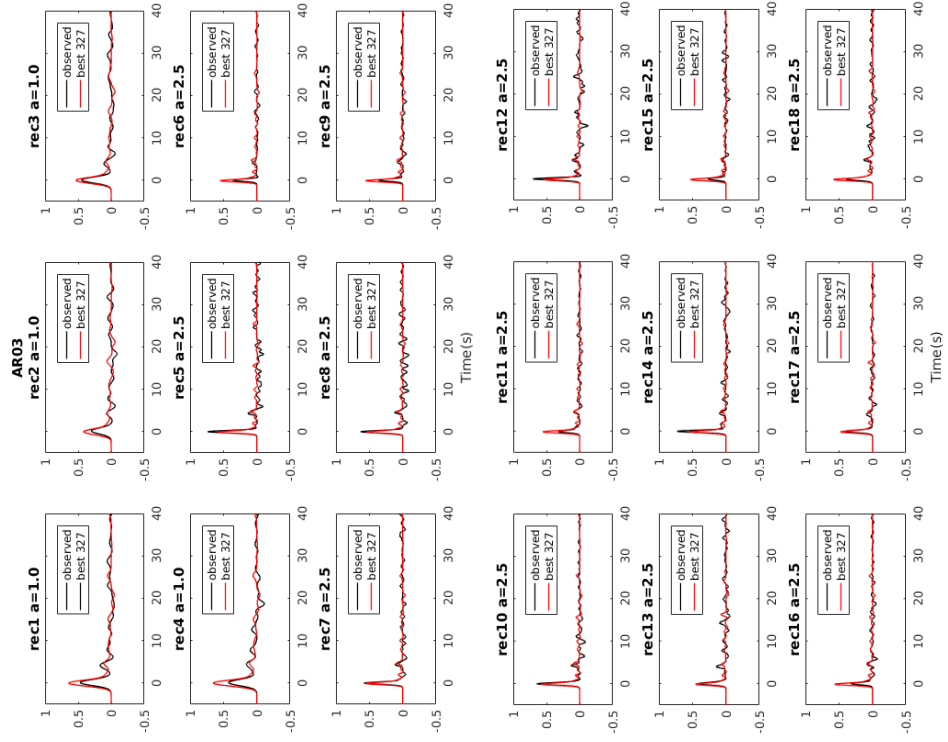


Figure 8: Station AR03.



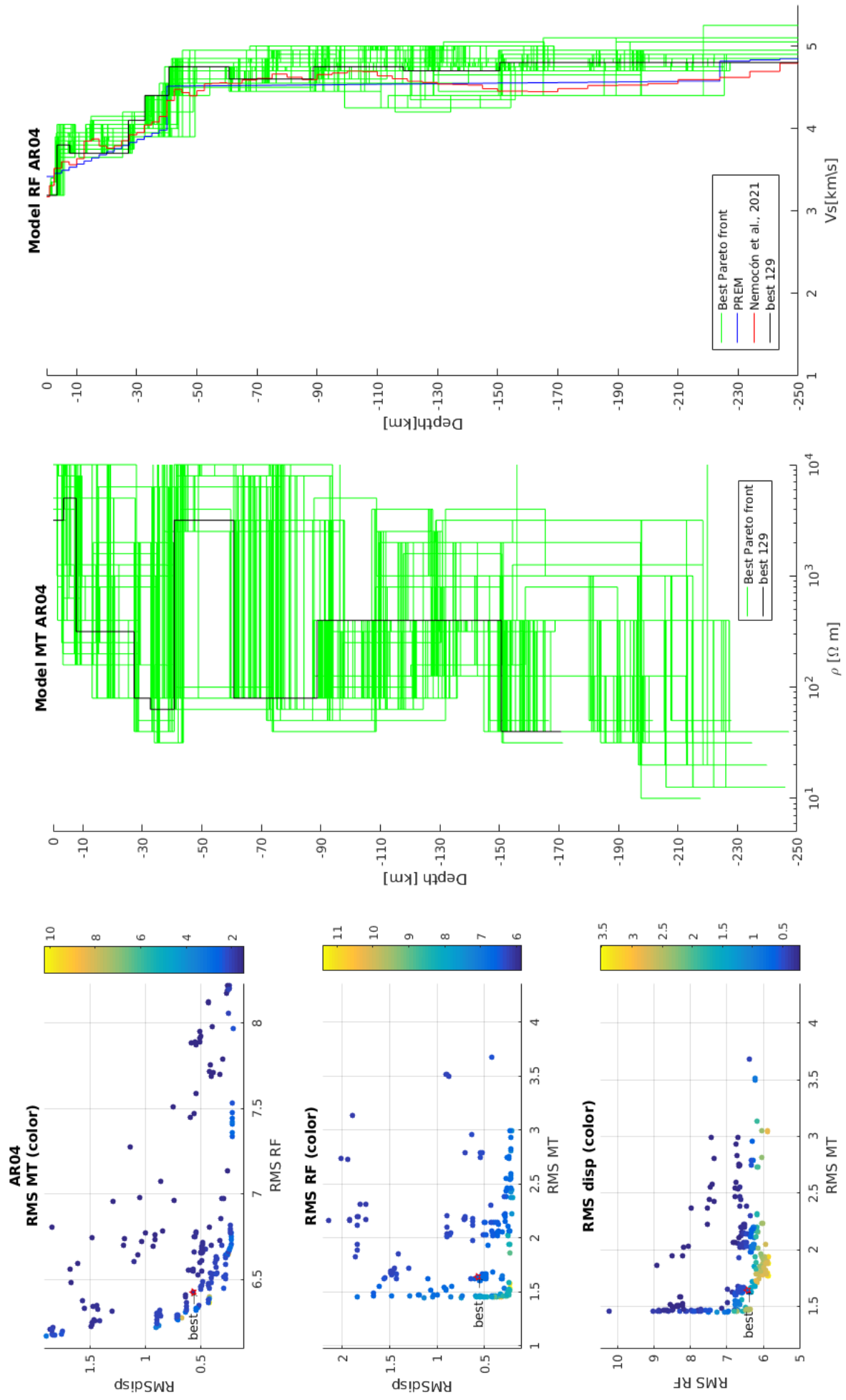


Figure 9: Station AR04.

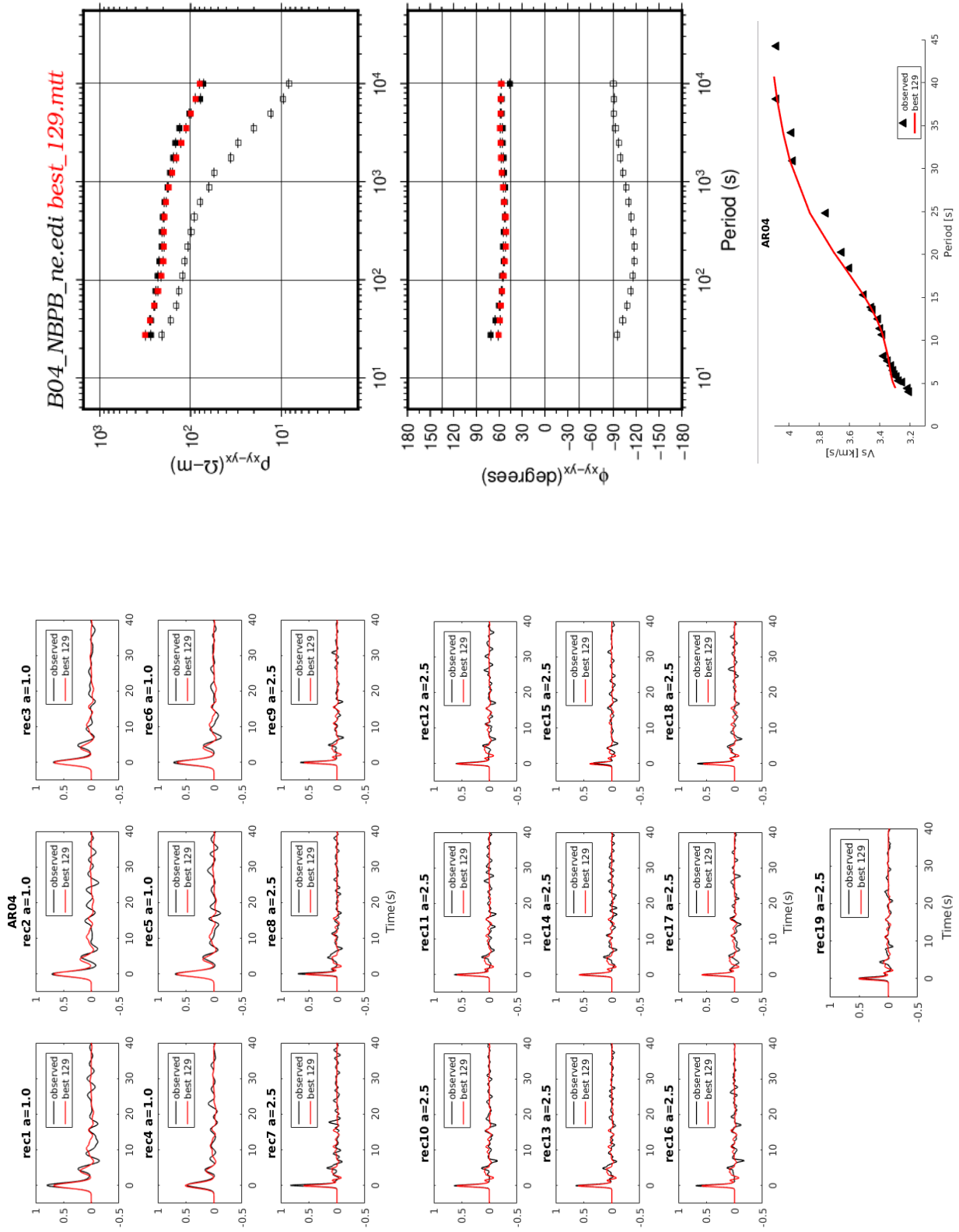


Figure 10: Station AR04.

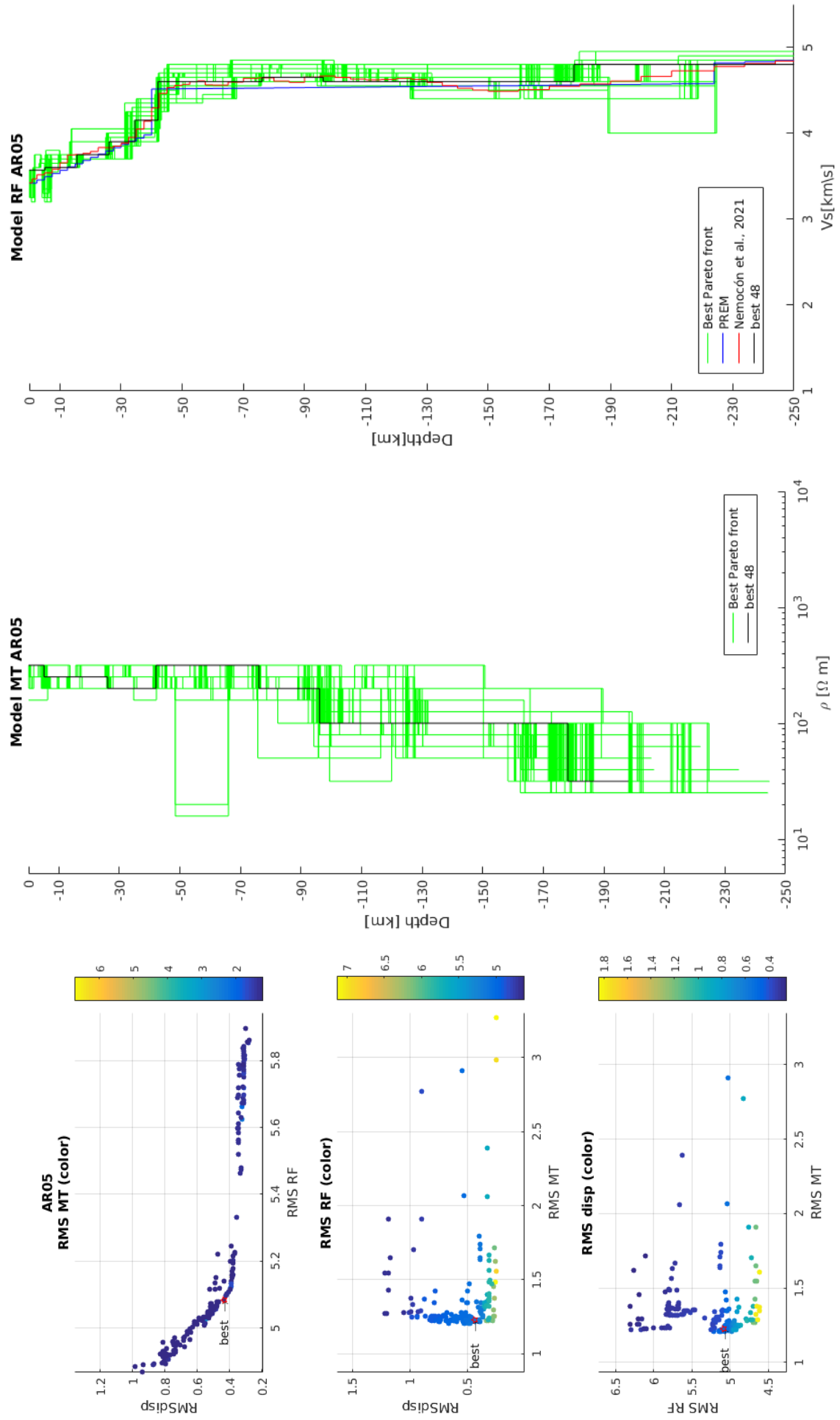


Figure 11: Station AR05.

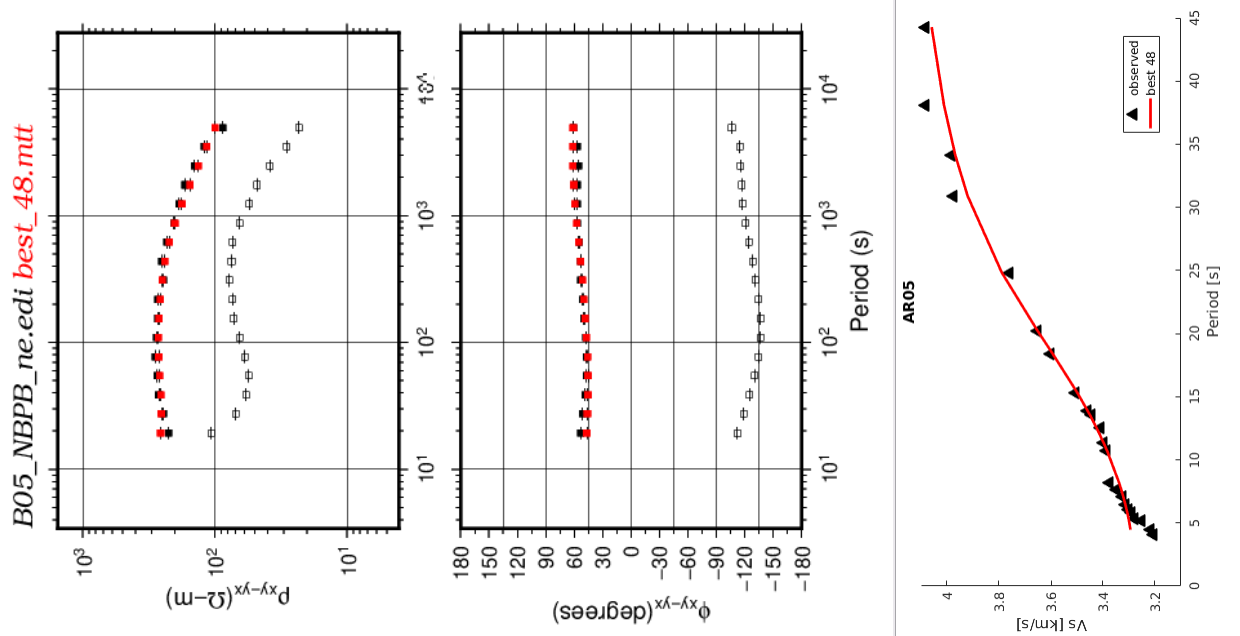
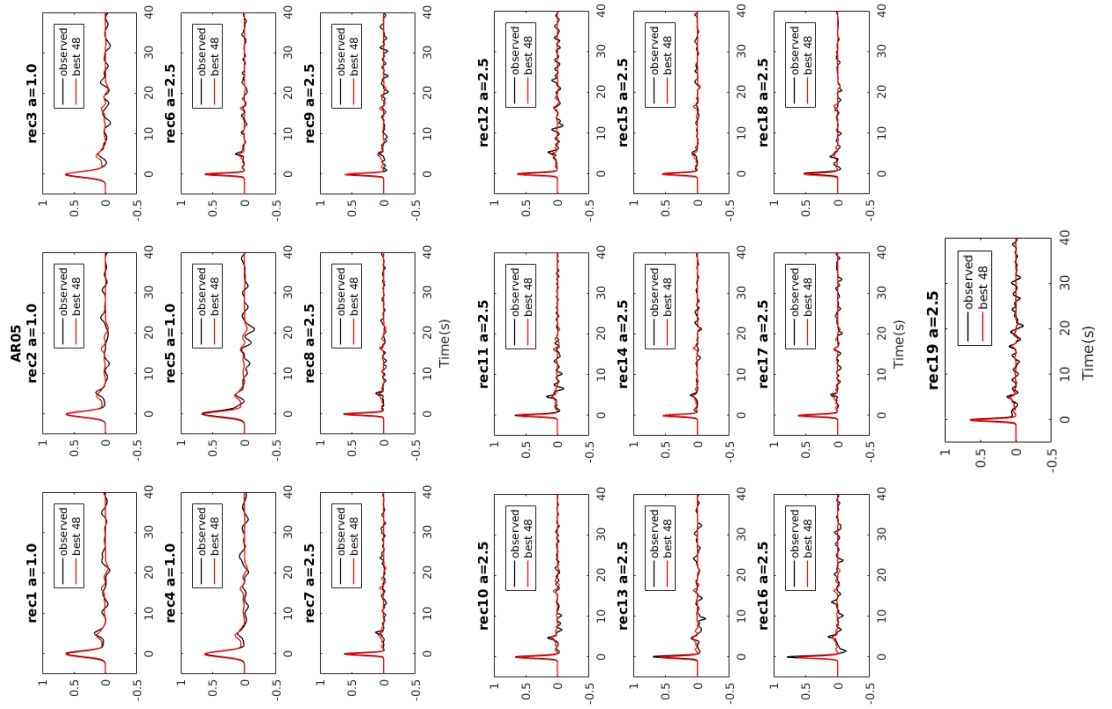


Figure 12: Station AR05.



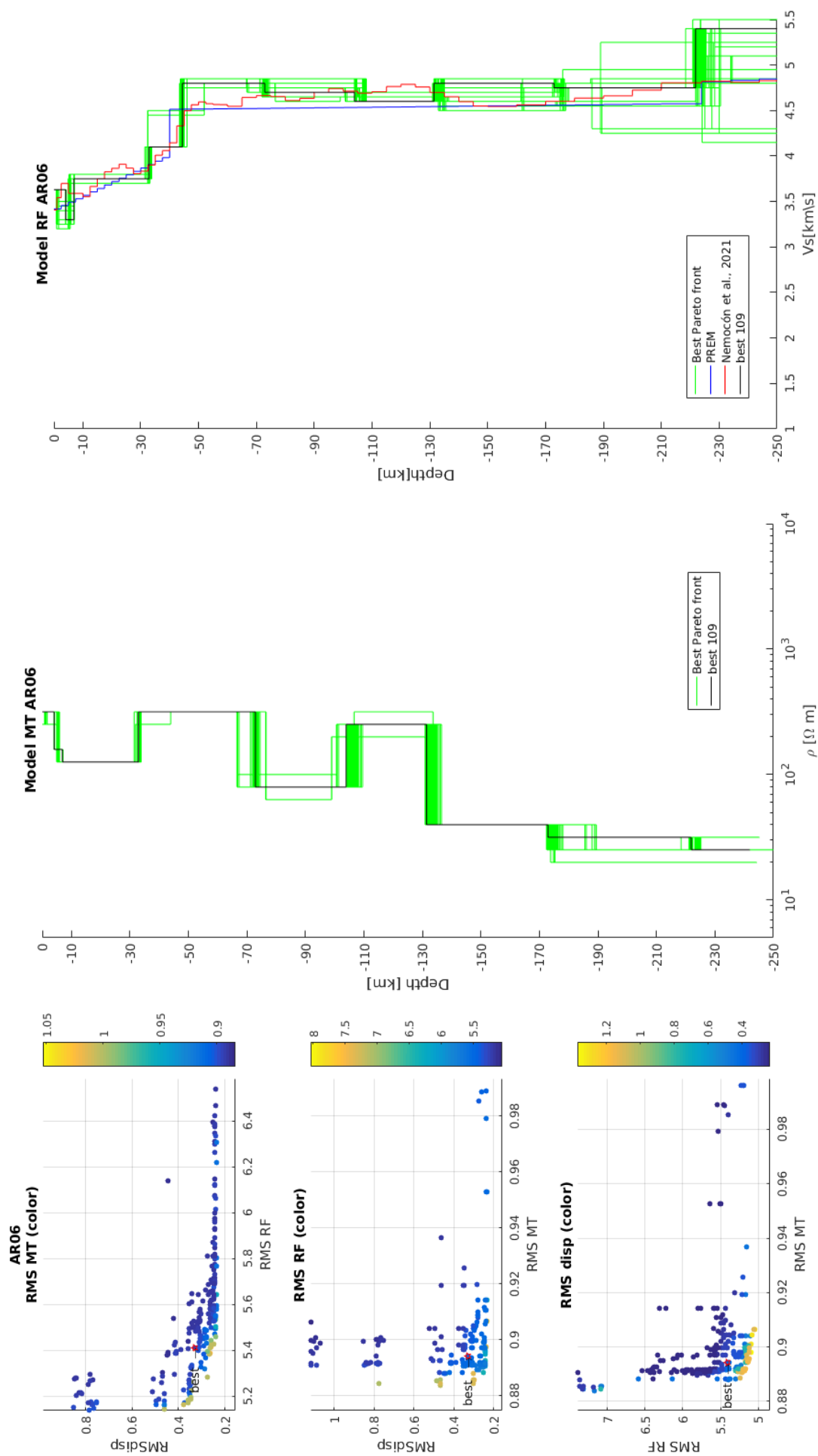


Figure 13: Station AR06.

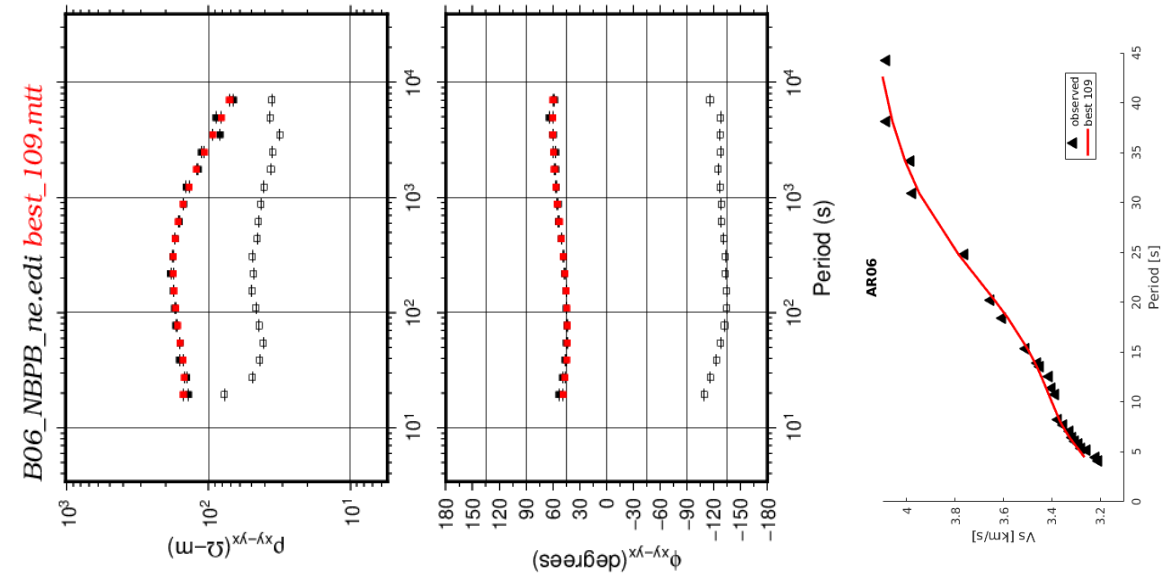
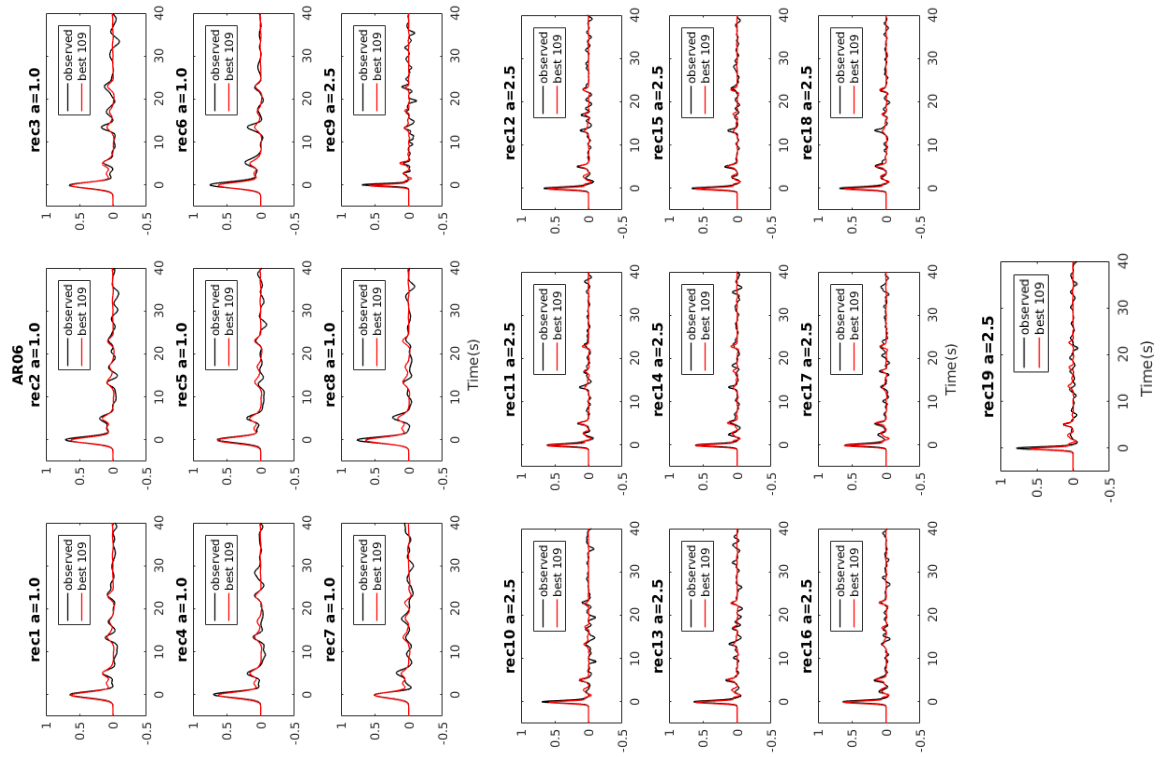


Figure 14: Station AR06.



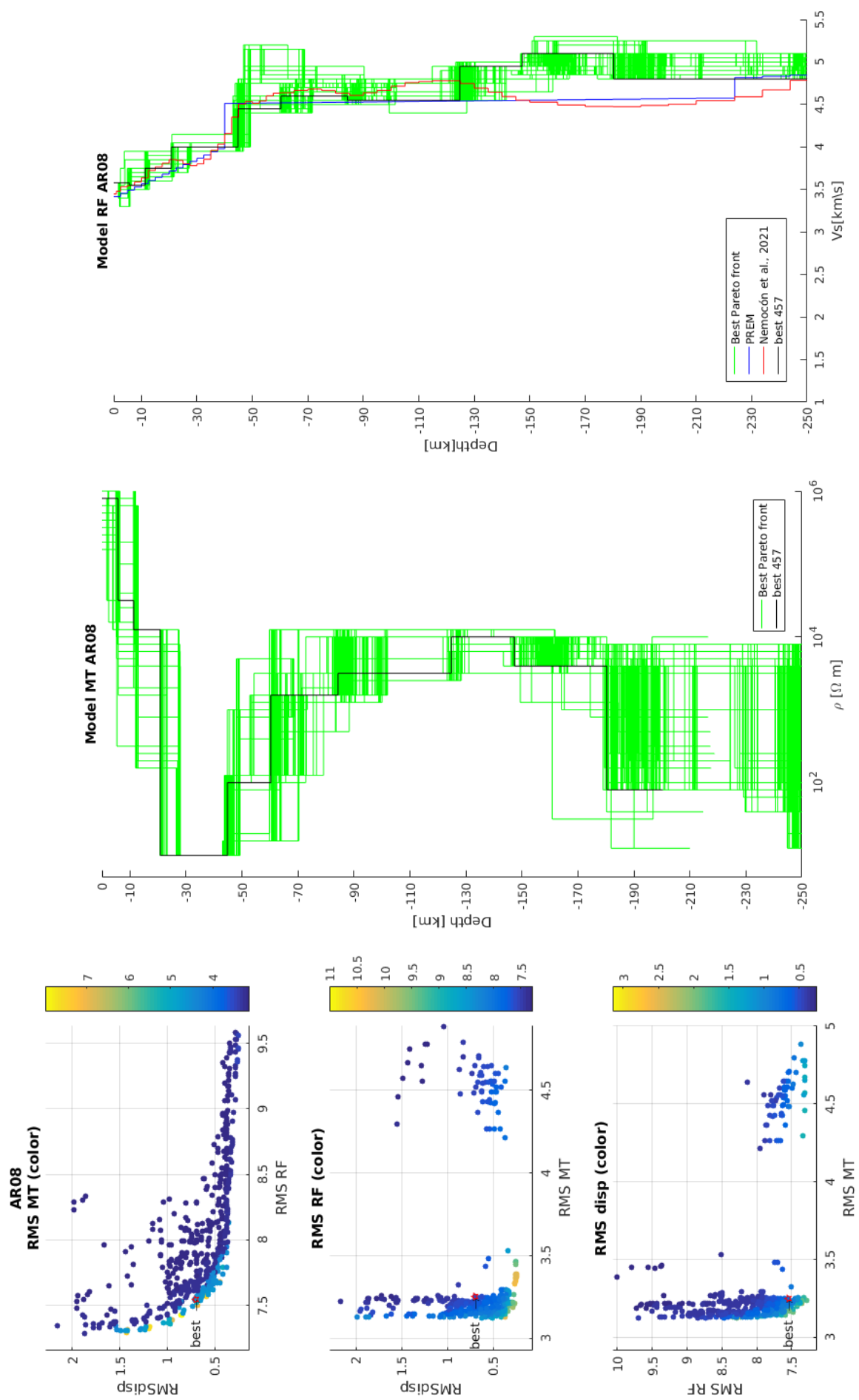


Figure 15: Station AR08.

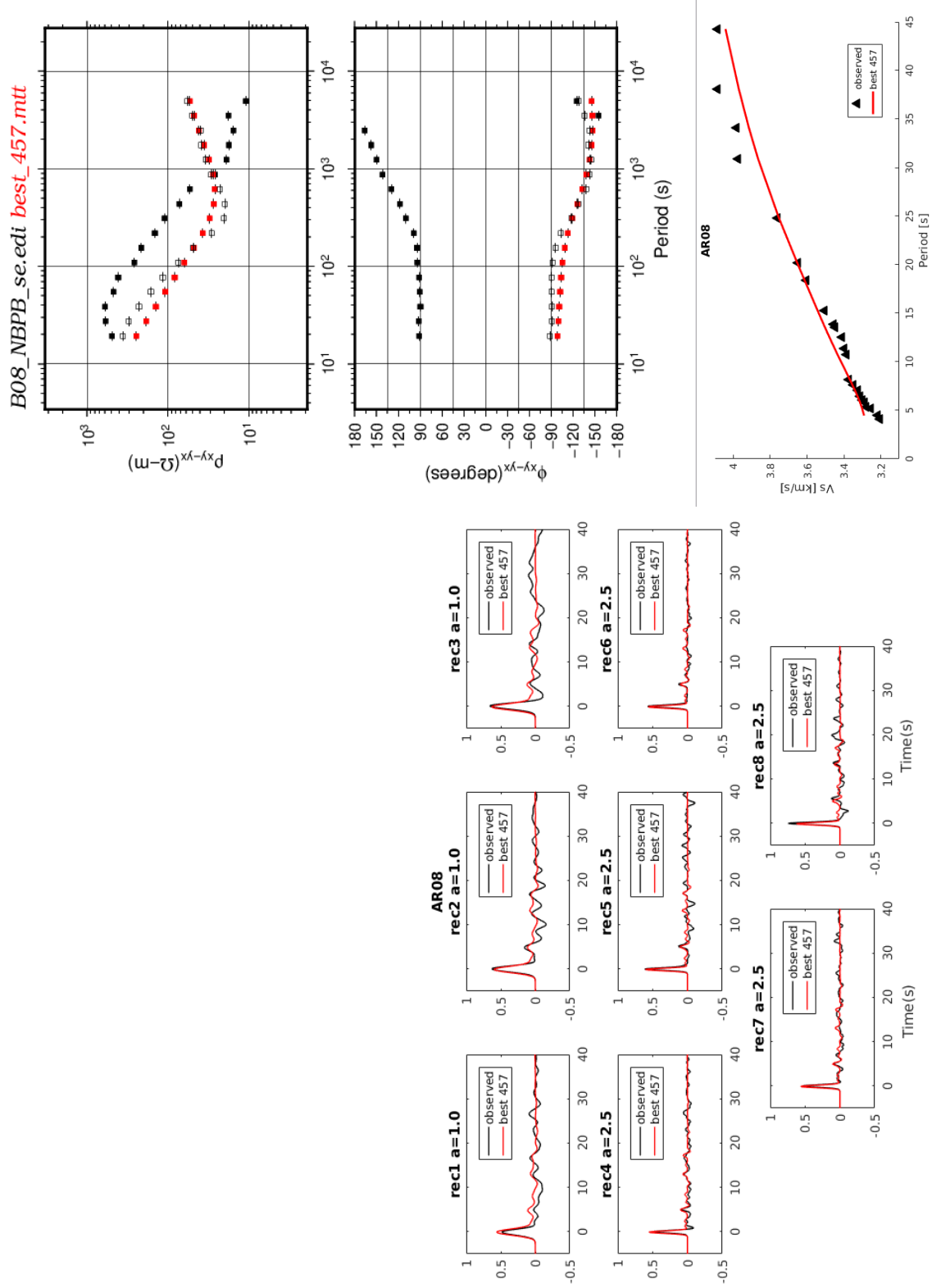


Figure 16: Station AR08.

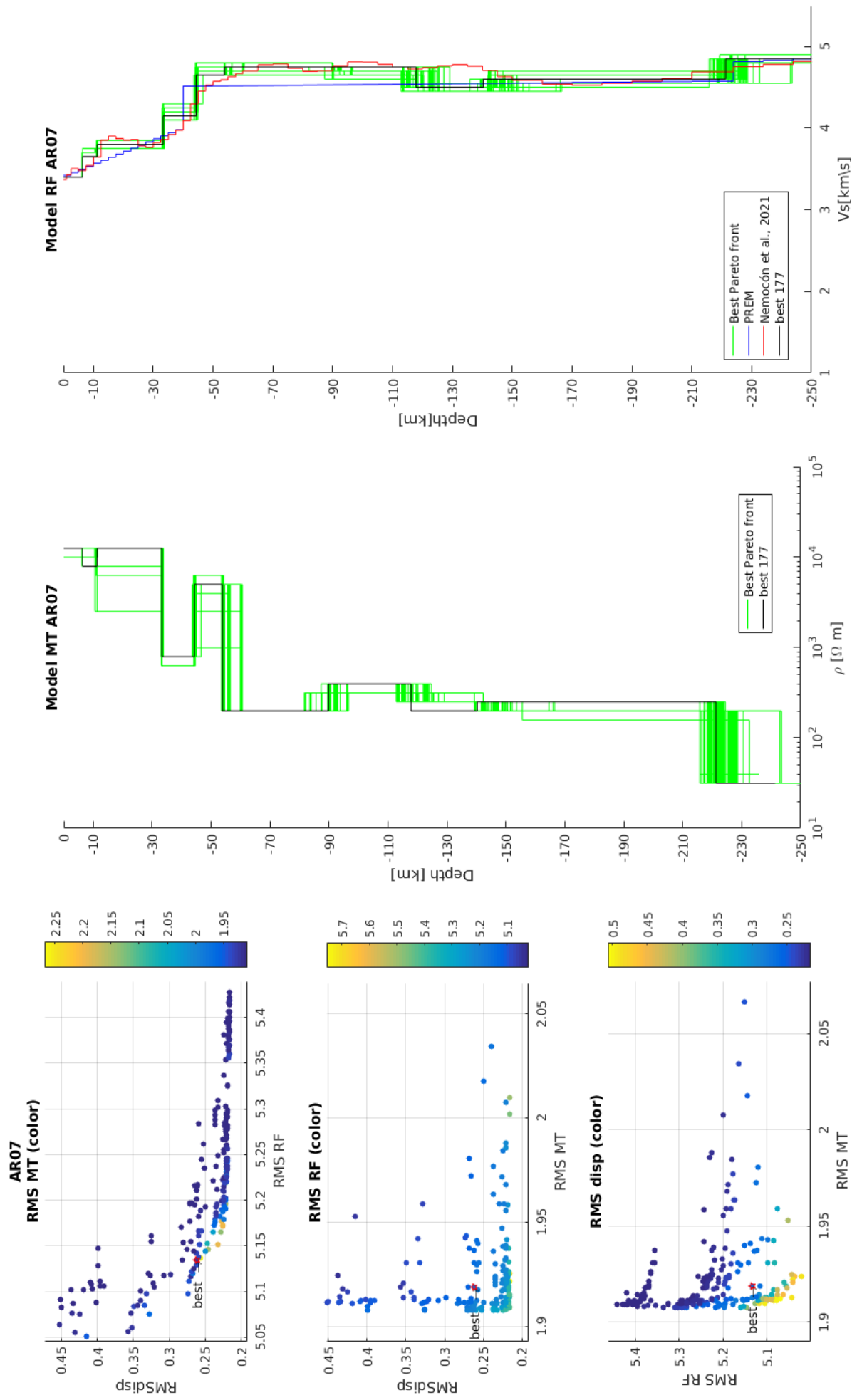


Figure 17: Station AR07.

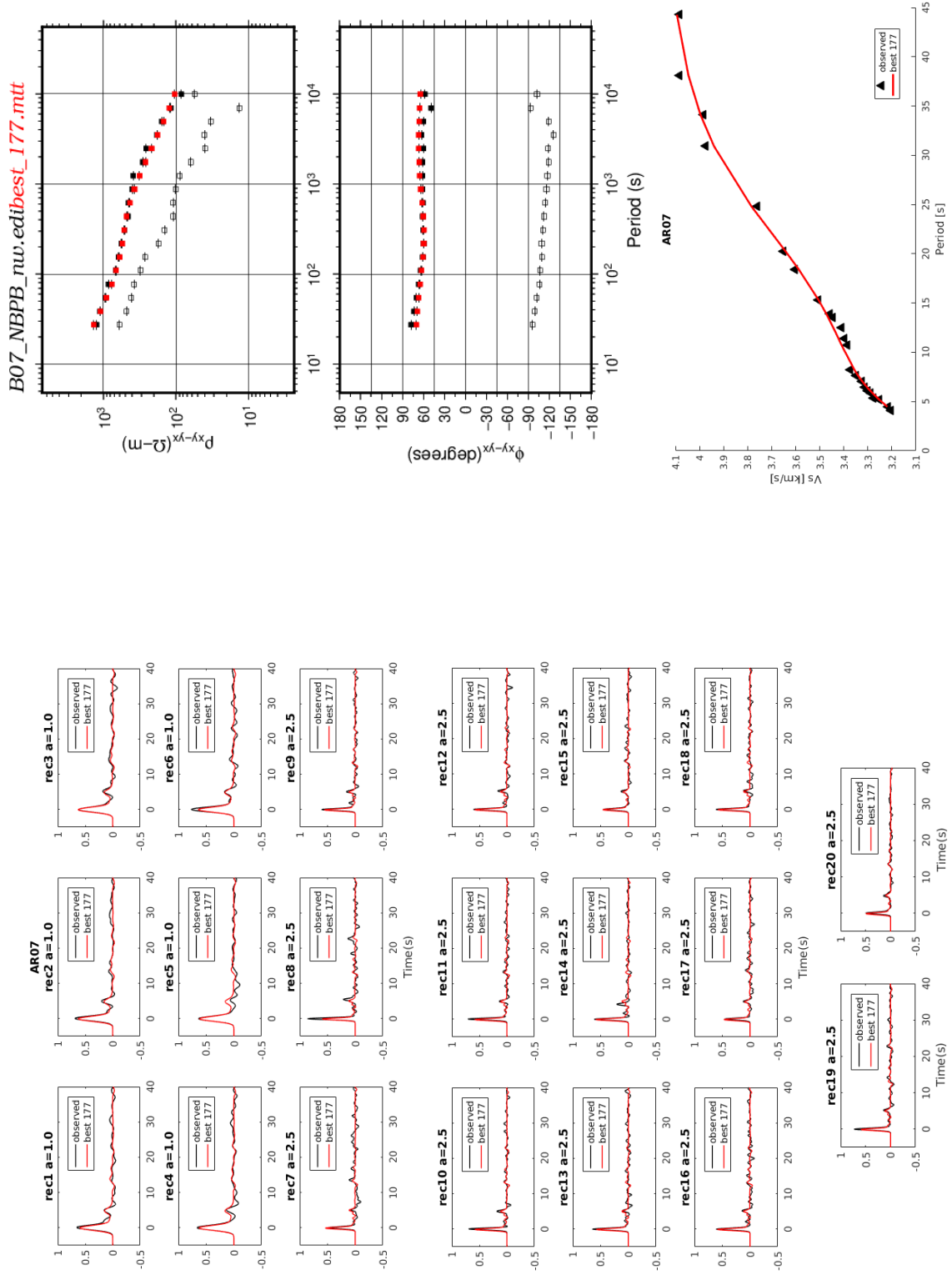


Figure 18: Station AR07.

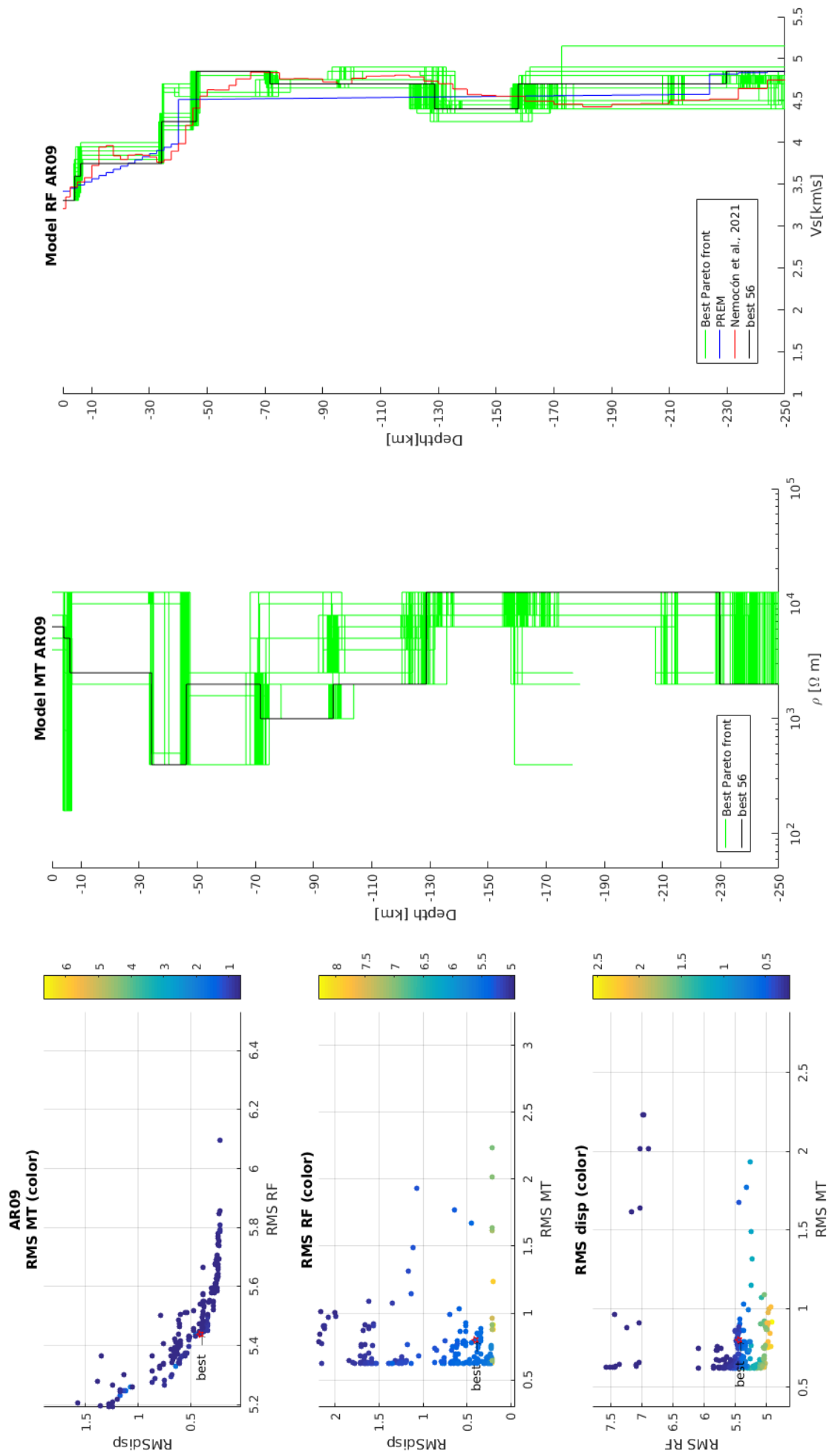


Figure 19: Station AR09.

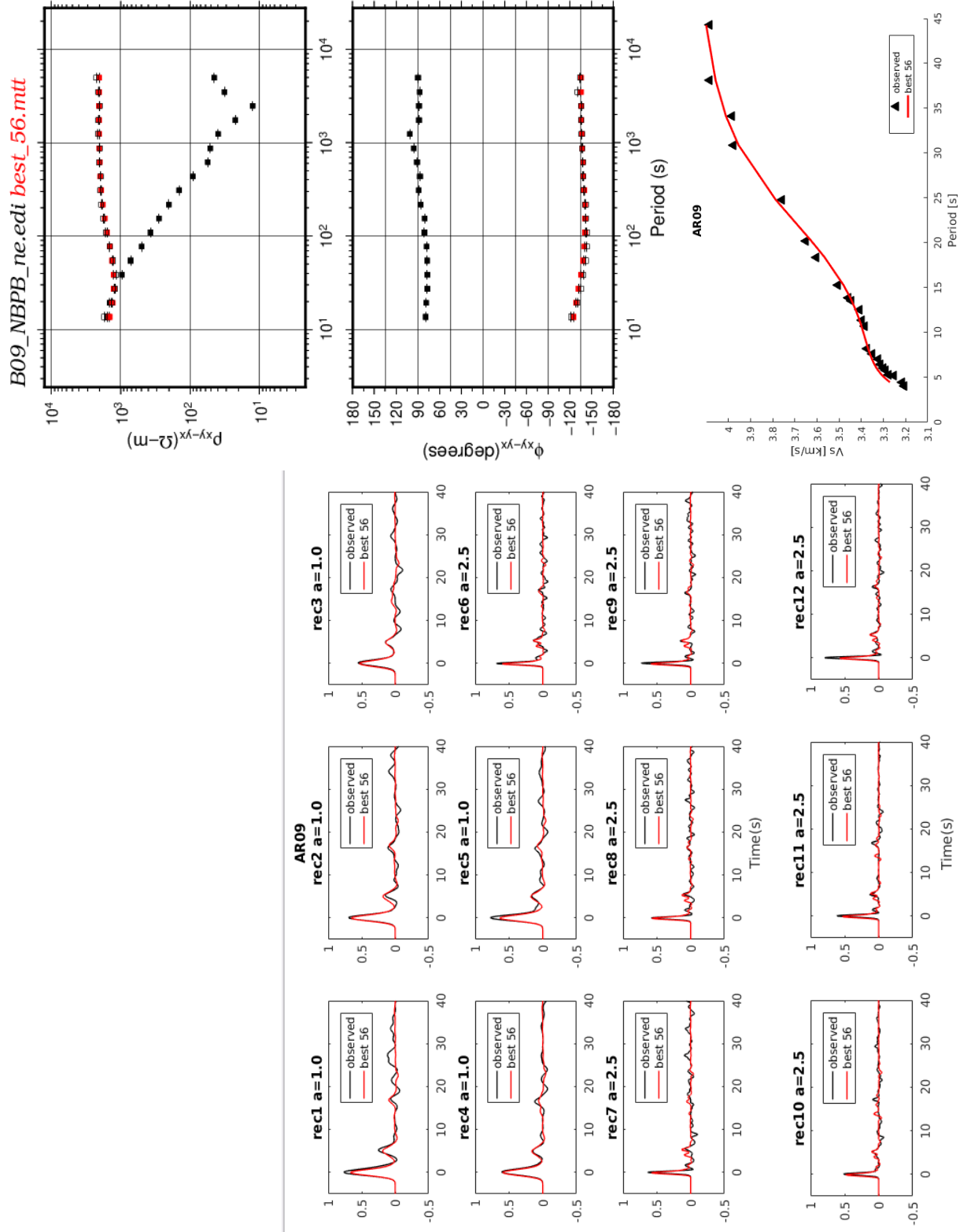


Figure 20: Station AR09.

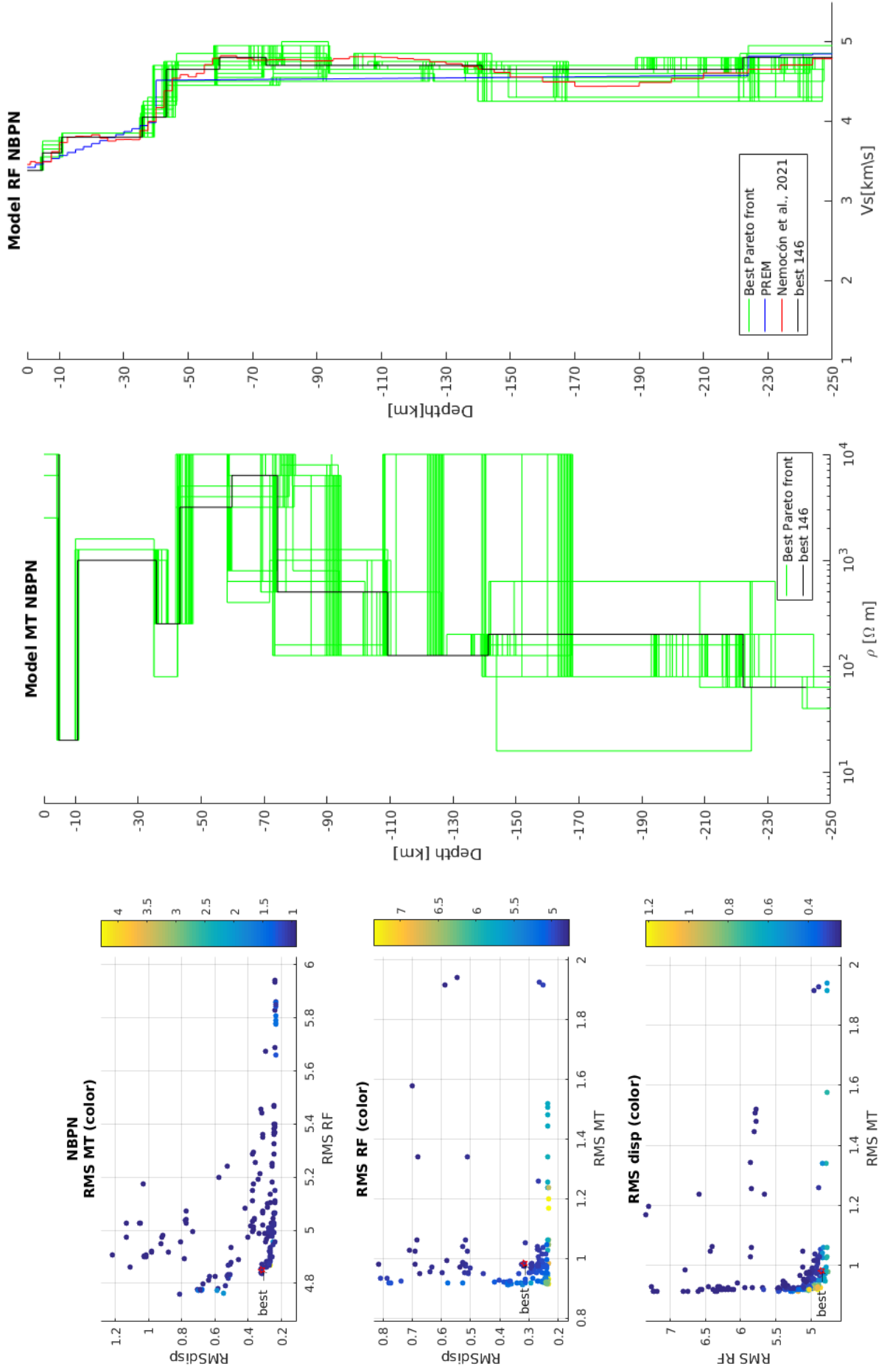


Figure 21: Station NBP.

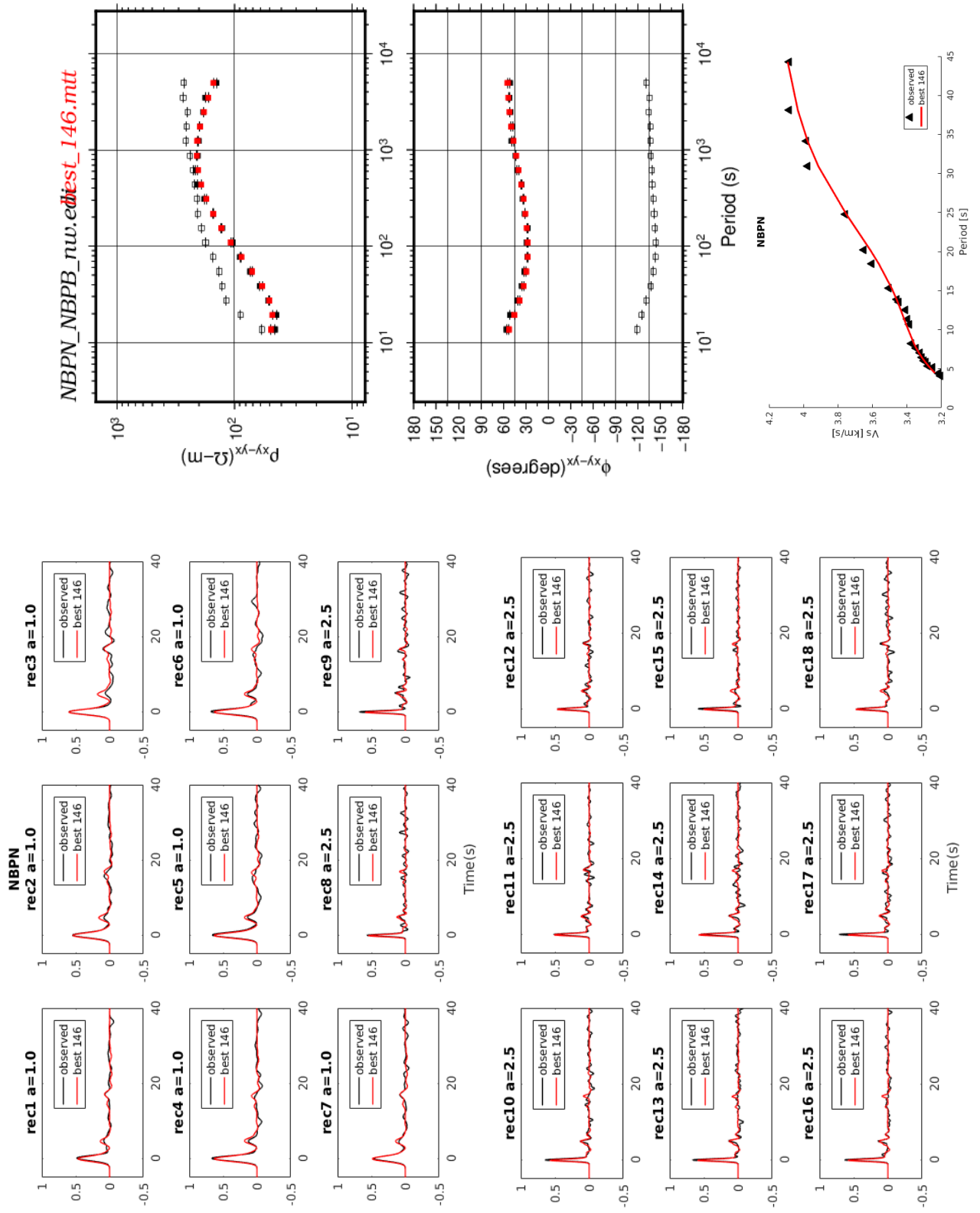


Figure 22: Station NBPB.