Magnetotelluric evidence for layered mafic intrusions beneath the Vøring and Exmouth rifted margins

David Myer *, Steven Constable, Kerry Key
Scripps Institution of Oceanography, 9500 Gilman Drive MC-0225 La Jolla, CA 92093-0225, USA

Abstract

Marine magnetotelluric (MT) surveys at two volcanic passive margins have revealed an enigmatic layer of extremely high conductivity ($<0.1$ S/m) at ~10 km depth. At the Vøring Plateau off the northwest shelf of Norway, 2D inversion of data from nine sites along a 54 km line resolves a layer with a conductance of $\sim10^6$ S/m. At the Exmouth Plateau off the northwest shelf of Australia, 2D inversion of 122 sites in 17 lines finds a similar layer at similar depth but an order of magnitude higher conductance. At both plateaus, the depth of the high conductivity layer coincides roughly with what seismic studies have identified as an assemblage of sills. We propose that the extremely high conductance is due to well-connected conductive cumulates (e.g., magnetite) precipitated in layered mafic intrusions. In contrast to sill emplacement, the nature of layered intrusion formation requires connection to a magma source over time. Such a connection would not be likely during rifting when the rift provides a preferential pathway for pressure release. This implies emplacement prior to or during a pause in the early stage of continental breakup.

1. Introduction

In this work, we present the results of two marine magnetotelluric (MT) surveys carried out on separate passive rifted margins: the Exmouth Plateau off the northwest coast of Australia, and the Vøring Plateau off the west coast of Norway. The Exmouth MT data are part of an extensive electromagnetic study of the Scarborough gas field (Myer et al., 2012) whose primary purpose was to investigate the shallow structure in the vicinity of the hydrocarbon-bearing strata. Here we present results from the deeper-sensing MT data which confirm the presence of an extremely conductive body at mid-crustal depths that was first observed by Heinsen et al. (2005). The Vøring Plateau data derive from a smaller pilot study for sediment mapping and serendipitously reveal a layer of similarly high conductivity and at the same approximate depth.

The tectonic environment of these two plateaus is substantially similar. They are large areas of thinned continental crust protruding into the transition zone between continental and oceanic crust that remain from the onset of rifting (i.e., “passive margins”). At both plateaus, the rifting is thought to have had a quick onset and been accompanied by voluminous volcanic output (White, 1952; White and McKenzie, 1989). In this model, a large thermal anomaly (e.g., a plume head) arrives underneath the continent and weakens its structure, greatly thinning it prior to the initiation of rifting. The thinned continental crust is typically uplifted and eroded during the early stage, then, after a rift has formed, subsides as the rift moves away. The unexpected presence of an extensive, mid-crustal conductive layer in two plateaus with theorized similar formation histories may provide a clue as to the early processes involved.

In this paper, we examine the data from each survey independently then look for a possible common explanation. First we present the Vøring Plateau survey, whose single line of high quality data provides a simple introduction to the extremely high conductivity, mid-crustal layer. This is followed by the Exmouth Plateau survey, whose extensive data coverage provides seventeen intersecting inversion lines. Finally we discuss possible interpretations of the conductive layer.

2. Vøring Plateau

2.1. Description of the plateau and survey

The Vøring Plateau (Fig. 1) is a volcanic passive margin remaining from the breakup of Greenland and Norway during the opening of the North Atlantic. The bulk of the plateau is between 1 and 1.5 km depth and is bounded to the east by the shallower Trøndelag platform leading up to the coast, to the southwest by the Jan Mayen Fracture Zone, and to the northwest and north by an outer margin high which dips steeply down to abyssal depths. The outer margin high is comprised of a thin sedimentary cover over flood basalts inter-
spersed with broken and dipping pieces of continental crust (Eldholm et al., 1989a,b; Mutter, 1985; Mutter et al., 1982). The plateau itself has undergone many episodes of extension since the Carboniferous period and is consequently only 15–20 km thick. There is no evidence for a continuous detachment zone cross-cutting the entire crust to accommodate extension (Gernigon et al., 2003), instead it is composed of many buried basins bounded by listric faults which are roughly perpendicular to the eventual spreading direction (Skogseid and Eldholm, 1989).

The sedimentary section contains a major unconformity of late Cretaceous–Paleocene age and a series of Eocene tuffs which date the breakup of this margin at about 57 Ma. The Cretaceous section is heavily cut by normal faults and intruded by numerous sills at depths from 2 to 10 km. These sills are so common across the plateau that they have inhibited seismic resolution below them, requiring the use of ocean bottom seismometers for deeper imaging (Mjelde et al., 1997). Many of these sills appear to have chimney structures leading to numerous small vent-like craters in the Eocene tuffs which appear as “eyes” in the seismic profiles (Skogseid et al., 1992). The sills generally shallow in depth toward an escarpment that marks the boundary between the plateau and outer margin high where they appear to change from intrusive to extrusive in nature (Skogseid and Eldholm, 1989). The entire outer margin high and a portion of the plateau is underlain by a lower-crustal high-velocity body which is interpreted as magmatic underplating under the continental crust and thickened layer 3 under the oceanic crust (Gernigon et al., 2003).

In 2010, we deployed five broadband electromagnetic receivers (Constable et al., 1998) on the Vøring Plateau as part of a test of the viability of carrying out a seafloor MT survey during seismic operations. Each instrument was deployed twice for 8–9 days per deployment along a north–south trending line near the center of the plateau. The instruments were spaced 6 km apart for a total line length of 54 km, and all but one of the final deployments were recovered with data. The survey line is 9–40 km from four exploration wells (6605/1-1, 6605/8-1, 6605/8-2, 6706/11-1) whose logs show resistivities gradually increasing from 1 to 10 $\Omega$m over several km depth and a typical temperature gradient of $\sim$30°C/km down to 4 km bsl (Norwegian Petroleum Directorate, unpublished data). The heat flow regime of the plateau is generally homogeneous (Fernández et al., 2004).

MT apparent resistivity and phase data were derived from the time series using the multi-station processing algorithm developed by Egbert (1997) and are shown in Fig. 2. High quality data were obtained for periods from 4 to 10,000 s. The dimensionality of the data as expressed in phase tensor invariants (Caldwell et al., 2004), polarization ellipses, and the Swift skew (Swift, 1967) is very low, indicating that 1D and 2D interpretation is valid. The dimensionality begins to climb at periods longer than $\sim$1000 s for the southern end of the line (sites 1–3) and $\sim$100 s for the

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**Fig. 1.** A map of the Vøring Plateau with bathymetry contours every 500 m to 2500 m total depth. Deeper contours are omitted for clarity. The location of our MT survey is marked by the filled circles. The MT ‘survey line strikes 19° east of north.
northern end. Consequently we expect there to be some difference in the deeper structure between the two ends of the line as determined by inversion.

The apparent resistivity curves are similar for all sites, rising to a high at around 200 s. This characteristic is typical of marine geology in which water saturated conductive sediments of a few kilometers thickness overlie a resistive basement. However, a significant feature of this dataset is a steep drop in apparent resistivity at periods longer than 200 s for both modes of the data, indicating the presence of a highly conductive body at depth. Note that in 2D, MT data are typically rotated into two orthogonal modes known as the transverse electric (TE) and transverse magnetic (TM) modes. Throughout this paper, we define these modes such that “transverse” is the horizontal direction perpendicular to the survey line direction.

At the southern end of the line, the TE and TM mode apparent resistivities are nearly identical and a significant difference does not develop until site 6. Similarity of the two modes indicates a substantially 1D environment. The greater difference between the TM and TE modes for the northern sites compared to the southern sites indicates that the nature of the deep structure in the north is different. For example, our line of sites may cross the edge of a buried body.

2.2. Inversion

We used the Occam 2D MT inversion code of deGroot-Hedlin and Constable (1990) to invert the apparent resistivities and phases from the nine sites in this survey. This code uses a dual mesh approach (Wannamaker et al., 1987): a triangular finite-element mesh for the forward solver and a rectangular model grid for the inversion. The forward mesh was discretized more densely than the model grid to ensure accurate finite element computation. For this survey, we used a forward mesh with 101,920 triangular elements and an inversion mesh with 9332 model blocks (6613 of which are free parameters). We used simplified bathymetry from Smith and Sandwell (1997) for the plateau and the abyssal depths to the north and south. The model blocks are especially fine near the sites (<1/6th of the skin depth of the shortest period), growing larger with depth and lateral distance from the sites. The resolution of the MT method decreases with distance from the observation sites, so there is no need to mesh finely at depth.

We ran a variety of 2D inversions to examine the properties of the data. Apparent resistivities and phases were inverted with a minimum error floor of 1%. In every case discussed below, the inversion ran to a target misfit of RMS 1.0 within a few iterations,
then proceeded for a few further iterations to minimize the roughness of the model.

In Fig. 3 we show the result of inverting all the data. The conductive sediments extend to between 1 and 2 km below the sea-floor and the resistive basement is 5–7 km thick with resistivity values ranging between 2 and 100 Ωm, though this is likely underestimated since MT is not as sensitive to resistors as it is to conductors. The shallow sediments appear disconnected because of the large site spacing (~6 km) compared with the layer thickness. The resistive plug at ~20 km horizontal range and the deeper concentration of conductivity are most likely due to the inversion taking advantage of the lack of constraint caused by the absence of data from site 4. The difference between the south (left) and north (right) sides of the model reflects the split in the TM and TE modes observed in sites 6–10. It is unlikely that the difference in modes is caused by the coast effect (Heinson and Constable, 1992) because coast-perpendicular modeling indicates that attenuation from the mid-crustal conductive layer limits the periods at which the coast is sensed here to >4000 s.

The inversion converges quite rapidly to what is essentially a 4 layer model: 1–2 km thick conductive sediments, 5–7 km thick resistive basement, 6–8 km thick mid-crustal conductor which may pinch out in the north, and a deeper resistive half-space. Model responses are plotted on top of the data in Fig. 2. For reference, we have also plotted the responses for a model in which we have stripped out the mid-crustal conductor to show that it is very robust feature of the data.

To a certain extent, MT inversions may trade between conductivity and thickness to achieve substantially similar apparent resistivity and phase values. Thus anomalies are often characterized by their conductance (conductivity times thickness). Because the smoothness constraint of the inversion can smear out an anomaly and make it difficult to calculate conductance, we ran an additional inversion in which we added a “prejudice”: a penalty for deviation from an a priori background value. By using a resistive prejudice, we can bias the inversion to produce a model with the minimum conductance value for the conductive mid-crustal layer required to fit the data.

The prejudiced inversion squeezes the conductive layer down to what is effectively a single layer of model blocks at ~10 km depth while still fitting the data to a RMS misfit of 1.0. This result is as expected considering that in the 1D sense, MT data can be well fit by a series of delta functions in conductivity (Parker and Whaler, 1981) and our data have low dimensionality in the higher frequencies. The conductance of our layer varies laterally from a high of ~3 x 10^4 S in the south to about 10^3 S in the north. By way of comparison, the conductance of the overlying ocean is ~3 x 10^5 S and the conductance of the remarkably conductive graphite deposits at the KTB borehole in Germany is merely ~200 S (Emmermann and Lauterjung, 1997; Haak et al., 1991; Stoll et al., 2000).

3. Exmouth Plateau

3.1. Description of the plateau

The Exmouth Plateau is a large block of thinned continental crust off the northwest Australian margin remaining from the breakup of Gondwanaland, surrounded on three sides by abyssal depths (Fig. 4). Several periods of extension occurred in the Triassic and Jurassic before the final rift formation in the Early Cretaceous (Veevers and Johnstone, 1974). Rifting that formed the Argo Abyssal Plain north of the plateau occurred in the late Jurassic at an oblique angle to the final rifting direction as expressed by the western and southern edges of the plateau. The result is that while the southern transform boundary is linear and relatively sharp, the northern boundary is comprised of fractured, rotated blocks and voluminous volcanic output (Boyd et al., 1992; Exon et al., 1982, 1988; Larson et al., 1979). The early periods of extension led to large-scale normal faulting roughly perpendicular to the eventual spreading direction, and to the formation of numerous near-shore sub-basins now buried beneath sediment fill. These sub-basins may be the expression of an early failed rift arm of a triple junction that has since moved off and subducted (Direen et al., 2008; Larson et al., 1979).

The Exmouth Plateau crust is approximately 20 km thick (Mutter and Larson, 1989), the top 3–5 km of which are Cretaceous and younger sediments overlying Triassic and Paleozoic rock. The plateau was uplifted and truncated during the Jurassic forming a major unconformity in seismic sections. The layers above this unconformity are generally not cut by faulting. Older layers are cut by steeply dipping faults which become listric or sole out at mid-crustal depths between 8 and 15 km (Boyd and Bent, 1992; Boyd et al., 1992; Driscoll and Karner, 1998; Exon et al., 1982).

The plateau has undergone thinning by at least a factor of two which Driscoll and Karner (1998) model using depth-dependent extension having brittle deformation in the upper crust layered over ductile deformation in the lower crust. They propose that the change in deformation style is accommodated by a sub-horizontal detachment zone at ~15 km depth. A magnetotelluric and geomagnetic depth sounding survey by Heinson et al. (2005) identified a highly conductive body (~<0.1 Ωm) in the mid-crust which they interpreted as being due to well-connected carbon accumulated along the proposed detachment. We shall discuss this hypothesis in more detail in a later section, but it is important to note at this juncture that Stagg et al. (2004), in a summary of research relating to the Exmouth Plateau, find that there is significant evidence to contradict Driscoll and Karner’s proposed detachment zone. Instead, volcanic sills intrude extensively throughout the mid-crust at 8–12 km including in the area of the plateau beneath our survey (Rey et al., 2008). Joint interpretation of seismic reflection, seismic refraction, gravity, and magnetic data

![Fig. 3. 2D inversion results for the Voring Plateau. Data were inverted with a 10% error floor and converged to an RMS misfit of 1.0. There is no vertical exaggeration. North is to the right.](image-url)
show that an interpretation of massive sills on the order of several km thick (i.e. the scale of a layered intrusion) is consistent with the data (Direen et al., 2008).

Our survey involved the deployment of broadband electromagnetic receivers in a 144 site grid to collect both controlled-source (all sites) and MT data (122 sites) over the Scarborough gas field. The orientation of the survey is such that lines are roughly either parallel or perpendicular to the fossil spreading direction. The experimental design and data quality are fully reported in Myer et al. (2012). Here we note that the Exmouth Plateau is periodically swept by strong, tide-related deep currents (Lim et al., 2008; Van Gastel et al., 2009) which caused considerable shaking of some instruments. Consequently, MT data collected here were not as high quality as that shown above for the Vøring Plateau.

Impedances were successfully derived from 112 sites for the period range from 20 to 2000 s. Impedance polarization ellipses are nearly circular for all data and the Swift skew is low as well. There are two remarkable features in these data. The first is that they are almost entirely rotationally invariant at every site. This homogeneity of the two modes indicates a primarily 1D environment. Second, the downturn in resistivity observed in the Vøring data at around 300 s is also observed here but is much sharper.

Fig. 5 shows the envelope of all the Exmouth apparent resistivity and phase values for both TM and TE. The Vøring data are overlaid on the figure for comparison. The downward turn of apparent resistivities toward a conductive lower crust is sharper and steeper than for the Vøring data, which suggests that inversion of the Exmouth data will resolve an even more conductive body. The similarity across all sites makes it straightforward to estimate that the basic geologic structure is a conductive sediment package underlain by a resistive basement which rests on top of an extremely conductive layer.

The wide band in the Vøring envelope at longer periods obscures the fact that there is a distinct upward turn in the TE mode apparent resistivities at about 2000 s. The Exmouth data may also turn upwards, but data at these long periods have large enough uncertainties that it is not strictly required. In the inversions which follow, very few data above 2000 s have been used because of the large uncertainties.

3.2. Inversion of line 1

We inverted the southernmost line of the Exmouth survey ("line 1") using the same code as for Vøring. This is the longest line in the Exmouth dataset and is composed of 34 sites with inter-site spacing between 0.5 and 4 km and covering 50 km. It is coast perpendicular, so the model space includes both the shallowing continental slope and coast in the east as well as the distal plateau slope.
and abyssal plain in the west. The forward mesh has \( \approx 152,000 \) finite element triangles and the inversion uses 8695 model bricks, 7816 of which are free parameters. As before, the model bricks near the seafloor are sized to be less than \( 1/6 \) of a skin depth for the shortest period and grow in size with depth and lateral distance from the survey line. We inverted apparent resistivities and phases for both TE and TM modes (1328 data points) with a minimum 10% error floor. The inversion converged to RMS 1.0 and produced the model shown in Fig. 6a. As predicted from the data, the model is primarily three layered: shallow 1 m conductive sediments over 10 m crustal basement, both underlain by a conductor whose resistivity values vary between 0.1 and 0.01 m. Also as predicted, our data do not resolve the bottom of the conductive layer, which extends in all directions to the edges of the model space.

Surprisingly, the top surface of the conductive layer contains large amplitude "ripples" (3–5 km high \( \times 10–15 \) km wide) which a casual examination of apparent resistivities and phases from site to site does not reveal. To determine whether the ripples are an artifact of a poor fit to a few data, we examined the misfit for each site separately. We found that there is no correlation between the misfit and the ripples; data from sites over the ripples are fit just as well as data from other sites. A similar examination of misfit by data type (i.e. TE and TM apparent resistivity and phase) and by period shows no dominant source of misfit skewing the resulting model. These are robust features of the data.

Though we did not obtain long enough period data to resolve the bottom of the conductive layer, we can still conduct the same prejudiced inversion experiment as on the Vøring data in order to determine conductance. From the resulting model (Fig. 6b) we find that the conductive layer must have a conductance of at least \( 10^5 \) S in order to fit the data to an RMS misfit of 1.0. Because there is no definitive resistive up-turn in the MT data, we cannot place an upper bound on this already extremely high value.

### 3.3. Inversion of all lines

We inverted all seventeen lines of Exmouth data in 2D and show the results in a fence plot in Fig. 7. These models all converged to an RMS misfit between 1.0 and 1.1 and in each case the misfit is distributed evenly among the various components, frequencies, and sites as discussed for line 1. In order to provide a good view of the surface of the conductive layer, the top portion of each model has been cut away to the depth at which the conductivity dips below 0.3 m. The depth of the conductive surface away from the ripples is 11–12 km.

From the fence diagram, it appears that the conductive layer contains an elevated feature like a ridge-line which strikes north-east, approximately parallel to the distal edge of the plateau. Note however that with the exception of line 1, the east–west lines include only 6 sites each, positioned where they cross the north–south lines. Therefore, the lateral position and amplitude of the ripple is not well constrained in these lines. It appears to be at least 25 km long and dissipates to the north. Referring back to the ambiguity of the line 1 prejudiced inversion, note that this feature may be a disconnected conductive body above the conductive layer.
Fig. 6. 2D inversion results for the long E-W trending line of the Exmouth Plateau survey. (a) Inversion of all data with a uniform roughness penalty. The inversion is unable to resolve the thickness of the conductive layer due to the dearth of long period data. The roughness penalty smooths the conductor to all edges of the model. (b) Inversion with a uniform roughness penalty and a penalty for deviating from 10 Ωm. The long wavelength depth variations in the surface of the mid-crustal conductor are resolved into localized conductive bodies. The conductive layer is thinned to a single model brick and yields the minimum conductance required to fit the data.

Fig. 7. 2D inversion models for all seventeen lines of the Exmouth data. The top of each model is stripped off down to the depth at which the resistivity drops below 0.3 Ωm. Filled circles are the site locations in each inversion, draped on the surfaces for clarity. The east–west lines consist of 6 sites each except for Line 1 which has 34. The north–south lines have 10 or more sites each. All inversions converged to a misfit between RMS 1.0 and 1.1.
4. Mid-crustal conductors

Lower and mid-crustal conductors are found in a wide variety of tectonic settings. Values of 20–30 $\Omega$m in mid-to-lower continental crust and 1–10 $\Omega$m in mid-to-lower oceanic crust are usually considered “low” (Bedrosian, 2007), so our extremely low values of 0.1 to 0.01 $\Omega$m are unusual. We rule out systematic problems with the receiver instruments or their calibration since we have used these receivers in other areas of the world without detecting deep conductors (e.g. Constable et al., 2009), and values in the same range were derived by Heinson et al. (2005) with a completely different equipment system.

The possible causes of mid-crustal conductors have been reviewed by Simpson (1999) and Bedrosian (2007) and in general are: melt, pore fluids, or mineral phases. Of this latter, some possibilities are graphite precipitated along fault zones, well-connected serpentinization products (e.g. magnetite), and conductive minerals precipitated along layers in mafic sills. For these two plateaus, we can confidently eliminate melt and pore fluids. To obtain the very low resistivity values modeled, 100% basaltic melt would be required (Roberts and Tyburczy, 1999; Tyburczy and Waft, 1983) and there is no evidence of an extant melt body below either plateau (i.e. their heat flow regimes are average). Regarding pore fluids, Heinson et al. (2005) point out that to produce the very low resistivities modeled here either the pore fluid conductivity would need to be unrealistically low ($10^{-5}$ $\Omega$m) or the porosity unreasonably high (50%) for the 10–11 km depth.

Serpentinized peridotite may produce low conductivities if magnetite has been precipitated along grain boundaries forming a well connected matrix (Stesky and Brace, 1973). Magnetite by itself can be much more conductive than 0.1 $\Omega$m (Drabble et al., 1971; Kakudate et al., 1979; Lorenz and Ihle, 1975; Verwey and Haayman, 1941); however in a mineral assemblage with serpentinized peridotite, conductivities this low have not been observed in the lab. This is possibly due to the irreversible destruction of the magnetite matrix during exhumation of the samples studied, as suggested by Frost et al. (1989) in the context of graphite films. We note that a recent study claims to have definitively measured serpentinite resistivity in the $10^4$ $\Omega$m range (Reynard et al., 2011), but in this case the samples were first ground up and pressed into larger samples, destroying any magnetite connectedness and making these measurements only valid for the bulk serpentinite minerals (hydrated magnesium silicates).

Even supposing that in situ serpentinized peridotite could produce low enough resistivities, seismic velocity structure does not favor it as an explanation for the Exmouth and Voring anomalies. The p-wave velocity structure for the Exmouth Plateau, for example, is less than 5.0 km/s down to ~10 km then 5.7–6.4 km/s for another ~10 km (Fomin et al., 2000; Mutter and Larson, 1989). If the conductive anomaly at ~11 km depth represents serpentinized peridotite, then the velocity structure indicates it is part of a 10 km thick layer which is ~70% serpentinized (Christensen, 2004; Horen et al., 1996) – i.e. the currently supposed lower crust is actually exhumed, serpentinized mantle. Serpentization of peridotite increases its volume which, at depth, has the effect of decreasing the porosity and sealing off further peridotite from serpentinization. In order to serpentinize a 10 km thick layer, it must be continuously extended and fractured to promote water circulation. Such fracturing would presumably break any magnetite connectivity and preclude the existence of a highly conductive body. Also, at the non-volcanic Iberian passive margin where massively serpentinized peridotite has been confirmed by scientific drilling (Girardeau et al., 1988; Seifert and Brunotte, 1996), the velocity structure contains only a few km of low velocities (5–6 km/s) over a much higher velocity lightly-serpentinized basement (~7.6 km/s) (Boillot et al., 1992; Chian et al., 1999; Whitmarsh et al., 1999). This same structure has been observed at other rifted margins such as Newfoundland (Reid, 1994) and the Great Australian Bight (Sayers et al., 2001), indicating that large thicknesses of serpentinized mantle peridotite are generally in the >7 km/s range. The velocity structures of the Exmouth and Voring Plateaus do not fit this pattern suggesting that our enigmatic conductive layer is not due to serpentinization.

Heinson et al. (2005), who originally discovered the extremely conductive layer at Exmouth, suggested it was due to graphite films precipitated along the detachment zone proposed by Driscoll and Karner (1998). They suggest that the source of the carbon is either mantle carbonate that decomposes as it shallows and is preferentially precipitated along a redox front represented by the brittle–ductile transition, or methane rich sediments that precipitate carbon along shear zones. However, there are several difficulties with this explanation. We place the conductor nearly 5 km shallower than the proposed detachment zone at the Exmouth Plateau, making its association with the brittle–ductile transition unlikely. Because our survey is more densely spaced and contains higher frequencies, our shallower depth is more robustly determined than the Heinson et al. (2005) result. Also, the surface of the conductive anomaly may contain a structural rippie that is not consistent with a detachment interface unless there was post-rifting deformation. The overlying strata contain no evidence for this deformation, which would likely break the graphite connectivity anyway. Alternatively, the structural rippie may be discrete conductors at different depths, which are also not explained by a redox front argument.

A detachment-related explanation has even more difficulties in the context of the Vøring Plateau where there is no evidence for a continuous detachment zone cross-cutting the entire crust (Gernigon et al., 2003). Instead, this plateau is composed of many buried basins bounded by listric faults which are roughly perpendicular to the eventual spreading direction (Skogseid and Eldholm, 1989). If the cause of the conductive anomaly is the same at these two plateaus, which seems reasonable given their similar formation histories and conductivity structures, it is very likely not detachment related.

Since both plateaus are heavily faulted down to the lower crust, one may still argue that the conductive anomaly represents graphite films precipitated along these faults in a narrow depth range (Frost et al., 1989; Mathez et al., 2008). The dominant mode of faulting is sub-vertical listric faults. If carbon films form along these faults, they would need to be well-connected along the fault plane for many tens of kilometers and ubiquitous along all faults in order to present such a well-connected model in the grid of MT inversions at Exmouth. Furthermore, the pressure–temperature argument required to constrain graphite connectedness to a particular depth would not explain the depth variations of the Exmouth data.

We propose an explanation involving well-connected mineral phases in layered intrusions injected into the mid-crust as part of continental breakup. Both plateaus have features identified as sills at the depths corresponding to our conductive anomalies. Given connection to a magma source over time scales of $10^3$–$10^5$ years, mafic sills can form layered intrusions in which minerals precipitate into discrete layers. Such layers are observed to extend laterally over distances of ~100 km (Lee, 1996; Naslund and McBirney, 1996). Also, layers of magnetite have been observed with thicknesses ranging up to several meters in large layered intrusions such as the Bushveld Complex in South Africa (Eales and Cawthorn, 1996; Lee, 1996), the Bjerkreim-Sokndal Intrusion in Norway (Wilson et al., 1996), the Skærgaard Intrusion in Greenland (McBirney, 1996), and the Windimurra Complex in Australia (Mathison and Ahmat, 1996).
The Heinson et al. (2005) study reported a conductance of $2 \times 10^{-5}$ S, which is compatible with our results. Considering that the conductivity of pure magnetite is about $10^{-3} - 10^{-4}$ S/m (Drabble et al., 1971; Kakudate et al., 1979; Lorenz and Ihle, 1975), this is equivalent to a $1 - 10$ m layer or assemblage of layers in each plateau. Ferré et al. (2009) examined borehole cores from the Inisizwa and Bushveld layered intrusions in South Africa and the Great Dyke in Zimbabwe. Their results showed that magnetite layering occurs throughout the sills and magnetite cumulates are present in most layers.

Regarding lateral extent, the 54 km long Voring Plateau survey line is located near the center of the plateau and a few tens of kilometers landward of the Voring Escarpment, which marks the transition from intrusive to extrusive volcanism. It seems unlikely that the conductive anomaly would continue all the way over to the Escarpment due to the shallowing of emplacement depth creating conditions unfavorable to the formation of layered intrusions. Oceanic crust is known to be resistive ($\sim 1000$ $\Omega$m) (Constable and Cox, 1996; Naif et al., 2013) and we suspect that the continent–ocean transition region with its large, often subaerially-placed basalt flows intermixed with continental crust is similarly resistive.

At the Exmouth Plateau, our survey covers a 40 km by 50 km area and finds a mid-crustal conductive layer comparable to that found by the 400 km long transect of Heinson et al. (2005) which is $>200$ km further north. This similarity implies that the conductive anomaly covers all or a large section of the plateau. Our survey is approximately midway between the continent–ocean transition and the inner sub-basins that represent an earlier failed rifting event. The 400 km transect crosses the sub-basins up onto the shallow shelf. If the conductive anomaly does extend across the entire plateau as implied, then it is reasonable to assume that sills were injected throughout the plateau, probably from both the earlier failed rifting event and the terminal rifting event. White et al. (2008) note a ratio of intrusive to extrusive volcanism of approximately 2:1 at two separate volcanic margins which each contain failed riffs. It is possible that the Exmouth Plateau has experienced a similar ratio.

We cannot definitively conclude that the conductive anomaly at the Voring and Exmouth Plateaus is due to conductive minerals in layered intrusions. Laboratory conductivity measurements of cumulates from layered intrusions at appropriate pressures and temperatures are particularly lacking. However, the other evidence for this hypothesis is reasonable: sills are observed at the appropriate depths in seismic data at both plateaus. Sub-aerially exposed layered intrusions around the world are observed to contain well-connected magnetite layers whose aggregate thicknesses are sufficient to explain the conductance values we observe. Magnetite, in particular, can be very conductive and it may be that the pressure and temperature conditions at this depth of burial are sufficient to provide electrical connection across large distances.

If our layered intrusion hypothesis can be confirmed, we expect that it would slightly modify the formation history for these two volcanic passive margins. In each case, terminal rifting was preceded by a number of extensional events that failed to proceed to rifting, instead creating a number of extensional basins and sill intrusions. The formation of a layered intrusion is more complicated than a simple sill in that it requires access to a magmatic source which stays open in order to provide a driving force for the internal convection which results in the eventual deposition of highly defined layers (Naslund and Mcirney, 1996). This requires that pressure be maintained in the system and not relieved by rifting or extension. If our conductive anomaly is due to magnetite in a layered intrusion, then extension must cease or be exceptional slow for a period of some tens to hundreds of thousands of years.

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