Rifting of a passive margin and development of a lower-crustal detachment zone: Evidence from marine magnetotellurics

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[1] Rifting of passive margins has, in some places, led to the development of extensive submerged continental platforms. Crustal thinning by a factor of two or more is at least partially accommodated by high-angle normal faults that extend through the brittle upper crust to the ductile parts of the lower crust. However, an approximately horizontal detachment zone has been postulated to explain the width of the shelf and differences between the amounts of upper crust and lower lithosphere extension. Here, we present magnetotelluric (MT) and geomagnetic depth sounding (GDS) measurements across the Exmouth Plateau, a deep ($\sim$1500 m) and wide (>500 km) continental platform on the Northwest Shelf of Western Australia. MT responses across all of the Plateau in the bandwidth 10^2–10^4 s exhibit a pronounced apparent resistivity low of less than 1 Ω.m at $\sim 2 \times 10^3$ s. Two-dimensional smooth inversion of all MT and GDS data images a very low resistivity layer ($\sim$0.1 Ω.m) at a depth of 10–15 km, dipping less than 1° landward from the continent-ocean boundary. Below this layer, resistivity increases to $>10^3$ Ω.m beneath the Moho at 20 km depth. The low-resistivity mid-crustal layer coincides with a laterally extensive seismic reflector at 10–15 km observed through much of the Exmouth Plateau. We suggest that this layer represents a decoupling zone between ductile and brittle continental crust, with a concentration of conducting mineralisation along shear zones. The most probable conducting mineral is graphite; this is known to precipitate along shear zones, and would significantly facilitate extensional strain. **Citation:** Heinson, G., A. White, and F. E. M. Lilley (2005), Rifting of a passive margin and development of a lower-crustal detachment zone: Evidence from marine magnetotellurics, *Geophys. Res. Lett.*, *32*, L12305, doi:10.1029/2005GL022934.

1. Introduction

[2] The Exmouth Plateau is a type example of an offshore rifted continental platform resulting from multiple phases of extension [*Exon et al.*, 1982]. It is a broad plain approximately 500 km wide, bounded to the north and south by transform faults that separate continental crust from oceanic crust, and is mostly about 1.5 km deep (Figure 1). Near-shore continental shelf regions (<200 m deep) contain a number of deep sedimentary basins, such as the Barrow, Dampier and Exmouth Sub-Basins that contain significant hydrocarbon resources.

[3] Seismic reflection studies [*Mutter et al.*, 1989; *Stagg and Colwell*, 1994], and wide-angle reflection and refraction [*Goncharov et al.*, 2000] have defined a seismic velocity cross-section across the Exmouth Plateau. From the earlier reflection surveys the Moho is about 20 km thick for most of the Plateau, deepening to 40 km onshore beneath the Pilbara craton, indicating crustal thinning by a factor of two. To accommodate such crustal thinning a number of hypotheses have been proposed [*Lister et al.*, 1986; *Buck*, 1991; *Lister et al.*, 1991; *Etheridge and O'Brien*, 1994; *Driscoll and Karner*, 1998; *Gartrell*, 2000]. Most authors agree that much of the extension is accommodated by high-angle normal faults that extend through the brittle upper crust to the ductile parts of the lower crust [*Mutter et al.*, 1989; *Stagg and Colwell*, 1994]. However, associated with the lateral motion is the concept of an approximately horizontal detachment zone to explain the width of the shelf and differences between the amounts of upper crust and lower lithosphere extension [*Driscoll and Karner*, 1998].

[4] Extensive seismic surveys indicate the presence of a bright reflector at 9–10 s TWT ($\sim$15 km), and normal faults are observed to sole at this boundary [*Stagg and Colwell*, 1994]. This reflector is suggested to be a mylonitic boundary [*Mutter et al.*, 1989], indicating a major change in acoustic impedance, and a relatively reflector-free lower crust with velocity of 5.7–6.4 km/s [*Goncharov et al.*, 2000]. The presence of a strong reflective layer does not necessarily imply that coherent displacement over hundreds of kilometres has occurred [*Driscoll and Karner*, 1998; *Gartrell*, 2000].

[5] In this paper we present MT and GDS data that yield additional constraints on crustal structure of a rifted and extended continental margin. Such constraints provide fresh

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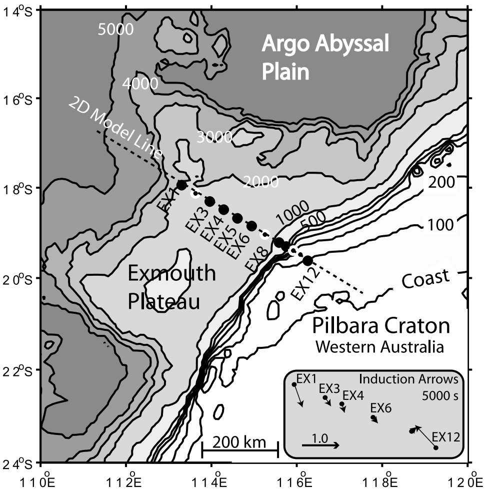

Figure 1. Bathymetry of the Exmouth Plateau showing the MT instruments deployment locations. Solid circles show instruments that returned useful data. The inset figure shows in-phase induction arrows for a period of 5000 s, reversed in the Parkinson convention.

evidence for an extensive layer that may support displacement, and which appears to dip landwards.

2. Experiment

[6] Figure 1 shows the location of seafloor instrumentation (listed in Table 1), deployed for one month between August and September 2000. At most sites, we recorded time series of natural variations of electric and magnetic fields. Data were sampled at 10 s intervals, with a resolution of 0.1 nT in magnetic and 0.05 μV/m in electric field. Standard MT tensor and GDS responses were obtained over a bandwidth of $10^2 - 10^4$ s using a robust remote-reference method [*Chave and Thomson*, 1989]. Data quality from the continental shelf was significantly degraded due to movement of the instruments caused by strong tidal motions (10 m tides in $\sim$200 m of water), and only EX12 produced useful data.

[7] Distortion analysis was carried out using a Mohr circle approach [*Lilley*, 1998]. It was found that MT data were approximately 1D at the shortest periods, changing to 2D at longer periods, but showed no significant electric field distortion. The dominant 2D strike was parallel with the coastline and the edge of the continental shelf margin and western edge of the Exmouth Plateau, but the deep ocean to the north introduced some complexity into the responses.

[8] The mode of induction with electric field parallel to the coastline is denoted TE; the orthogonal mode is TM. In the same orientation, GDS data are also a TE mode response. Figure 2 shows MT data from site EX6 expressed as apparent resistivity and phase for Zxy and Zyx components of the impedance tensor, corresponding to the TE and TM modes respectively. Components Zxx and Zyy are much smaller, and indicate that the data have no significant 3D component over the entire bandwidth. The response at EX6 is typical of sites from EX3 through EX8. Apparent resistivities from the TE and TM modes are similar between 10^2 to 2×10^3 s, and decrease from 3 Ω.m to 0.7 Ω.m. At periods $>2 \times 10^3$ s, the split between TE and TM modes is indicative of a change from 1D to 2D structure, with a trend to higher resistivities at depth. Responses at sites EX1 and EX12 show more variation, but are still predominantly 2D. For site EX1 apparent resistivities are generally higher than at all other sites, between 4 and 20 Ω.m, and do not show a characteristic low resistivity at 2×10^3 s.

[9] GDS responses at sites EX1 and EX12 are much larger than at all other sites, as shown in Figures 1 and 3, and are reversed in sign across the Exmouth Plateau. Large GDS responses are indicative of lateral variations in resistivity, and a reversal in sign requires a concentration of electric current between sites. Expressed as Parkinson in-phase induction arrows [*Parkinson*, 1962] in Figure 1, the arrow at EX1 points away from the deep ocean, while the arrow at site EX12 points away from the coast. That EX1 arrow is orientated towards the coast implies that the conductance (depth-integrated conductivity) of the continental crust must be greater than that of 5 km of seawater, which is 1.5×10^4 S. In the next section we show that this is indeed the case.

3. Results

[10] Given that the data display 1D and 2D characteristics, it is valid to carry out 2D inversions using the smooth Occam algorithm [*DeGroot-Hedlin and Constable*, 1990]. We inverted the entire MT and GDS data set (592 data points) with jack-knife derived errors [*Chave and Thomson*, 1989] of 5% or less for most periods to an rms error of $\sim$2 (Figure 4). Model responses at site EX6, in the middle of the Exmouth Plateau, are shown in Figure 2, while Figure 3 shows the fits to all the GDS data. Misfits at periods of $>10^4$ s at sites EX1 and EX3 probably reflect the influence of the deep ocean to the north and northwest

Table 1. Location and Deployment Details

Site	Latitude, S	Longitude, E	Depth, m	Notes
EX1	17°56.63′	113°18.29′	1902	
EX2	18°07.59′	113°37.84′	1819	no data
EX3	18°18.49′	113°57.72′	1549	
EX4	18°29.33′	114°17.37′	1420	
EX5	18°40.90′	114°36.62′	1604	mag. flooded
EX6	18°51.19′	114°56.62′	1732	
EX7	19°02.41′	115°15.22′	1512	lost
EX8	19°13.00′	115°35.17′	542	
EX9	19°18.32′	115°45.13′	314	tidal movement
EX10	19°23.64′	115°55.13′	214	tidal movement
EX11	19°29.00′	115°05.10′	132	tidal movement
EX12	19°37.20′	115°15.48′	104	

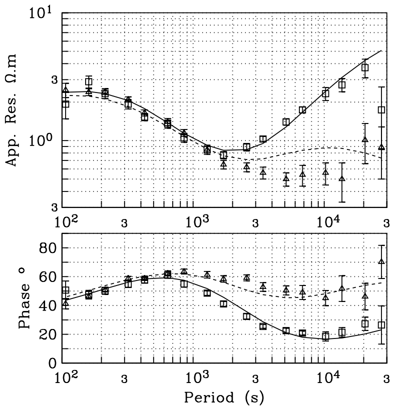

Figure 2. Apparent resistivity and phase data at site EX6. The square symbols represent the TE mode with electric currents flowing parallel to the coastline; triangles represent the TM mode with the magnetic field parallel to coast. Solid and dashed lines represent the model responses for the 2D inversion in Figure 4.

[11] In Figure 4 we have indicated primary structural boundaries from the seismic velocity model [*Stagg and Colwell*, 1994]. Within the upper 5 km we image low resistivities of 1–10 Ω.m, due to porous Tertiary sediments, including the Dampier Sub Basin beneath site EX12. From Archie's Law, a bulk sediment resistivity of 1–10 Ω.m would imply a porosity of between 50 and 16% for seawater resistivity of 0.25 Ω.m

[12] The most striking feature is the low-resistivity layer (~0.1 Ω.m) at a depth of between 10–15 km, approximately the depth of the brittle-ductile zone. The layer is at a similar depth to the bright seismic reflector [*Stagg and Colwell*, 1994]. The dip is about 5 km in 500 km distance, or about 1° of slope, landwards towards the Pilbara Craton. The thickness of this layer is hard to determine by the nature of smoothing inherent in the inversion, and diffusive nature of EM fields. However, a conductance can be estimated as

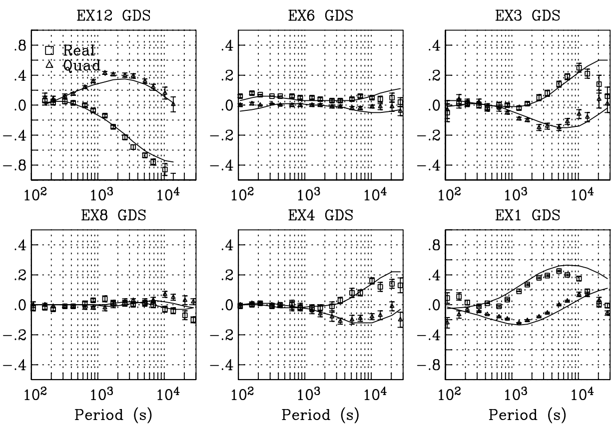

Figure 3. Vertical field GDS responses at sites with useful data indicating the ratio of the vertical magnetic field to the horizontal magnetic field perpendicular to the coastline (along the direction of the 2D inversion). Square symbols represent the in-phase component of the transfer function; triangles are for the out-of-phase component. The lines represent the model responses for the 2D inversion in Figure 4.

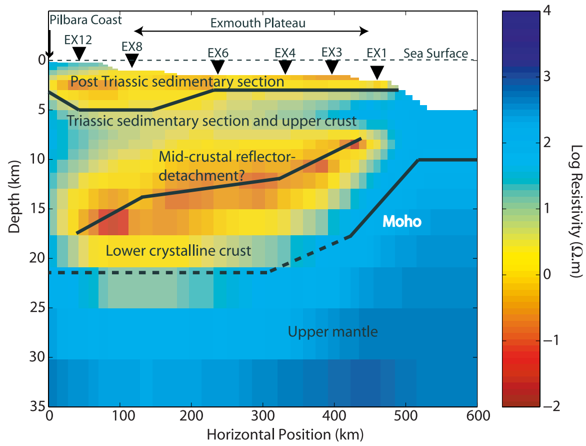

Figure 4. A smooth model from a 2D Occam inversion of all MT and GDS responses. Solid and dashed black lines indicate the main structural boundaries [adapted from *Stagg and Colwell*, 1994]. Fits to the MT and GDS data are shown in Figures 2 and 3, respectively.

2×10^4 S. We note that this is higher that the conductance of 5 km of seawater of about 1.5×10^4 S and hence is consistent with the observation that the induction arrows at site EX1 are more sensitive to the continental crust than the deep ocean.

[13] Beneath the low resistivity layer, resistivity increases rapidly to $>10^3$ Ω.m beneath the Moho. With only six sites, mapping the mantle boundary beneath continental and oceanic crust is somewhat tenuous, but it is most probable that the uppermost mantle is resistive, based on studies of olivine at temperatures of <1000°C [*Constable et al.*, 1992].

4. Discussion and Conclusion

[14] Mechanisms to cause low resistivity in mid to lower continental crustal settings have been extensively reviewed [*Simpson*, 1999]. There are three primary candidates, namely graphite [*Frost*, 1989; *Glover*, 1996], free fluids [*Wei et al.*, 2001] and other mineralisation such as magnetite associated with serpentinization [*Stesky and Brace*, 1973]. We cannot unambiguously discriminate between these mechanisms without petrophysical sampling, but it is possible to make some general comments on the feasibility in each case.

[15] Serpentinisation occurs due to the interaction of water with olivine. Magnetite grains form halos around the original olivine grains during hydration, resulting in bulk resistivity <100 Ω.m [*Stesky and Brace*, 1973]. A serpentinised region of under-plating at the continental-ocean boundary (beneath site EX1 in Figure 1) with velocity of 7.2 km/s at a depth of 20–25 km has been identified [*Stagg and Colwell*, 1994; *Goncharov et al.*, 2000], but there is no equivalent region of low resistivity at this depth in Figure 4.

[16] Saline fluids are also unlikely as a mechanism for conduction at the brittle-ductile zone [*Simpson*, 2001]. A bulk resistivity of 0.1 Ω.m and 1% porosity would imply a fluid resistivity of 10^{-5} Ω.m, which is unrealistic, even at temperatures of several hundred degrees Celsius. For a more realistic saline fluid resistivity of 0.025 Ω.m at 350°C a porosity of 50% would be required, which is

also unrealistic. Continuity of the low resistivity zone implies that fluid would have to be extremely well connected laterally, but impermeable in the vertical direction, and would have to have been present since the break up.

[17] The most probable explanation is due to graphite mineralisation [*Glover*, 1996]. Carbon films 50–200 Å thick around grain boundaries can reduce resistivity to <100 Ω.m [*Frost*, 1989], and thicker graphite flakes along shear zones may reduce resistivity to a significantly greater extent. There are two possible sources of graphite.

[18] One explanation is that mantle degassing of volatiles occurred during phases of extension. *Keppler et al.* [2003] argue that carbon is stored in the mantle as a separate carbonate phase that is mobilized during low degrees of partial melting during extension. Upon decompression below 2.5 to 3.5 GPa the carbonate phases decompose and releases CO_2, to migrate upwards through the crust. *Glover* [1996] suggests that crustal iron-titanium rich oxides react with CO_2 under reducing conditions to precipitate graphite C, as:

$$6Fe_2TiO_4 + CO_2 \rightarrow 2Fe_3O_4 + 6FeTiO_3 + C \qquad (1)$$

[19] The process begins at temperatures of 600–650°C in the crust and produces magnetite, ilmenite and graphite

[20] It is possible that the brittle-ductile decoupling represents a redox front at which equation (1) precipitates a large amount of graphite. As the Exmouth Plateau has experienced a near uniform strain over 500 km [*Stagg and Colwell*, 1994], then precipitation of carbon could be widespread and hence produce a continuous conductor.

[21] An alternative would be that methane-rich sediments precipitate carbon as flakes along shear zones through the reaction [*Glover*, 1996]

$$CH_4 + CO_2 \rightarrow 2H_2O + 2C \qquad (2)$$

[22] This reaction occurs at temperatures <300°C, but requires a source of methane. Given the depths and lack of subduction history, it is unlikely to have a source of biogenic carbon to produce methane. Carbonate rich fluids may be present along normal faults, but the carbon would be precipitated at shallow depths due to the strong redox gradients near the surface [*Glover*, 1996]. The amount of graphite required to reduce resistivities to 0.1 Ω.m over many hundreds of kilometres implies a high degree of continuity. This implies that since emplacement there would have to have been little to break up such continuity.

[23] Figure 4 indicates landward dip of less than 1° of the conductive zone, coincident with the seismic reflector [*Stagg and Colwell*, 1994; *Goncharov et al.*, 2000]. Our observations are in agreement with *Driscoll and Karner* [1998, Figure 5], who argue that a landward dipping detachment is required from observations of an eastward decrease in subsidence from the continent-ocean boundary. The depth of the detachment is thermally determined, and shoals in the region of maximum heat input. At the point of rifting (the future continent-ocean boundary), upwelling of asthenosphere occurs, and the detachment will therefore dip towards both margins. *Driscoll and Karner*'s [1998, Figure 5] model suggests that the detachment will ramp down beneath the Australian continent, merging with the Moho at 40 km. This hypothesis could be tested with MT sites on the Pilbara and on the continental shelf.

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