Multigenetic Origin of the X-discontinuity Below Continents: Insights from African Receiver Functions

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Key Points:

- P-to-S converted receiver functions reveal the X-discontinuity beneath the East African Rift System, Morocco, Cameroon, Hoggar and several ocean islands
- Observations are collocated with recent surface magmatism suggesting widely distributed chemical heterogeneity below Africa
- No relationships exist between depth and amplitudes of observed X and estimated temperatures, suggesting multiple causal mechanisms

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Abstract

Constraints on chemical heterogeneities in the upper mantle may be derived from studying the seismically observable impedance contrasts that they produce. Away from subduction zones, several causal mechanisms are possible to explain the intermittently observed X-discontinuity (X) at 230–350 km depth: the coesite-stishovite phase transition, the orthorhombic enstatite to high-pressure clinoenstatite phase transition and/or carbonated silicate melting, all of which require a local enrichment of basalt. Africa is host to a broad range of terranes, from Precambrian cores to Cenozoic hotspots with or without lowermost mantle origins. With the absence of subduction below the margins of the African plate for >0.5 Ga, Africa presents an ideal study locale to explore the origins of the X.

Traditional approaches used to map the spatial distribution of horizontal discontinuities observed by receiver functions (RFs), like common conversion-point stacking, ignore the slowness information crucial for discriminating converted upper mantle phases from surface bouncing multiples. By manually assessing depth and slowness stacks for overlapping bins of 1° radius, a normalized vote mapping approach of RF stacks is used to robustly assess the spatial distribution of converted upper mantle phases. The X is mapped beneath Africa between 233 and 340 km depth, revealing patches of heterogeneity proximal to mantle upwellings in Afar, Canaries, Cape Verde, East Africa, Hoggar, and Réunion with further observations beneath Cameroon, Madagascar, and Morocco. There is a lack of an X beneath the whole of southern Africa, and strikingly, the magmatic eastern rift branch of the southern East African Rift. With no relationships existing between depth and amplitudes of observed X and estimated mantle temperatures, multiple causal mechanisms are required across a range of continental geodynamic settings.

Plain Language Summary

Local variations in the mineral chemistry of the upper mantle results in sharp changes in velocity and density. Seismic waves that convert from compressional-to-shear wave propagation (P-to-S) in the upper mantle are sensitive to these jumps in velocity and density, revealing variations in mineral chemistry. One such jump in velocity and density, the X-discontinuity, has several proposed explanations and detecting its presence across a range of mantle conditions allows us to test these possible hypotheses.

We search for observations of the X-discontinuity, where P-to-S conversion occurs between 230–350 km depth, beneath the continent of Africa. Our observations are found beneath many regions of surface magmatism across Africa, suggesting that the X-discontinuity has multiple origins.

1 Introduction

1.1 Overview

Mineral phase transitions cause abrupt jumps in local velocity and density structure of the mantle. The seismically observable impedance contrasts resulting from these transitions are termed seismic discontinuities. Due to estimated pressure-temperature dependence of discontinuity depth, the uplift or depression of some global seismic discontinuities can be studied as a thermometer for local mantle structure (e.g., Helffrich, 2000, 2002). On the other hand, non-global, localized seismic discontinuities can map the local enrichment of chemical heterogeneity, such that the impedance contrast of the seismic discontinuity renders it seismically observable. One such non-global discontinuity at 230–350 km depth is the X-discontinuity (X; e.g., Revenaugh & Jordan, 1991; Deuss & Woodhouse, 2004; Schmerr, 2015) that may be used as a tracer for upper mantle chemical heterogeneity, the distribution of which may shed important insights into mantle dynamics and the extent of chemical equilibration.

Prior observations of the X-discontinuity have been concentrated in the oceans (Deuss & Woodhouse, 2002; Schmerr et al., 2013; Schmerr, 2015; Pugh et al., 2021). While observations of the X at subduction zones are thought to track hydration reactions (Revenaugh & Jordan, 1991) and eclogite (Schmerr et al., 2013) in downgoing plates, importantly, observations of the X above mantle plumes may inform us about both the distribution of chemical heterogeneity in the lower mantle and its subsequent transportation to the upper mantle (e.g. C. D. Williams et al., 2019; Pugh et al., 2021). In regions of elevated mantle temperatures, several causal mechanisms have been proposed to explain observations of the X including the coesite-stishovite (Co-St) phase transition (Q. Williams & Revenaugh, 2005), orthorhombic enstatite to high-pressure clinoenstatite pyroxene (OEN-HCEN) phase transition (Woodland, 1998) and the deep formation of partial melts (e.g., Dasgupta et al., 2013). For further discussion of causes of the X-discontinuity away from elevated temperatures see Schmerr (2015), Kemp et al. (2019) and Pugh et al. (2021).

In contrast to the oceans, studies of the X-discontinuity in continental hotspot settings have received comparatively little attention (e.g., Rein et al., 2020), due to sampling deficiencies in reflected phase studies (e.g., Schmerr, 2015) and the distribution of overlying stations required for converted phase studies. Africa presents the perfect study locale to search for the X because it is host to widely distributed Cenozoic magmatism (e.g., Ebinger & Sleep, 1998; Furman et al., 2006; Pik et al., 2006; de Gouveia et al., 2018), with debated thermochemical nature, origin/scale depth (e.g., Simmons et al., 2007; Civiero et al., 2015; Boyce et al., 2021) and geodynamic causal mechanism (Fairhead & Binks, 1991; Ebinger & Sleep, 1998; King & Anderson, 1998; King & Ritsema, 2000; Gallacher & Bastow, 2012; Milelli et al., 2012). For example, Boyce et al. (2021) propose that a hot, chemically distinct upwelling beneath the southern East African Rift System (EARS; Figure 1b) is sourced from the African LLVP, while magmatism in Ethiopia may lie above an additional purely thermal upwelling whose depth extent is uncertain. Elsewhere, smaller volume magmatism along the Cameroon Volcanic Line (CVL; Figure 1b) in west Africa lacks the age progression associated with a classic mantle plume (Montigny et al., 2004), and has been explained instead by several alternative mechanisms: shear zone reactivation, small-scale convection, and lithospheric delamination (e.g., Fairhead & Binks, 1991; King & Ritsema, 2000; Reusch et al., 2010, 2011; Gallacher & Bastow, 2012; Milelli et al., 2012; De Plaen et al., 2014).

Previous observations of the X beneath Africa have been limited to a few widely scattered SS precursor observations (Deuss & Woodhouse, 2002) and localized scatterers in regional P'P'$_{df}$ precursors studies (Xu et al., 1998) and P-to-S receiver functions (RFs) (Owens et al., 2000; Rein et al., 2020; Pugh et al., 2021). Such point observations do not span competing models of hotspot tectonism, limiting their ability to interrogate the causal mechanisms of the X in the upper mantle. A detailed, continent-wide study of the X is required to understand its exact geographical distribution and the mechanisms that lead to observation. Using seismic network data that sit atop sites of African mantle upwellings, Cenozoic magmatism, and cratonic lithosphere, we map the presence of the X-discontinuity beneath the African continent using receiver functions.

1.2 African Cenozoic Magmatism

Cenozoic magmatism across the African continent has been linked to multiple mantle upwellings of varying scale, geodynamic causal mechanism and thermochemical na-

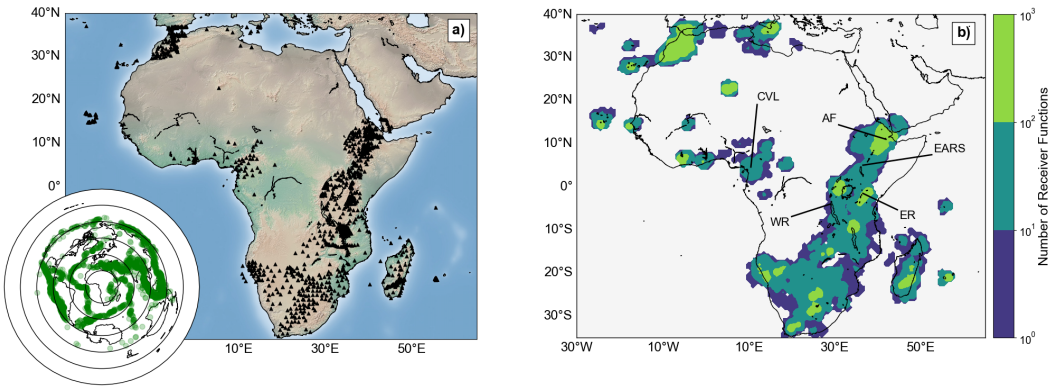

Figure 1. a) Station distribution (black triangles) across Africa and its surrounding islands. The inset globe shows the earthquake distribution (green circles) for this study and black circles represent distance intervals of 30° at 30–180° epicentral distance from the centre of Africa. b) Receiver function distribution across Africa for piercepoints of P300s at 300 km depth using raypaths through PREM (Dziewonski & Anderson, 1981) calculated using the TauP toolkit (Crotwell et al., 1999). AF: Afar, CVL: Cameroon Volcanic Line, EARS: East African Rift System, ER: Eastern Rift, WR: Western Rift.

ture. Uplifted plateaux in northeast, central and southern Africa (Lithgow-Bertelloni & Silver, 1998) and the 45 Ma Ethiopian flood basalts (e.g., Furman et al., 2006; Rooney, 2017) have been associated with ongoing rift-related magmatism along the EARS. One or more mantle upwellings have been invoked to explain uplift, elevated $^{3}He/^{4}He$ anomalies, the surface distribution of magmatism, and the observed mantle transition zone structure (Ebinger & Sleep, 1998; Furman et al., 2006; Pik et al., 2006; Rooney, 2017; de Gouveia et al., 2018; Chang et al., 2020; Boyce & Cottaar, 2021). Global tomographic models reveal a broad, low-wavespeed anomaly extending from the core-mantle boundary beneath southern Africa to the surface below the EARS (Ritsema et al., 1999, 2010; Chang et al., 2015; French & Romanowicz, 2015). Some continental to regional scale models decompose this broad anomaly into multiple smaller low-wavespeed anomalies in East Africa linked to mantle plume activity (Emry et al., 2019; Chang et al., 2020; Boyce et al., 2021). However, the exact number of upwellings is debated with up to three being proposed to explain surface magmatism (Chang et al., 2020) and further suggestions that these could comprise multiple smaller-scale upwellings in the upper mantle (Furman et al., 2006; I. Bastow et al., 2008; Civiero et al., 2015, 2016).

The linear trend of Cenozoic magmatism across Morocco has previously been attributed to edge-driven convection (King & Anderson, 1998; King & Ritsema, 2000; Missenard & Cadoux, 2011; Kaislaniemi & van Hunen, 2014), delamination of the root of the Atlas mountains (Bezada et al., 2014) and diversion of the Canarian mantle plume (Duggen et al., 2009; Mériaux et al., 2015; Miller et al., 2015). Whilst several tomographic models suggest the Canaries sit atop a whole mantle plume (French & Romanowicz, 2015; Marignier et al., 2020), geochemical evidence suggests that Moroccan and Canarian magmatism do not share a single deep origin and are attributable to several upper mantle upwellings (Lustrino & Wilson, 2007; van den Bogaard, 2013). However, recent tomographic models show these upper mantle upwellings are connected at depth (Civiero et al., 2018), with upwelling beneath the Azores and Cape Verde also connected to the same common deep source (Saki et al., 2015).

The CVL in west Africa is a linear chain of volcanoes oriented NE/SW including four islands offshore in the Gulf of Guinea (Fitton, 1980; Déruelle et al., 1991). Despite being observed in at least one global tomographic model (e.g., French & Romanowicz, 2015), the robustness of a whole mantle plume beneath Cameroon in tomographic models is debated (Emry et al., 2019; Marignier et al., 2020; Boyce et al., 2021). With no age progression to magmatism along the CVL (Montigny et al., 2004), several alternative mechanisms have been invoked to explain magmatism including edge-driven convection (King & Anderson, 1998; King & Ritsema, 2000; Reusch et al., 2010, 2011), lithospheric delamination and/or fault zone reactivation (e.g., Milelli et al., 2012; De Plaen et al., 2014; Fairhead & Binks, 1991; Gallacher & Bastow, 2012), and lateral flow of plume material from East Africa (Ebinger & Sleep, 1998).

In central Madagascar, magmatism has been linked with uplift, lithospheric thinning and intercontinental extension (Melluso et al., 2016; Cucciniello et al., 2017). An alternative view connects central Madagascan magmatism to lateral flow of plume material from East Africa as suggested beneath Comoros and northern Madagascar (Ebinger & Sleep, 1998). Recent seismological studies find a thin mantle transition zone (MTZ) beneath south and central Madagascar (Boyce & Cottaar, 2021) and low wavespeed anomalies extending to the lower mantle (Boyce et al., 2021; Tsekhmistrenko et al., 2021), suggesting the presence of a thermal upwelling in this region. In nearby Réunion, magmatism has been shown to be underlain by a mantle plume from seismic tomography (French

& Romanowicz, 2015; Tsekhmistrenko et al., 2021) and anomalously high $^3He/^4He$ ratios (Graham et al., 1990).

2 Data and Method

2.1 Data

We extend the receiver function (RF) data sets of Boyce and Cottaar (2021) and Pugh et al. (2021) using data recorded up until October 2021 downloaded from the Incorporated Research Institutions for Seismology (IRIS) Data Management System for teleseismic earthquakes with magnitude (M_W) $\geq$ 5.5 at epicentral distances of 40-90°. We capitalize on the new TRAILS data set in the Turkana Depression (I. D. Bastow, 2019; Ebinger, 2018; Kounoudis et al., 2021), and a further data set in northeast Uganda (Nyblade, 2017) where there is a paucity of station coverage in our RF data set along the East African Rift (EAR). This results in ∼200,000 event-station pairs recorded between January 1990 and October 2021, recorded at >1,800 stations. The distribution of stations and events is displayed in Figure 1a and a full list of networks used in this study can be found in the Open Research Section and Table S1.

2.2 Receiver Functions

We use P wave RFs to highlight P to S converted phases (Pds; where d denotes the depth of conversion) from the upper mantle. Pds converted from the X are herein referred to as PXs. SV waves converted from an incident P wave are radially polarized. RF analysis (Langston, 1979) constitutes deconvolution of the vertical component seismogram from the radial component assuming that the vertical component represents a convolution of the earthquake source, instrument response and some noise. Subsequently the RFs can be stacked to emphasize the low-amplitude Pds arrivals. RFs record the discontinuity structure at depth near a seismometer. We use the iterative, time-domain deconvolution method (Ligorria & Ammon, 1999) to construct RFs, which iteratively adds Gaussian pulses to reduce the least squares misfit between the predicted and observed radial seismograms. Traces are windowed 25 s before and 150 s after the P-wave to include depth phases in the source deconvolution and remove correlated noise between the vertical and radial traces. Data are bandpass filtered with corner frequencies 0.01 and 0.4 Hz, isolating the frequency band with the largest X amplitudes (Pugh et al., 2021).

Approximately 10% of RFs remain after automatic quality control (see Pugh et al., 2021, for details), leaving 20,630 RFs. Of these, 18,017 remain after further manual inspection to remove obvious low quality RFs (Section S3 and Figure S1a).

RF are initially converted from time-to-depth using the 1D velocity model PREM (Dziewonski & Anderson, 1981). However, 1D time-to-depth conversions can result in >20 km of error on upper mantle discontinuities (Pugh et al., 2021). To perform 3D velocity corrections SEMUCB_WM1 (French & Romanowicz, 2014), and a recent model of the African continent, AF2019 (Celli, Lebedev, Schaeffer, & Gaina, 2020), are used. While tomographic models suffer from sparse and uneven data coverage beneath Africa, the resolution should be greatest where receiver functions are located. This condition is valid especially for the recent model of Celli, Lebedev, Schaeffer, and Gaina (2020), which incorporates body waves from recently available seismic arrays.

2.3 Stacking and Vote Mapping

The amplitudes of Pds on individual RFs are <10% of the incoming P wave and will typically be lower than the noise on an individual trace. As such, high-quality RFs with overlapping sensitivity in upper mantle are stacked to amplify coherent Pds and suppress noise. Following the spiral distribution of Rakhmanov et al. (1994), equidistant bins of radius 111 km ($\sim 1^\circ$ at the equator) are defined across the globe overlapping by $\sim 0.5^\circ$. All RFs that traverse the upper mantle within a bin, estimated using their ray theoretical pierce point at 300 km depth from the TauP toolkit (Crotwell et al., 1999), are stacked together in the depth and time-slowness domains (see Pugh et al., 2021, for details). Visual inspection shows that the majority of stacks with <30 RFs are of poor quality. Therefore, stacks containing >30 RFs are assessed by the visibility of Pds phases in slowness stacks as outlined in Pugh et al. (2021) to determine robust (Figure 2a and b), potential (Figure 2c and d) and null (Figure 2e and f) observations of the X-discontinuity.

Common conversion point stacking (Dueker & Sheehan, 1997) has previously been used to map the 410 km and 660 km discontinuities across Africa, as well as the presence of mid-mantle discontinuities (Thompson et al., 2015; Reed et al., 2016; Boyce & Cottaar, 2021). This technique relies on stacking all depth converted RFs that traverse the upper mantle within a defined Fresnel zone width at each point in a regular grid. Whilst this provides a powerful tool for analyzing these surfaces, slowness information is not con-

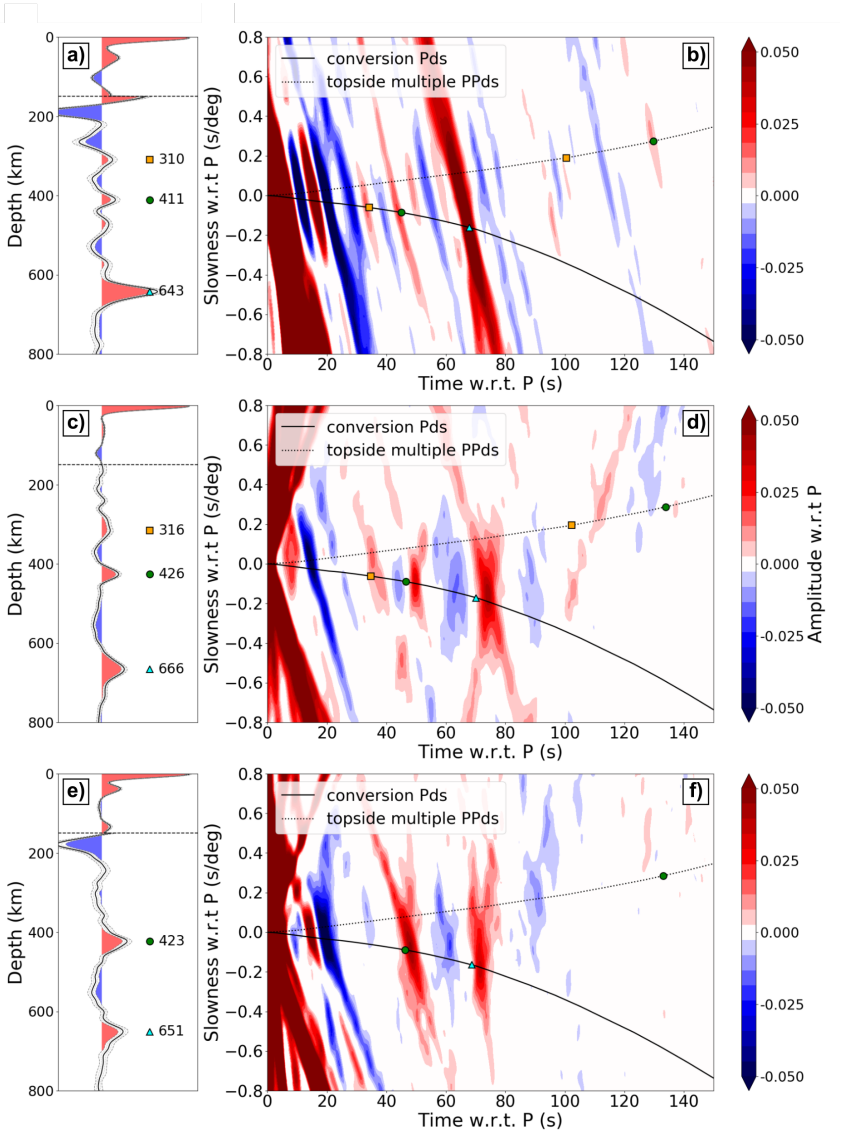

Figure 2. Example stacks for X-discontinuity classifications of Robust (a and b), Potential (c and d) and Null (e and f) respectively. Depth (a, c and e) and slowness stacks (b, d and f) with 130, 370 and 137 RFs, filtered at 0.01–0.4 Hz. Depth stacks (a, c and e): Time-to-depth converted RFs are linearly stacked with the black line marking amplitude (normalized to P) and dashed lines marking 2 σ_M. Amplitudes are multiplied by 5 below the horizontal dashed line at 150 km depth. Stacks are converted from time-to-depth using SEMUCB_WM1. Colored symbols mark significant peaks from PXs (orange squares), P410s (green circles), and P660s (cyan triangles). Slowness stacks (b, d and f): RFs with amplitude >2 σ_M normalized to P stacked in the time-slowness domain. Predicted time-slowness curves are shown for the direct (Pds) and multiple (PPvds) phases. The colored symbols correspond to predicted times and slownesses for direct arrivals and PPvds multiples for significant arrivals in the depth stacks computed from PREM.

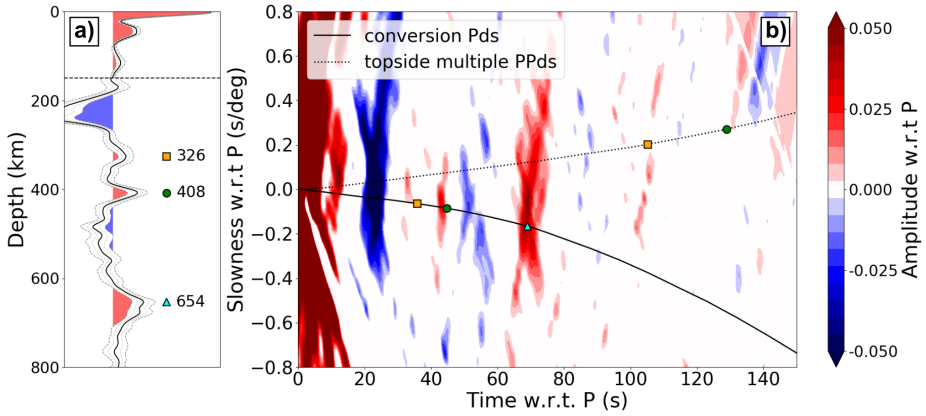

Figure 3. Example of a contaminating PPvds multiple. Depth (a) and slowness stack (b) for 31 RF, filtered at 0.01–0.4 Hz. Depth stack: Time-to-depth converted RFs are linearly stacked with the black line marking amplitude (normalized to P) and dashed lines marking 2 σ_M. Amplitudes are multiplied by 5 below the horizontal dashed line at 150 km depth. The stack is converted from time-to-depth using SEMUCB_WM1. Colored symbols mark significant peaks from PXs (orange squares), P410s (green circles), and P660s (cyan triangles). Slowness stack: RFs with amplitude >2 σ_M normalized to P stacked in the time-slowness domain. Predicted time-slowness curves are shown for the direct (Pds) and multiple (PPvds) phases. The colored symbols correspond to predicted times and slownesses for direct arrivals and PPvds multiples for significant arrivals in the depth stacks computed from PREM.

sidered. With the presence of shallow upper mantle PPvds phases contaminating the depth and slowness interval considered for the X (e.g., Figure 3), common conversion point stacking is deemed unsuitable.

Here, a vote mapping procedure (e.g., Lekic et al., 2012) is carried out whereby each bin that intersects a 0.5°× 0.5° region counts a "vote" towards that region, with robust, potential and null bins voting +1, 0, or -1 to that region respectively. Regions with ≥2 votes in Figure 4 are normalized by the number of votes to account for a heterogeneous data coverage. Regions with <2 votes are masked. We interpret the resulting smooth map as an increasing likelihood that an X is present (positive values) or not present (negative values). Using this approach of a vote map with overlapping bins reduces the dependence of our results on a specific choice of grid. Whilst the depth information of common conversion point stacking is lost, this provides a robust representation of the dis-

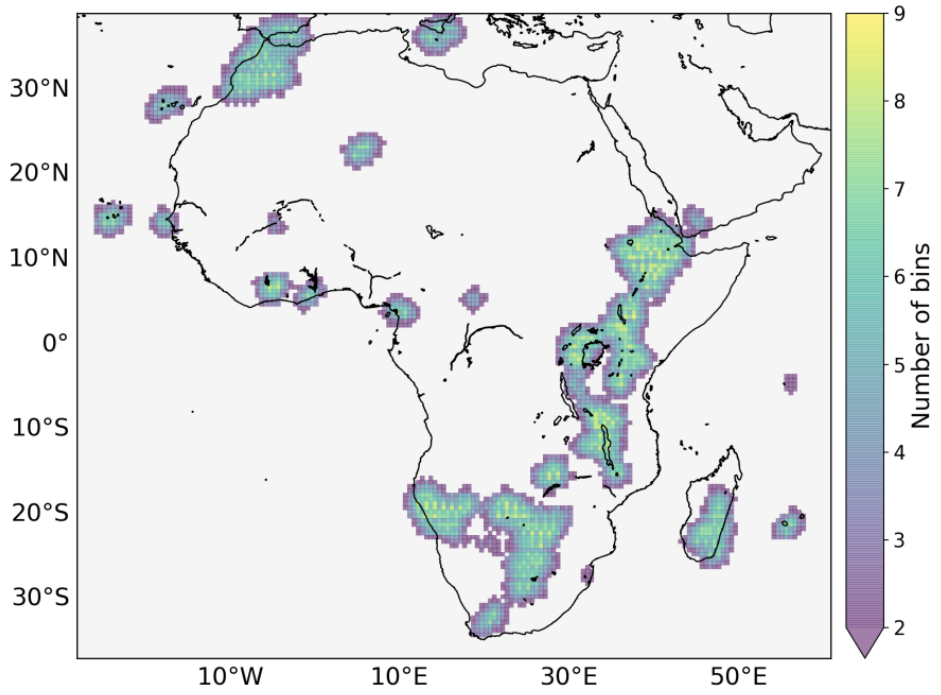

Figure 4. Number of 1° radius bins intersecting each 0.5° × 0.5° bin for normalized vote mapping.

tribution of the X and its potential length scales across the continent including depth and slowness information.

3 Results

3.1 Geographical distribution of the X-discontinuity

The upper mantle beneath the African continent is imaged using >18,000 RFs. Depth and slowness stacks are computed to identify observations of the X for nearly 600 overlapping bins of 1° radius. Using a normalized vote mapping approach in Figure 5, the geographical distribution and length-scales of the X-discontinuity are displayed across the African continent. X observations cluster around the Canaries, Cameroon, Cape Verde, Ethiopia, the Hoggar volcanic province, southernmost Madagascar, Morocco, and Reunion with intermittent observations appearing along the Western Rift (WR). Null observations are prevalent in southern Africa, along the Eastern Rift (ER), and in north-

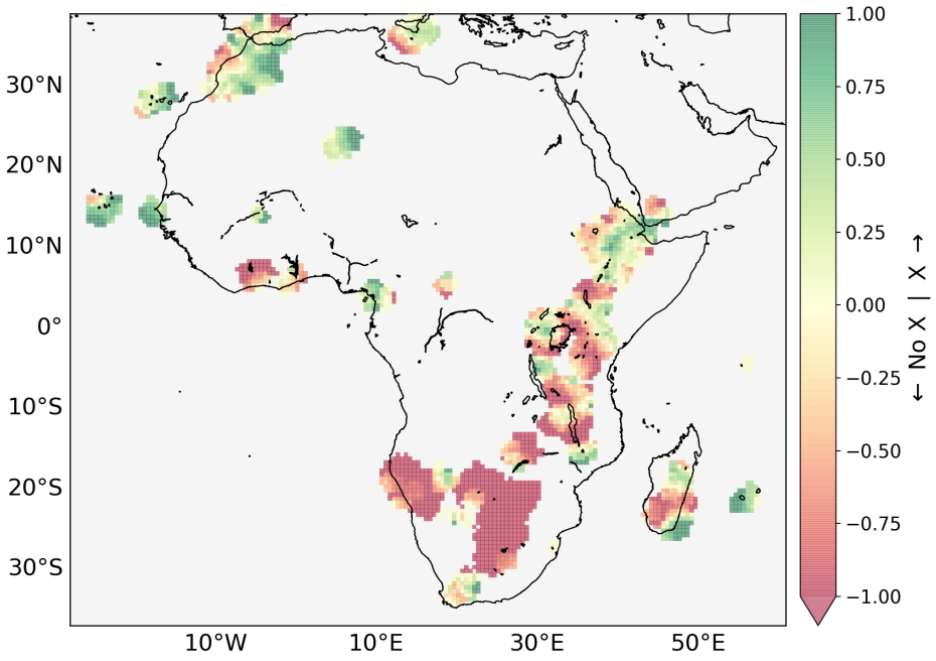

Figure 5. Normalized vote map of X-discontinuity observations for 597 overlapping 1° radius bins with ≥2 votes on a 0.5°× 0.5° grid with darkening green colours showing an increasing likelihood of the X and darkening red colours showing an increasing likelihood of no X. Votes are based upon the presence of the X-discontinuity in slowness stacks.

ern Madagascar. This distribution is explored further in the context of magmatism in Section 4.2.

The number of RFs in a stack has little bearing on its classification (Figure S2). However, the backazimuth distribution of RFs has a strong control on epicentral distance distribution, thus determining the streakiness of slowness stacks and potentially masking X observations (Figure S4; Section S4).

3.2 Depth distribution of the X-discontinuity

Nearly 600 depth and slowness stacks with > 30 RFs are visually inspected for the presence of the X-discontinuity. Upper mantle positive conversions are present in the depth range of 212–377 km, 233–340 km, and 235–347 km depth for 262 stacks using PREM, SEMUCB_WM1 and AF2019, respectively. Using the criteria set out in Pugh et al. (2021), 172 stacks are classed as robust with the correct slowness for PXs, 121 stacks are classed

as potential, 303 stacks are classed as null and only one stack is classed as poor quality; 59 stacks contain robust observations of PPvXs.

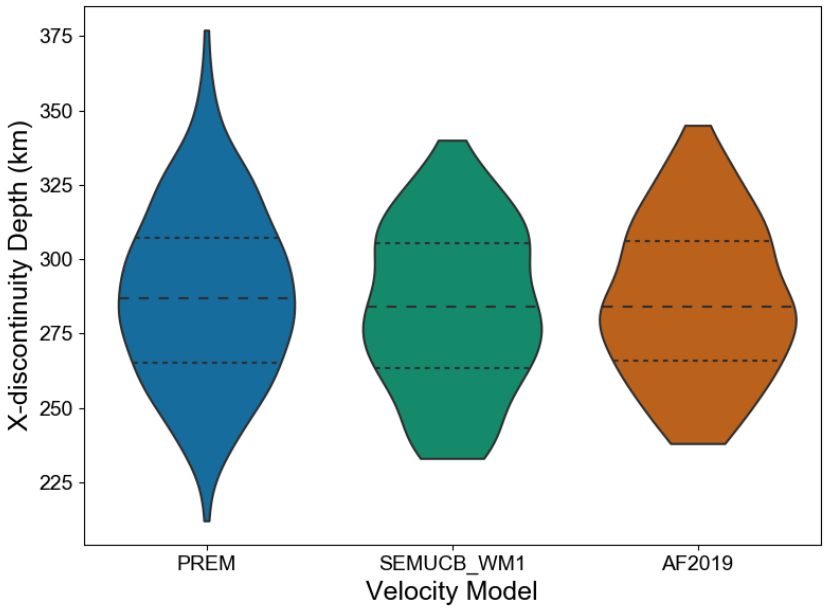

Figure 6. Violin plots of X-discontinuity depths observed in depth stacks cut to the minimum and maximum observations for the three velocity models used in this study: PREM, SEMUCB_WM1 and AF2019. Dashed lines mark the 1^{st}, 2^{nd} and 3^{rd} quartiles of each violin.

Including slowness information, robust X-discontinuity depths range between 233 and 340 km, centered around a mean depth of 284 km with lower and upper quartiles of 264 and 305 km respectively for SEMUCB_WM1 (Figure 6). Though the total data distribution shifts deeper when using AF2019 (238-345 km depth), the mean depth and quartiles (284 km, 266 km and 306 km) are remarkably similar to those from SEMUCB_WM1.

SEMUCB_WM1 is preferred for depth correction as temperature variation and topography calculated in Section S5 suggest it better accounts for upper mantle velocity structure. Subsequently, depths reported below are as converted using SEMUCB_WM1 and displayed in Figure 7.

Only 34% of the depth and slowness stacks that contain PXs also contain PPvXs arrivals. It is logical that the percentage of stacks containing PPvXs is lower beneath the continents than the 60% percent in Pugh et al. (2021), where stacks are predominantly made beneath ocean islands. The heterogeneous nature of the continental crust

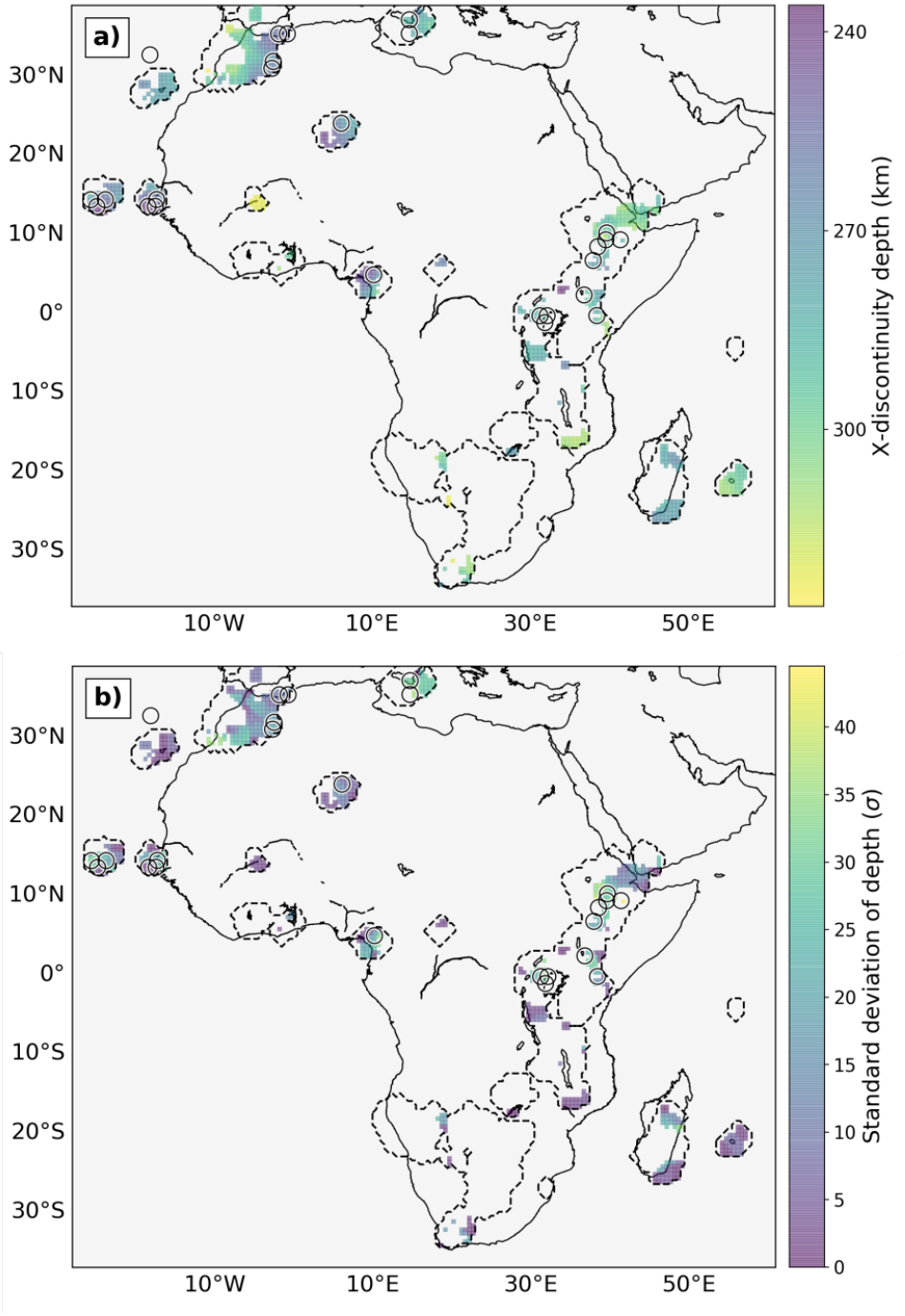

Figure 7. Vote maps of a) X depth and b) standard deviation of depths (σ) for 597 overlapping 1° radius bins with ≥2 votes, and an average vote ≥ 0.25 (Figure 5), on a 0.5° × 0.5° grid. Black circles mark bins where two X observations are made and black dashed contours mark the enter of the vote map (Figures 4 and 5).

and lithosphere may inhibit coherent stacking of PPvXs, especially considering the greater/variable thickness of the continental lithosphere.

4 Discussion

4.1 Potential trends in the X-discontinuity

We explored discontinuities in the upper mantle beneath the African continent and surrounding ocean islands in ∼600 equally spaced depth and slowness stacks (Figure 8). Scattered observations of the X exist in nearly 30% of the stacks over a broad depth range of 233–340 km. Figure 8a shows that the depth and amplitude distribution of these results is comparable to those in Pugh et al. (2021) with the largest amplitudes of ≥8% of the main P wave arrival found at ∼280 km depth and smaller amplitudes of ≤3% below 330 km depth. No clear correlation exists between the depth of the X and its amplitude (Figure 8a). We calculate average upper mantle temperatures at 200–400 km depth for every stack at ±1° latitude and longitude using the temperature deviations found in the geophysical-petrological inversion of Fullea et al. (2021). No correlation can be found between the depth of the observed X and these local thermal perturbations (Figure 8b) and the X is no more readily observed at elevated upper mantle temperatures than at depressed temperatures (Figure S6). However, null observations of the X are more readily found at depressed mantle temperatures (Figure S6). Without a trend in the temperature-depth space, it is impossible to infer a Clapeyron slope for the X. With no Clapeyron slope, and such a broad range in depths and amplitudes, we conclude multiple causal mechanisms are responsible for the X below continents.

X discontinuity observations are broadly found beneath the thinnest lithosphere across the African continent (Figure S7) using three global tomographic models (Schaeffer & Lebedev, 2013; Debayle et al., 2016; Priestley et al., 2018). X observations closely track the thin lithosphere of the Main Ethiopian Rift and are present beneath several ocean islands. However, the X is absent beneath thick cratonic lithosphere. One explanation for the lack of X observations below thick lithosphere concerns multiples. PPvds multiples are expected to arrive at seismic stations at the same time as PXs conversions when d ≈ 80–100 km, similar to the depth of the mid-lithospheric discontinuities reported in S wave RF studies beneath the Tanzanian (Wölbern et al., 2012) and Kalahari (Sodoudi et al., 2013) cratons, though these discontinuities are observed with the opposite polar-

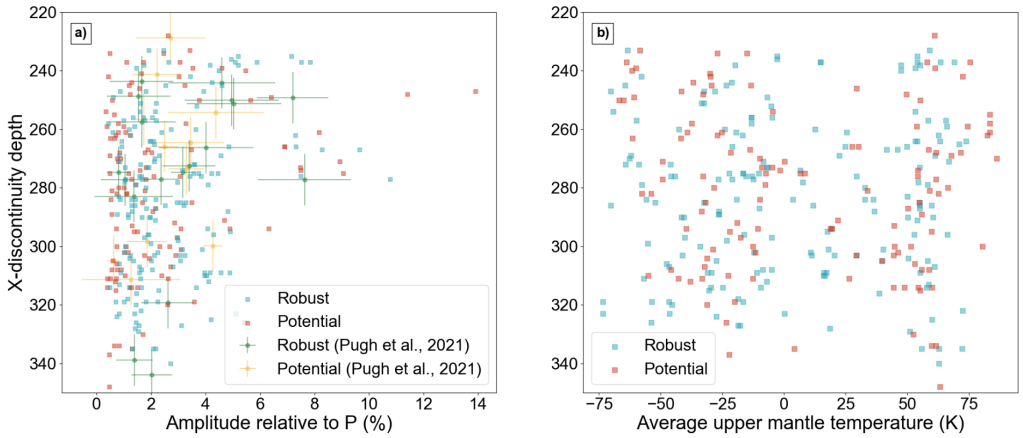

Figure 8. a) Depth and amplitude distribution of 172 robust X-discontinuity observations, and 121 potential X-discontinuity observations plotted alongside the observations of Pugh et al. (2021). Depths are converted using SEMUCB_WM1. Error bars are calculated in depth and amplitude using 10 jackknife resamples with 90% of the data in each sample. Amplitude error bars represent $\pm 2\ \sigma_M$ of the mean of each stack whereas depth error bars represent the mean width of the PXs $+2\ \sigma_M$ peak at the PXs amplitude across all stacks. Error bars are not displayed for observations from this study to avoid overcrowding the axes. b) Depth and average upper mantle temperature distribution of 172 robust X-discontinuity observations, and 121 potential X-discontinuity observations. Temperatures deviations are taken from Fullea et al. (2021) and averaged at 200–400 km depth.

ity to the X. As for the lithosphere-asthenosphere boundary, only 13–21% of 0.5° × 0.5° bins in our vote map sample lithosphere between 80 and 100 km thick according to maps derived by Hoggard et al. (2020) using velocity models SLNAAFSA (Schaeffer & Lebedev, 2013, 2014; Celli, Lebedev, Schaeffer, & Gaina, 2020; Celli, Lebedev, Schaeffer, Ravenna, & Gaina, 2020), CAM2016 (Ho et al., 2016; Priestley et al., 2018) and 3D2015-sv (Debayle et al., 2016). As such, interfering phases are unlikely to be a persistent issue for RF stacks. While thick heterogeneous lithosphere could result in the incoherent stacking of RFs, very low standard error in null stacks beneath cratons suggests that, for the thickest lithosphere, this does not occur.

In line with our correlation between null observations and depressed upper mantle temperatures, we suggest a cooler asthenospheric mantle underlying thicker lithosphere may present unfavorable conditions of X formation/visibility. This is, however, challeng-

ing to prove with our results alone. For stacks classed as potential, limited backazimuth and epicentral distance distribution resulting in streaky slowness stacks seems to have the largest control on the X observation robustness (Section S4).

While no quantitative relationship is found between the depth of the observed X and estimated temperatures in the upper mantle, most regions of Quaternary-recent magmatism are associated with an X observation: they are located proximally to several ocean islands (Canaries, Cape Verde, Réunion), Morocco, Cameroon, the East Africa Rift and Madagascar, and overlap with our previous RF stacks (Pugh et al., 2021). While robust X observations were found in the Canaries and Cameroon in Pugh et al. (2021), this current study highlights the importance of studying the X over short wavelengths with potential observations in Cape Verde, Hoggar, Afar and Réunion now found to be robust. The non-robustness of X signals in these regions in Pugh et al. (2021) may have been the result of unimodal backazimuth and epicentral distance distributions causing streaky slowness stacks (e.g., Figure S4) or topography across the X on short wavelengths as can be seen for Cape Verde, Hoggar and Réunion (Figure 7). These locations all host Quaternary volcanoes and/or Cenozoic magmatism (Figure 9). Alongside X observations in regions of ongoing subduction (e.g., Revenaugh & Jordan, 1991; Schmerr et al., 2013), these observations suggest the X to be related to recent upwelling or downwelling, with chemical heterogeneity mixed into the mantle during subsequent mantle convection. The causal mechanisms and implications of these observations are discussed below to explore the cause of the X-discontinuity across upwellings of variable geodynamic origin.

4.2 Links to Surface and Geodynamic Features

4.2.1 East Africa

The presence of the X beneath East Africa may be related to chemical heterogeneity introduced by mantle upwellings in the region (Simmons et al., 2007; Rooney, 2017; Boyce & Cottaar, 2021). Robust X observations underlie several sections of the EARS from Afar in the north, through Ethiopia and two patches beneath the WR, to a small patch in Mozambique at depths of 270-320 km (Figure 10a). Notable null results are seen beneath the Turkana Depression, along the ER, and under the Tanzanian craton where substantial volumes of data exist. In Ethiopia, X depths gradually increase northwards to Afar (Figure 11). If the X is controlled by a common causal mechanism beneath the

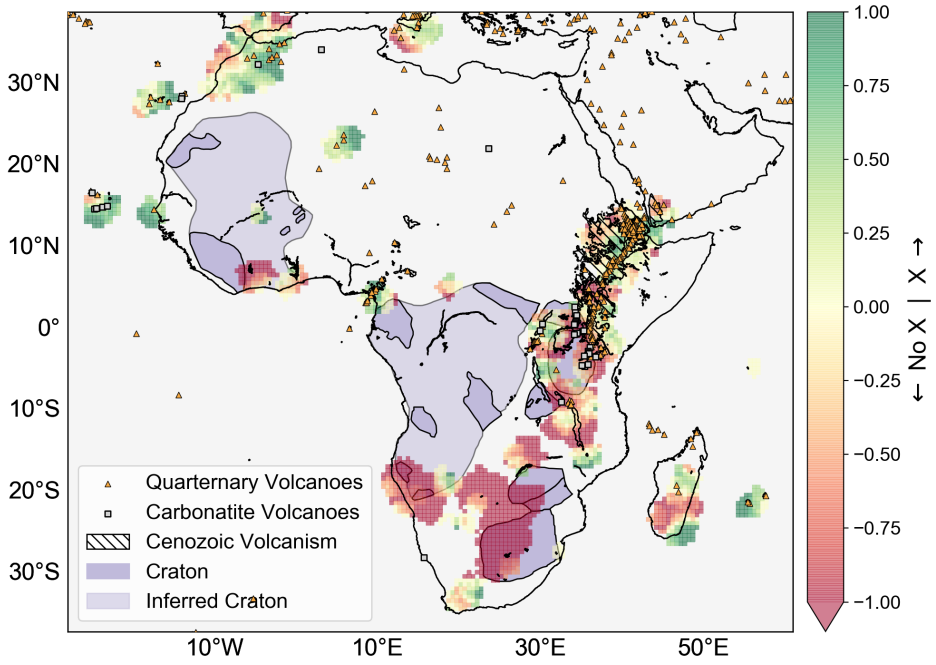

Figure 9. X-discontinuity vote map (Figure 5) plotted above cratons and Cenozoic magmatism adapted from (Begg et al., 2009; Boyce et al., 2021; Kounoudis et al., 2021). Carbonatite volcanoes plotted are ∼45 Ma-recent (Woolley & Kjarsgaard, 2008; Muirhead et al., 2020).

EARS, the positive Clapeyron slope of the Co-St (Akaogi et al., 2011) would suggest an increase in temperature towards Afar, consistent with reductions in seismic wavespeed (Boyce et al., 2021). However, the impact of the variable presence of melt on seismic wavespeeds makes isolating thermal controls on seismic heterogeneity difficult to isolate (e.g., Rooney et al., 2012; I. Bastow et al., 2008).

There is significant debate as to the number of plumes that exist in the upper mantle in East Africa and whether they are commonly or uniquely sourced. One to three whole-mantle plumes of variable thermochemical nature or multiple upper mantle plume heads have been proposed to explain surface magmatism on the strength of seismological and geochemical evidence (Ebinger & Sleep, 1998; Furman et al., 2006; Pik et al., 2006; Civiero et al., 2015, 2016; de Gouveia et al., 2018; Chang et al., 2020; Boyce & Cottaar, 2021). Robust X observations closely follow the Main Ethiopian Rift from Afar to southern Ethiopia, and reappear along parts of the western rift marked by Quaternary volcanism (Figures

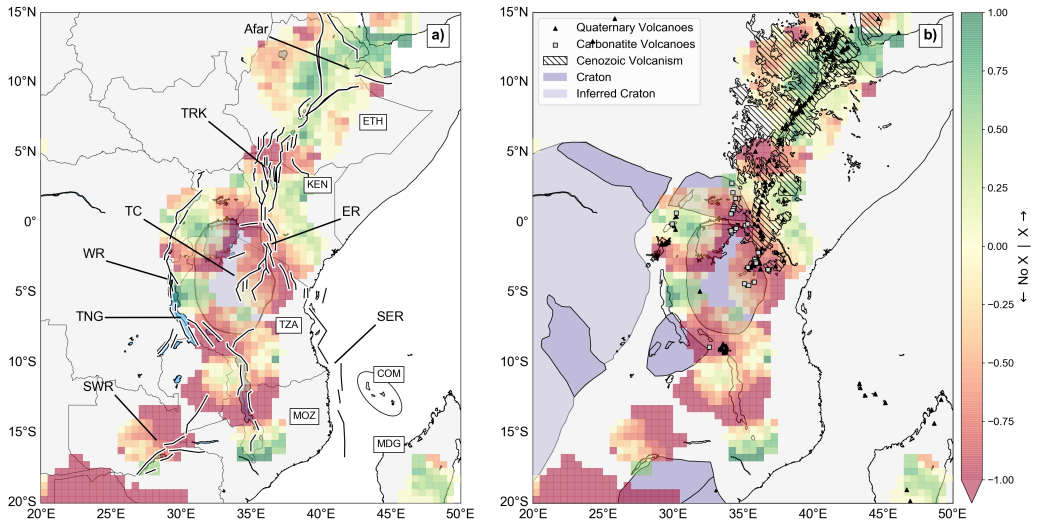

Figure 10. a) X-discontinuity vote map (Figure 5) plotted above the Tanzanian Craton (TC - purple) and lakes (blue) along the EARS (TRK, Lake Turkana; TNG, Lake Tanganyika). Major faults adapted from Jones (2020) are marked with thick black and white lines and reveal the Eastern, Western, Southeastern and Southwestern rifts (ER, WR, SER, SWR). National borders are marked with thin black lines and countries referred to in text are labelled (COM, Comoros Archipelago including Mayotte; ETH, Ethiopia; KEN, Kenya; MDG, Madagascar; MOZ, Mozambique; TZA, Tanzania). b) X-discontinuity vote map (Figure 5) plotted above cratons and Cenozoic magmatism adapted from (Begg et al., 2009; Boyce et al., 2021; Kounoudis et al., 2021). Carbonatite volcanoes plotted are ∼45 Ma-recent (Woolley & Kjarsgaard, 2008; Muirhead et al., 2020).

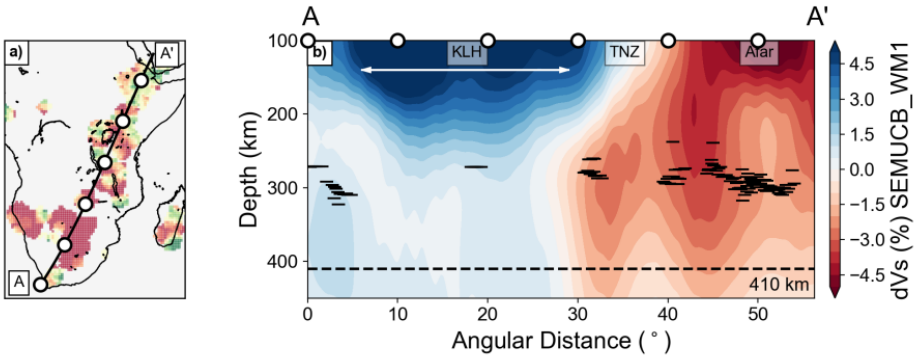

Figure 11. Along profile a), a cross-section b) of X-discontinuity depths of robust observations from South Africa to Afar plotted above the tomographic model SEMUCB_WM1 (French & Romanowicz, 2015). X-discontinuity depths are taken from bins in Figure 7 ≤250 km from the line of section. KLH, Kalahari Craton; TNZ, Tanzanian Craton. The white arrow marks the horizontal extent of the Kalahari Craton.

10). Widely distributed X observations suggest chemical heterogeneity is pervasive throughout East Africa around 300 km depth and so presents no clear support for multiple small scale upwellings that have been reported by some workers (Civiero et al., 2015, 2016) or underlying plumes of variable thermochemical nature (Boyce & Cottaar, 2021), at least at X depths. We therefore broadly support the notion that plume signatures in the East African upper mantle are well mixed and the upper mantle pervasively hosts material transported from depth by the African Superplume (e.g., Rooney, 2017). However, we note that some scatter in X observations in East Africa may be associated with the variable presence of CO_2 assisted silicate melting that is required to explain the discrepancy between mantle potential temperature estimates and slow seismic wavespeeds below depths commonly associated with decompression melting (Rooney et al., 2012). Previous workers show the basal impedance contrast from such a carbonate silicate melt layer presents a viable explanation for the X (Dasgupta et al., 2013).

Moving southwards from Ethiopia, X observations terminate abruptly north of Lake Turkana (Figure 10) forming a WNW–ESE band of null observations that interrupt the broad trend of robust observations below the East African Rift. Given the presence of Quaternary volcanoes and Cenozoic magmatism in much of the Turkana Depression (Figure 9), this result may indicate a locally different geodynamic environment to that ob-

served below the Ethiopian and East African plateaus beneath which two separate plumes have been proposed (e.g., Pik et al., 2006). However, recent tomographic models present no clear evidence for a break in slow wavespeeds (and therefore dynamic support) at the upper mantle depths to which our data are sensitive (Hansen et al., 2012; Emry et al., 2019; Celli, Lebedev, Schaeffer, & Gaina, 2020; Boyce et al., 2021; Kounoudis et al., 2021). Intriguingly, a fast wavespeed band at lithospheric depths in southernmost Ethiopia in the seismic tomographic study of Kounoudis et al. (2021), coincident with a broadly (∼500 km-wide) rifted zone (Figure 10) co-exists with our zone of absent X. This anomalous region, interpreted by Kounoudis et al. (2021) as refractory Proterozoic lithosphere, is not associated with Quaternary volcanism, perhaps resulting in a lack of melt ponding below the region at X depths. Complex lithospheric seismic structure, both associated with the Kounoudis et al. (2021) fast wavespeed band, and with the failed Mesozoic Anza rift immediately to the south of it in the Turkana Depression, may be precluding our view of the X.

South of the Turkana Depression, the EARS splits into the ER, WR, Southwestern and Southeastern rift zones (Figure 10a), which developed in Proterozoic lithosphere between thick cratonic lithosphere (e.g., Chorowicz, 2005; Mulibo & Nyblade, 2016; Ebinger et al., 2017; Daly et al., 2020). Owens et al. (2000) observe Pds arrivals at 250–300 km across Tanzania, but they note that they cannot discriminate between a Pds phase or a shallower multiple from velocity analysis alone. This region overlaps with regions of X observations east of Lake Tanganyika along the WR and null observations due east along the ER. Unlike the ER that has experienced 30 Ma-Recent magmatism along its length, the WR is characterized by isolated, volumetrically small magmatic provinces (e.g., Ebinger, 1989; Chorowicz, 2005; Roberts et al., 2012). The Southwestern rift zone has no known magmatism, whereas the Southeastern rift zone offshore between Africa and Madagascar has experienced ∼20 Ma-recent magmatism (e.g., Michon, 2015; Courgeon et al., 2016, 2017; O'Connor et al., 2019; Berthod et al., 2022). Therefore, it is striking that the ER is underlain by null X observations while patchy X observations underlie the WR (Figure 9a). The two patches along the WR show consistent X depths of ∼290 km, however, at the surface, the northern patch is colocated with carbonatite magmatism while the southern patch underlies no surface magmatism. Furthermore, the role of the Tanzanian Craton, separating the two branches remains enigmatic. Whether through edge-driven convection (King & Anderson, 1998; King & Ritsema, 2000) or lateral di-

version of plume material around the cratonic keel as suggested beneath the Kalahari Craton (Forte et al., 2010; Tepp et al., 2018), it remains uncertain whether the Tanzanian Craton could divert chemically heterogeneous plume material to the WR. However, the Tanzanian Craton has a shallower depth extent than other cratons globally (e.g., Priestley et al., 2018; Celli, Lebedev, Schaeffer, & Gaina, 2020). Consequently, it is unclear whether it would have a similar impact on upper mantle flow compared to cratons of typical thicknesses ($\geq$250 km). Whether craton induced flow would result in our null observation beneath the ER is also elusive.

4.2.2 Canaries

Robust X observations occur at 270–280 km depth beneath the Canaries (Figure 7a) with the most robust results appearing to the north and east (Figure 5). The Canaries have been shown to overlie a whole mantle plume in tomographic models (French & Romanowicz, 2015; Marignier et al., 2020), meaning, similarly to East Africa, the X beneath the Canaries may also result from the introduction of chemical heterogeneity from the deep mantle. Canarian shield stage lavas contain the signatures of both old (>1 Ga; Thirlwall, 1997; Gurenko et al., 2006, 2009) and young (<1 Ga; Widom et al., 1999; Geldmacher & Hoernle, 2000) recycled oceanic crust. Should the Canarian mantle plume recycle oceanic crust to the surface, this may provide necessary chemical heterogeneity to explain the X via a single causal mechanism here, similar to other ocean island hotspots (e.g. Kemp et al., 2019; Pugh et al., 2021).

4.2.3 Morocco

The X is widespread beneath Morocco. Observations of the X at 250–310 km depth (Figure 7a) show the greatest variation in depth over short spatial distances for our study region. Here the X has been observed in several previous studies (e.g. Deuss & Woodhouse, 2002; Rein et al., 2020), with potential links to the Canaries mantle plume (e.g. Rein et al., 2020). We observed the X at depths of ~310 km on the western coastline shallowing monotonically eastwards to ~250 km (Figure 7a). However, to the south of this region, Rein et al. (2020) observe the X deepening ~eastwards from ~310 km to ~350 km. Considering the Clapeyron slope of the Co-St phase transition invoked by Rein et al. (2020), it would be expected to deepen with increasing temperature. Wavespeed anomalies in this region in SEMUCB_WM1 and AF2019 transition from slow in the west, to fast in

the east. Interpreted in terms of temperature, this would suggest a cooling trend west to east, and a shallowing X in line with our observations. Double X observations observed in this region (Deuss & Woodhouse, 2002; Rein et al., 2020) may explain discrepancy with our shallowing results. Although we do not observe two X observations, there is a large standard deviation in X depth between stacks (Figure 7b). Large standard deviation in depth is a reasonable indicator that two X observations may occur, often being colocated with two X observations elsewhere in the African continent (Figure 7), though the secondary arrival may be a multiple.

Duggen et al. (2009) present geochemical analyses to show that the Canaries plume may have deflected to the northeast beneath the Moroccan lithosphere, a suggestion supported by plume modelling (Mériaux et al., 2015). However, several studies (Lustrino & Wilson, 2007; van den Bogaard, 2013) show that the geochemistry of magmatism across the region does not fit with a single origin deep sourced mantle plume and is more readily reconciled by multiple upper mantle upwellings. This remains a topic of ongoing debate with a recent regional tomographic study suggesting that these multiple upwellings may have a common deep source (Civiero et al., 2018). Rein et al. (2020) use evidence of old and young recycled oceanic crust in lava samples from the Canaries (e.g., Thirlwall, 1997; Gurenko et al., 2006, 2009), and the proximity of subducted slabs in the Mediterranean, to conclude that multiple upwellings recycle basalt into the upper mantle, facilitating the Co-St phase transition as the causal mechanism for the X. From our extended data set, it is possible that recycling of basalt may be pervasive across this region, extending offshore of Morocco beneath the Canaries.

4.2.4 Cape Verde

Shallow X observations are found to the south of Cape Verde and offshore Senegal at 240–270 km depth, shallowing approximately north to south (Figure 7a). There are also a number of doubled X observations in this region. X observations beneath Cape Verde may be associated with hotspot magmatism and potentially also linked to the Canaries plume. Whilst French and Romanowicz (2015) classify Cape Verde as overlying a 'primary plume', Marignier et al. (2020) are less confident of a plume in this location compared with the majority of their 'primary plumes'. However, Cape Verde exhibits HIMU and EM geochemical signatures, and high $^{3}He/^{4}He$ ratios (Doucelance et al., 2003; Jackson et al., 2017, 2018), representing plume signatures. Geodynamic models (Davaille

et al., 2005) and precursor studies (Saki et al., 2015) found the Cape Verde plume to have a common source in the lower mantle with the Canarian plume to the north. Should this be the case, recycled basalt may be sourced from this singular upwelling ponded below the 660 km discontinuity (e.g., Davaille et al., 2005).

4.2.5 Hoggar

We observe the X-discontinuity beneath the Hoggar mountains at 270–280 km depth, deepening to the East (Figure 7a) where we find the most robust observations (Figure 5). While the X is present beneath Hoggar, its relationship to a potential mantle plume is uncertain. In the hotspot catalogue of Courtillot et al. (2003), Hoggar has one of the lowest probabilities of being a whole mantle plume (Marignier et al., 2020) and is classified as only 'somewhat resolved' by French and Romanowicz (2015). Further, the lavas in Hoggar are characterized by MORB-like $^{3}He/^{4}He$ ratios (Pik et al., 2006; Jackson et al., 2017) suggesting that they do not have a deep mantle source. The X is much deeper than the 150 km source depth of magmatism (Liégeois et al., 2005), thus separate processes may be invoked to explain the source of chemical heterogeneity and surface magmatism with little connection between the two. With limited station coverage it is difficult to assess the lateral extent of the X, and whether this observation is limited to the Hoggar mountains alone, or whether it is connected to the widespread X observations seen in Morocco.

4.2.6 Cameroon

Beneath Cameroon, we find the X colocated with the CVL. Robust X observations are made in western Cameroon, decreasing in confidence eastwards from positive to negative normalized votes. X observations are made at 250–290 km depth.

There is much debate as to the source of magmatism along the CVL, meaning a connection between magmatism and the X is uncertain here. Despite its linear morphology and a HIMU signature consistent with a lower mantle source (Lee et al., 1994), there is no age progression of magmatism along the CVL (e.g., Montigny et al., 2004). Further, maximum $^{3}He/^{4}He$ ratios are not distinguishable from MORB ratios (Barfod et al., 1999; Jackson et al., 2017). Continental and global tomographic models resolve a lower mantle plume (e.g., French & Romanowicz, 2015), suggest that there may be some lower

mantle contribution to magmatism (Emry et al., 2019; Boyce et al., 2021), or classify the likelihood of a mantle plume as 'unclear' (Marignier et al., 2020). Previous MTZ and regional tomographic studies of the CVL do find evidence of a thermal anomaly across the MTZ (Reusch et al., 2010, 2011), favoring edge-driven convection as the source of magmatism (King & Anderson, 1998; King & Ritsema, 2000). Other workers have also favored non-plume, low melt volume mechanisms for CVL development (e.g., lithospheric delamination or fault zone reactivation: Milelli et al., 2012; Gallacher & Bastow, 2012; De Plaen et al., 2014; Fairhead & Binks, 1991).

Boyce and Cottaar (2021) find complex MTZ behavior in their recent study, with reduced 410 amplitudes, 20–30 km of thinning and variable 660 km discontinuity behavior. With mechanisms of a water-rich MTZ (Buchen et al., 2018) and a high basalt fraction (sufficient for X observation; Kemp et al., 2019) available to explain a disappearing 410 km discontinuity, Boyce and Cottaar (2021) do not preclude a lower mantle contribution to magmatism along the CVL. Should such a large basalt accumulation be present atop the 410 km discontinuity, the Co-St phase transition would be a likely candidate cause of the X in this region, as seen in Hawaii (Kemp et al., 2019). However, the depth of the X is much shallower here than the 336 km reported by Kemp et al. (2019) and such large basalt fractions remain unattainable in geodynamic models (Monaco et al., 2022).

4.2.7 Madagascar

X observations are present in two distinct patches across Madagascar. South-easternmost Madagascar is underlain by robust X observations at 270-280 km depth, with a strong band of null results through central Madagascar separating a less certain region of X observations (normalized vote ≈ 0.5) to the north (Figures 5 and 7a).

Central and southern Madagascar are associated with substantial MTZ thinning and depression of both the 410 and 660 km discontinuities (Boyce & Cottaar, 2021), indicating a thermal upwelling across the MTZ. Recent tomographic studies image a low velocity anomaly extending from the surface beneath southernmost Madagascar to greater than 1000 km depth, connected to the African LLVP (Boyce et al., 2021; Tsekhmistrenko et al., 2021). Therefore, this anomaly maybe an upwelling branch of the African Superplume. An upwelling from the lowermost mantle south of Madagascar would explain the sharp transition from null to robust X observations from north to south (Figure 5), but

it is unable to explain moderately certain X observations in central Madagascar that underlie magmatism with a source depth at the base of the lithosphere ($\leq$130 km; depth Melluso et al., 2016).

Geochemical analyses of Cenozoic basalts in central Madagascar determine their provenance to be the Madagascan continental mantle, with uplift, lithospheric thinning and intercontinental rifting being the most likely processes to trigger melting (Melluso et al., 2016; Cucciniello et al., 2017). Magma has been suggested to have spread laterally as far as Madagascar from the EARS along pre-existing structures like the Davie ridge (Ebinger & Sleep, 1998; O'Connor et al., 2019). This is corroborated by plume-like signatures and age progression of volcanics in the Comoros islands (Emerick & Duncan, 1982; Deniel, 1998). Owing to a lack of resolution offshore Mozambique, these results are unable to discriminate between these two models of melt generation. It is also undetermined whether flow would be demarcated by X observations or whether this would be constrained to the uppermost mantle.

4.2.8 Mauritius and Réunion

Mauritius and Réunion present two further hotspot ocean islands underlain by an X. The X here is at 290–310 km depth, with the highest confidence southwest of Réunion and decreasing eastwards of Mauritius.

While Marignier et al. (2020) find evidence of a mantle plume to be 'unclear' in this region, French and Romanowicz (2015) find a plume to be 'clearly resolved' with high $^{3}He/^{4}He$ ratios (Graham et al., 1990) corroborating a lower mantle source. Regional seismic tomography reveals a slow wavespeed anomaly connected to the African LLVP (Tsekhmistrenko et al., 2021). Though Tsekhmistrenko et al. (2021) invoke individual blobs episodically detaching from the LLVP and subsequently ascending buoyantly, as opposed to a continuous plume conduit, this may provide a means to recycle basalt to the upper mantle in this region. Since plumes can only support a basalt fraction of 20% (Ballmer et al., 2013; Dannberg & Sobolev, 2015), ponding of eclogite has been invoked to match the seismically observable impedance contrasts of the X (Kemp et al., 2019). It is unclear how individually ascending blobs would affect the amount of basalt available to pond in the upper mantle.

4.2.9 Southern Africa

We largely observe a lack of the X in Southern Africa where the overlying lithosphere comprises several cratons. Figure 11 shows an abrupt termination of X observations at the margins of the Kalahari Craton which is characterized by high confidence null observations. As discussed before, it remains uncertain whether strong negative conversions from the base of the craton mask weaker converted phases in RFs or whether there is indeed a lack of chemical heterogeneity/necessary geodynamic conditions to observe the X beneath the thicker lithosphere extending to ≥200 km depth (e.g., Priestley et al., 2008; Fishwick, 2010; Adams & Nyblade, 2011).

In Pugh et al. (2021), six stacks beneath cratonic lithosphere from Canada, Brazil, Scandinavia, Siberia and Australia displayed null X observations to higher frequency than studied here, but strong negative conversions from the base of the craton still dominate the region for PXs conversions in depth and slowness stacks. Cooler than average mantle temperatures may be expected beneath the base of the cratons in southern Africa, as evidenced by thickening of the MTZ by 10-20 km (Blum & Shen, 2004; Boyce & Cottaar, 2021). This would raise the depth of the Co-St and OEN-HCEN transitions to ∼250 km depth (Schmerr, 2015). While cool mantle temperatures are not observed beneath all cratons (e.g., Thompson et al., 2011), for the thicker cratons, this may inhibit phase transitions associated with the X, preventing its observation (Figure S6). Additionally, in some locations, ponding of chemical heterogeneity may not be possible at the margins of the craton where complex flow is present due to edge-driven convection (King & Anderson, 1998; Currie & van Wijk, 2016).

5 Conclusions

X discontinuity structure beneath the African continent and surrounding ocean islands is vastly extended using widespread recordings of Pds RFs. The X is observed beneath the EARS, Morocco, Cameroon, Hoggar and several ocean islands (Canaries, Cape Verde, Madagascar, Réunion and Mauritius) at 233–340 km depth using a normalized vote mapping approach.

The X is recorded across a broad range of depths and amplitudes. With no apparent relationship to upper mantle temperature and widespread occurrence across variable geodynamic settings, a multigenetic origin of the X is required below continents. Ob-

servations of the X are typically collocated with surface regions of Cenozoic magmatism and Quaternary volcanoes, suggesting that surface magmatism is intrinsically linked to upper mantle chemical heterogeneity, however some notable exceptions exist. The broad connection may be explained by the presence of plumes of variable thermochemical nature beneath parts/all of the EARS, the Canaries/Morocco, Cape Verde, southern Madagascar and Réunion. However, it is difficult to explain the cause of upper mantle chemical heterogeneity beneath Cameroon, Hoggar and central Madagascar by a plume related mechanism where upper mantle processes are mostly likely responsible for surface magmatism. Whilst null observations dominate beneath cratons, it remains uncertain if this is linked to localized geodynamic conditions (e.g., lack of chemical heterogeneity or thermal anomaly) or is an artifact due to masking by shallower structure.

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Open Research

The facilities of IRIS Data Services, and specifically the IRIS Data Management Center, alongside the AUSPASS, GEOFON, ORFEUS and RESIF data centres, were used for access to waveforms, related metadata, and/or derived products used in this study. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience (SAGE) Award of the National Science Foundation under Cooperative Support Agreement EAR-1851048. The waveform data used in this study are from the following networks: 1C (Velasco et al., 2011), 2H (Keir & Hammond, 2009), 2L (Lange & Soler, 2019), 3D (Thomas, 2010), 4H (Hammond et al., 2011), 5H (Hammond, 2011), 6A (Heit, Yuan, Jokat, et al., 2010), 6H (Helffrich & Fonseca, 2011), 6R (I. D. Bastow, 2019), 7C (Vergne et al., 2014), 8A (Nyblade, 2015a), 9A (Weber et al., 2007), 9C (Heit, Yuan, & Mancilla, 2010), AF (Penn State University, 2004), G (Institut de physique du globe de Paris (IPGP) & École et Observatoire des Sciences de la Terre de Strasbourg (EOST),

1982), GE (GEOFON Data Centre, 1993), GH (Ghana Geological Survey, 2012), GT (Albuquerque Seismological Laboratory (ASL)/USGS, 1993), IB (Institute Earth Sciences "Jaume Almera" CSIC (ICTJA Spain), 2007), II (Scripps Institution of Oceanography, 1986), IU (Albuquerque Seismological Laboratory (ASL)/USGS, 1988), IV (INGV Seismological Data Centre, 1997), MN (MedNet Project Partner Institutions, 1988), NJ (Centre for Geodesy and Geodynamics, 2009), NR (Utrecht University (UU Netherlands), 1983), PF (Observatoire Volcanologique Du Piton De La Fournaise (OVPF) & Institut De Physique Du Globe De Paris (IPGP), 2008), PM (Instituto Português do Mar e da Atmosfera, I.P., 2006), TZ (Aubreya Adams, 2017), WM (San Fernando Royal Naval Observatory (ROA) et al., 1996), XA (Silver, 1997), XB (Wiens & Nyblade, 2005), XB (Levander et al., 2009), XC (Kind, 1998), XD (Owens & Nyblade, 1994), XI (Nyblade, 2000), XJ (Ebinger, 2013), XK (Gao et al., 2012), XM (The ARGOS Project, 2012), XV (Wysession et al., 2011), XW (Leroy et al., 2009), XW (Nyblade, 2017), Y1 (Ebinger, 2018), YA (Ebinger, 2012), YF (van der Lee et al., 1999), YH (Nyblade, 2010), YI (Gaherty & Shillington, 2010), YQ (Gaherty et al., 2013), YQ (Keir et al., 2017), YR (Levèque & RESIF, 2010), YV (Barruol et al., 2017), YY (Keranen, 2013), ZE (Ebinger, 2007), ZE (Tilmann et al., 2012), ZF (Fontaine et al., 2015), ZI (Wookey et al., 2011), ZK (Gao, 2009), ZP (Nyblade, 2007), ZS (Deschamps et al., 2007), ZT (Nyblade, 2015b) and ZV (Ebinger, 2014). Networks 1B, BX, TT, XJ, XJ, XM, YF, YJ, YK, YW, YZ, ZC, ZF, ZP, ZQ and ZU do not have DOIs and are detailed in Table S1. This study uses ObsPy (Megies et al., 2011). Scripts for receiver function creation and processing are available through SMURFPy (Cottaar et al., 2020).

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