2

Intraplate earthquakes in Australia

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Abstract

Relative to other intraplate areas of the world, Australia has a short recorded history of seismicity, spanning only a couple of centuries. As a consequence, there is significant uncertainty as to whether patterns evident in the contemporary seismic record are representative of the longer term, or constitute a bias resulting from the short sampling period. This problem can, in part, be overcome by validation against Australia's rich record of morphogenic earthquakes – Australia boasts arguably the richest late Neogene to Quaternary faulting record of any stable continental region. Long-term patterns in large earthquake occurrence, both temporal and spatial, can be deduced from the landscape record and used to inform contemporary earthquake hazard science. Seismicity source parameters such as large earthquake recurrence and magnitude vary across the Australian continent, and can be interpreted in a framework of large-scale neotectonic domains defined on the basis of geology and crustal setting. Temporal and spatial clustering of earthquakes is apparent at the scale of a single fault, and at the 1,000 km scale of a domain. The utility of the domains approach, which ties seismicity characteristics to crustal architecture and geology, is that behaviours can be extrapolated from well-characterised regions to poorly known analogous regions, both within Australia and worldwide.

2.1 Introduction

The Australian continent resides entirely within the Indo-Australian Plate, and is classified as a Stable Continental Region (SCR) in terms of its plate tectonic setting and seismicity (Johnston *et al.*, 1994; Schulte and Mooney, 2005). Such settings host approximately 0.2% of the world's seismic moment release (Johnston, 1994), and moderate to large earthquakes are rare. Analysis of focal mechanisms from SCR crust (Zoback, 1992;

Intraplate Earthquakes, ed. Pradeep Talwani. Published by Cambridge University Press. © Cambridge University Press 2014.

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Reinecker *et al*., 2003; Schulte and Mooney, 2005) indicates that compressive stress regimes dominate within continental interiors, with maximum compressive stresses oriented predominantly in accordance with absolute plate motion (see also Richardson, 1992). Australia is anomalous in this respect. While earthquake focal mechanisms suggest that the crustal stress field at seismogenic depths in Australia is everywhere compressive, with significant strike-slip components along the northwest margin (NWSSZ, Figure 2.1) and in the Flinders/Mount Lofty Ranges (FRSZ, Figure 2.1) (Leonard *et al*., 2002; Clark *et al*., 2003), stress orientations in southern Australia are typically not parallel to the north–northeast-directed plate motion vector (Coblentz *et al*., 1995; Hillis and Reynolds, 2000, 2003; Sandiford *et al*., 2004, 2005; Hillis *et al*., 2008).

In the southern half of the continent, the maximum horizontal stress orientation (S_{Hmax}) is essentially east–west in western and central Australia and rotates to northwest–southeast in eastern Australia (Figure 2.1). In the northern half of the continent, the stress field transitions from the generally east–west trend in the south, to a broadly northeast–southwest trend. To a first order these regional stress orientations are not influenced by tectonic terrane, crustal thickness, heat flow, regional structural trends, geological age, or by the depth at which orientations are sampled (e.g., Hillis *et al*., 2008; Sandiford and Quigley, 2009; Holford *et al*., 2011). The trends have been satisfactorily modelled in terms of a balance between plate driving and resisting torques generated at the margins of the Indo-Australian Plate (Cloetingh and Wortel, 1986; Coblentz *et al*., 1995, 1998; Reynolds *et al*., 2002; Burbidge, 2004; Sandiford *et al*., 2004; Dyksterhuis and Müller, 2008) (Figure 2.2).

Stratigraphic relationships establish that fault-related and presumably seismogenic deformation consistent with the present stress field commenced in the late Miocene, in the interval 10–6 Ma (Dickinson *et al*., 2002; Sandiford *et al*., 2004; Keep and Haig, 2010), as the result of complex evolving plate boundary conditions (Sandiford and Quigley, 2009). It has been proposed that the onset of deformation at specific locations may reflect rising stress levels related to the combination of all plate boundary forcings (Sandiford and Quigley, 2009), with variations in the thermal structure of the Australian lithosphere influencing the localisation of deformation at the regional or terrane scale (e.g., Celerier *et al*., 2005; Sandiford and Egholm, 2008; Holford *et al*., 2011). Analysis of the spatial and temporal patterns of contemporary seismicity might therefore be improved by studying the extraordinarily rich Neogene to Quaternary record of seismicity in the Australian landscape (e.g., Clark *et al*., 2012). In the sections that follow, earthquake occurrence in Australia is therefore examined from both the historic and prehistoric (Neogene and Quaternary) records.

2.2 Two centuries of earthquake observations in Australia

Between 1788, when European colonists reported the first earthquake felt in Australia (Historical Records of Australia, 1914), and the early 1900s, newspaper articles were the main source of information about earthquakes (e.g., Malpas, 1991), and continued to be important until the 1960s (McCue, 2004). The first seismographs in Australia for

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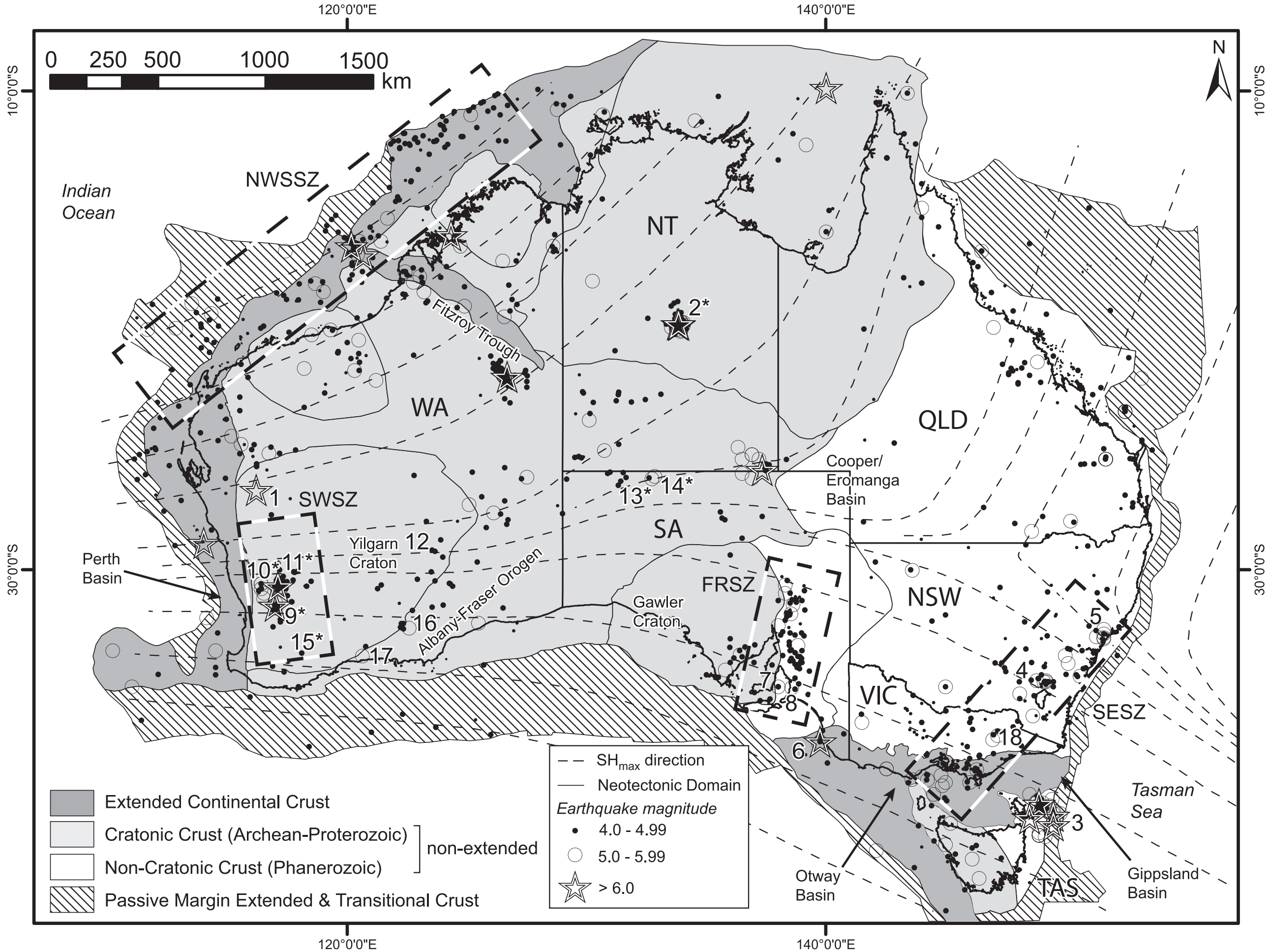

Figure 2.1 Map of the Australian continent plotting historical seismicity from the complete Australian catalogue for earthquakes $M \geq 4$. Zones of higher seismicity are shown: SWSZ, Southwest Seismic Zone; NWSSZ, Northwest Shelf Seismic Zone; FRSZ, Flinders Ranges Seismic Zone; SESZ, Southeast Seismic Zone. Significant earthquakes or earthquake sequences mentioned in text are indicated by numbers: 1, Meeberrie; 2, Tennant Creek; 3, offshore northeast Tasmania; 4, Dalton-Gunning; 5, Newcastle; 6, Beachport; 7, Warooka; 8, Adelaide; 9, Meckering; 10, Calingiri; 11, Cadoux; 12, Kalgoorlie-Boulder; 13, Ernabella; 14, Marryat Creek; 15, Katanning; 16, Norseman; 17, Ravensthorpe; 18, Mount Hotham. Numbers with asterisks indicate confirmed surface-rupturing earthquakes. The maximum horizontal stress (S_{Hmax}) direction is indicated by fine dashed lines. The continent is broadly divided on the basis of geology and tectonic history, showing the dominance of non-extended cratonic crust in the central and western parts, non-extended non-cratonic crust in the east, and extended crust around much of the continental margin.

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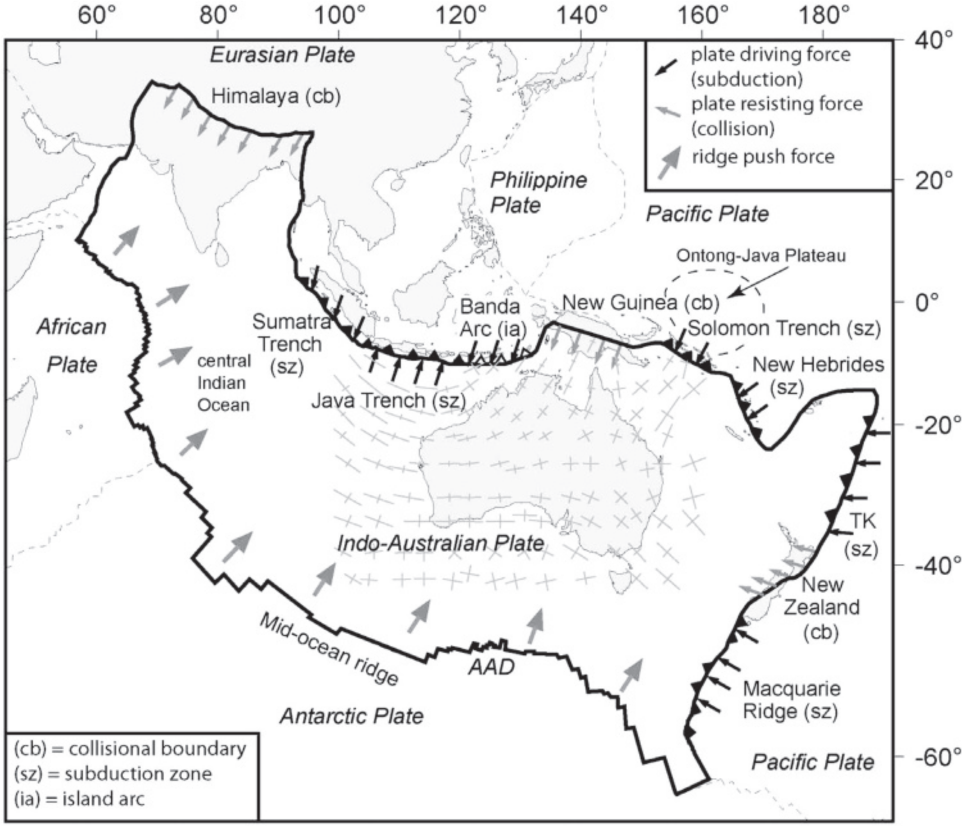

Figure 2.2 Indo-Australian Plate with plate boundary forces and orientation of modelled maximum and minimum horizontal stresses used in the finite element stress modelling of Reynolds *et al.* (2002). Much of the southern part of the continent has an east–west-oriented maximum horizontal compressive stress oriented at a high angle to the NNE-oriented plate velocity vector. Solid triangles indicate the direction of subduction and open triangles delineate the Banda Arc. TK, Tonga–Kermadec Trench; AAD, Australian–Antarctic discordance. (Modified after Reynolds *et al.*, 2002 and Hillis *et al.*, 2008.)

which records remain were four Milne seismographs established between 1901 and 1909 (Doyle and Underwood, 1965; McCue, 2004). The seismograph network remained sparse until 1962–3, after which time the Australian catalogue could be considered complete for magnitude (M) 5.0 and above earthquakes, and instrumental records largely replaced felt reports as the primary information source (McCue, 2004; Leonard, 2008). Doyle *et al.* (1968) summarised the history of earthquakes and seismological research in Australia up to 1966, and prepared a comprehensive reference list of historical earthquake sources. The authors tabulated over 160 Australian earthquakes, which formed the core of the preliminary digital Australian earthquake catalogue (Denham *et al.*, 1975). Despite enhancements resulting in the lowering of the completeness magnitude for the national catalogue to

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M 3.5+ by the 1980s, the Australian National Seismic Network (ANSN – http://www.ga.gov.au/earthquakes/seismicSearch.do) remains relatively sparse, with no more than 70 seismic stations suitable for general earthquake monitoring in a continent of a similar size to the conterminous United States or Western Europe (Leonard, 2008). This results in generally poor constraints being placed on the location of the more than 36,000 onshore earthquakes recorded in the current Australian earthquake catalogue (Allen *et al.*, 2012a).

Approximately 20 earthquakes of $M > 6.0$ have been recorded in Australian continental crust, at the rate of one every ~6–8 years (McCue, 1990; Leonard, 2008). On average, less than one $M > 5.0$ event has occurred per year (Leonard, 2008). The largest instrumentally recorded earthquake occurred in 1941 near Meeberrie, Western Australia (Everingham, 1982) (Figure 2.1 – 1). The event is associated with a surface wave magnitude of M_S 6.8 (Allen *et al.*, 2012a). In January 1988 three large "mainshock" earthquakes of magnitude M_S 6.3–6.7 (Choy and Bowman, 1990) occurred in a 12-hour period near Tennant Creek in the Northern Territory (Jones *et al.*, 1991) (Figure 2.1 – 2). The events were preceded by a year-long foreshock sequence, and followed by a continuing aftershock sequence (e.g., Bowman *et al.*, 1990). The most notable pre-instrumental earthquake sequence occurred in Bass Strait, off the northeast coast of Tasmania (Figure 2.1 – 3), in the 1880s–1890s. Over 2,400 events were felt during that period (Ripper, 1963), four of which caused damage in Tasmania, and two of which are estimated from felt effects to rival the Meeberrie earthquake in magnitude (Michael-Leiba, 1989). The Dalton–Gunning region of southeast Australia (Figure 2.1 – 4) continues to experience a less intense earthquake sequence (comprising more than 600 recorded events) that began with an M_L 5.6 event in the 1930s (Cleary, 1967; Denham *et al.*, 1981; Michael-Leiba *et al.*, 1994).

Earthquake clusters (e.g., Dalton–Gunning) are considered to be distinct from swarms, which are defined as occurring within a limited volume, lasting over a period from hours to months, with the largest event occurring well after the swarm commences, and not having a magnitude significantly greater than the second largest event (~0.5 magnitude unit; Gibson *et al.*, 1994; Dent, 2008). Geomechanical testing suggests that earthquake swarms occur preferentially in regions of extremely heterogeneous geological structure (Mogi, 1963). Increasing regional stress manifests as high stress concentration around numerous cracks and faults within the structured volume, resulting in failure on many local fractures at low stress. This has the effect of reducing the probability of failure on a single large fracture. Earthquake swarms are an important component of Australian seismicity, and can represent a large percentage of events in earthquake catalogues, particularly in regions of granitic geology (Dent, 2008). A non-exhaustive list of 42 earthquake swarms was compiled by Dent (2008, 2009), which includes perhaps the most significant of recent swarms, the 2001–2 Burakin Swarm (Leonard, 2002, 2003) in the Southwest Seismic Zone (SWSZ; Doyle, 1971) of Western Australia (Figure 2.1). This swarm involved six events in the magnitude range M_W 4.0–4.6 (Allen *et al.*, 2006) with over 18,000 smaller events, all purportedly occurring within a volume of ~5 km diameter (Leonard, 2003). The centre of activity for the swarm occurs approximately 20 km north of the surface rupture relating to the 1979 Cadoux M_S 6.0 earthquake (Figure 2.1 – 11). Several other swarms in the area (Kalannie, Beacon)

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Table 2.1 *Documented surface-rupturing earthquakes in Australia. Refer to Figure 2.1 for locations.*

Year	Location	Magnitude (M_W)	Reference(s)
1968	Meckering, WA	6.6	Gordon and Lewis (1980), Vogfjörd and Langston (1987), Johnston (1994)
1970	Calingiri, WA	5.5	Gordon and Lewis (1980), Johnston (1994)
1979	Cadoux, WA	6.1	Lewis *et al.* (1981), Fredrich *et al.* (1988)
1986	Marryat Creek, SA	5.8	Fredrich *et al.* (1988), Machette *et al.* (1993)
1988	Tennant Creek, NT*	6.3–6.6	Bowman *et al.* (1990), Chung *et al.* (1992), Crone *et al.* (1992)
2008	Katanning, WA	4.7	Dawson *et al.* (2008)
2012	Ernabella, SA	5.4	Clark *et al.* (2013)

* Tennant Creek involved a series of three consecutive events within one day.

suggest that the events might reflect continuing crustal adjustment relating to the Cadoux rupture (Dent, 2008, 2009). However, many other swarm sequences are apparently unrelated to larger events (Gibson *et al.*, 1994; Dent, 2008).

On 27 December 1989 an M_L 5.6 earthquake occurred near Newcastle, NSW (Figure 2.1 – 5), resulting in 13 fatalities and significant damage. The event was one of the most destructive and costly natural disasters to have occurred in Australia (McCue *et al.*, 1990; Sinadinovski *et al.*, 2002; Insurance Council of Australia, 2013), but not the first event to affect a populated centre. The 1897 Beachport, 1902 Warooka, and 1954 Adelaide events (McCue, 1975; Greenhalgh and Singh, 1988; Love, 1996) (Figure 2.1 – 6, 7, 8) caused damage in South Australia; the Warooka event also claimed two lives (McCue, 1975). The 1968 Meckering, 1970 Calingiri (Gordon and Lewis, 1980), 1979 Cadoux (Lewis *et al.*, 1981), and 2010 Kalgoorlie–Boulder (Bathgate *et al.*, 2010) (Figure 2.1 – 9, 10, 11, 12) events all caused damage in Western Australia, as did the 1988 Tennant Creek sequence in the Northern Territory (Bowman, 1992) (Figure 2.1 – 2).

Several of the damaging earthquakes noted previously also ruptured the ground surface. If the three 1988 Tennant Creek mainshocks (Bowman, 1988; Choy and Bowman, 1990; Bowman, 1992; Crone *et al.*, 1992) are treated as a single scarp-forming event, then the 23 March 2012 M_W 5.4 Ernabella earthquake (Clark *et al.*, 2013) (Figure 2.1 – 13) represents the seventh historic Australian event that can be unequivocally associated with surface rupture (Figure 2.1; Table 2.1). All seven of these events ruptured non-extended SCR cratonic crust (Figure 2.1) and involved a dominant reverse component to motion (Gordon and Lewis, 1980; Lewis *et al.*, 1981; Crone *et al.*, 1992; Machette *et al.*, 1993; Dawson *et al.*, 2008). As such, they are similar to the 1989 Ungava (Canada) (Adams *et al.*, 1991, 1992; Bent, 1994) and the 1993 Killari Latur (India) (Rajendran *et al.*, 1996, 2001; Seeber *et al.*, 1996; Rajendran, 2000) earthquakes.

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In addition to the few documented surface-rupturing events, isoseismal maps have been compiled for almost 400 historic Australian earthquakes (Everingham, 1982; Rynn *et al.*, 1987; McCue, 1996). McCue (1980) derived a relationship between Richter magnitude (M) and the radius of a circle of equivalent area to the Modified Mercalli Intensity III isoseismal for Australian earthquakes with recorded magnitudes ranging from 3.6 to 7.0 (reproduced in McCue, 2004). This permitted magnitudes to be assigned for pre-instrumental earthquakes or recent earthquakes that were widely felt but either too close to or too far from the nearest seismograph to measure the magnitude (McCue, 2004).

2.2.1 Mechanism, geographic distribution, and strain rate

The Australian crustal stress regime is generally considered to be compressive (Denham *et al.*, 1981; Hillis and Reynolds, 2000, 2003). This assessment is supported by the majority of earthquake focal mechanisms, which range from thrust to oblique strike-slip (e.g., Leonard *et al.*, 2002; Keep *et al.*, 2012). Notable exceptions, with a dominant normal component, are the 1985 Norseman and 2001 Ravensthorpe earthquakes (Figure 2.1 – 16, 17), both located along the Albany–Fraser Orogen margin of the Yilgarn Craton, with the latter being anomalously deep at 18 km (McCue, 1989; Clark, 2004). The 1966 Mount Hotham earthquake in the SESZ (Denham *et al.*, 1982) (Figure 2.1 – 18), several of the Burakin Swarm events in the SWSZ (Leonard *et al.*, 2002), and two Tennant Creek aftershocks (Clark and Leonard, 2003) also have a dominant normal component to their focal mechanisms. Note that Leonard *et al.* (2002) incorrectly transposes the P and T axes, and dilatant and compressional quadrants, for the Norseman 1985 event (cf. McCue, 1989).

Assuming a maximum possible magnitude ($M_{\max}$) of ~7.0 (refer also to Section 2.5), the maximum seismogenic strain rate estimates for the continent (averaged over the last ~50 years of complete data) are $\sim 10^{-17}$ to 10^{-16} s^{-1} (Leonard, 2008; Braun *et al.*, 2009; Sandiford and Quigley, 2009). Comparable results (i.e., 10^{-17} s^{-1}) are obtained from thin-plate finite element modelling using plate boundary conditions, heat flow, stress, and geodetic data as inputs (Burbidge, 2004). This has been equated to an east–west shortening rate across southern Australia of approximately 0.3–0.4 mm/yr (Leonard, 2008), which compares to estimates based upon laser ranging and geodetic GPS of 0.65–3.0 ± 2.0 mm/yr (Smith and Kolenkiewicz, 1990; Tregonning, 2003; Leonard, 2008).

However, earthquake epicentres are not randomly distributed across the Australian continent in time or space (Denham, 1988; Leonard, 2008; Sinadinovski and McCue, 2010). At the continental scale, concentrations of epicentres occur in four major "seismogenic zones" (Figure 2.1) (Hillis *et al.*, 2008; Leonard, 2008; Sandiford and Egholm, 2008), with the continental margins also demonstrating a comparatively high number of epicentres relative to the interior (Sandiford and Egholm, 2008). Strain rates calculated for the higher seismicity zones are no more than 10^{-16} to 10^{-15} s^{-1} (Leonard, 2008; Braun *et al.*, 2009). Sandiford and Quigley (2009) provide an example from the Flinders Ranges Seismic Zone (FRSZ, Figure 2.1), where the bulk strain rate of 10^{-16} s^{-1} implies a total shortening of

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~250 m/Ma across the ~100 km wide zone, which could be accommodated by 10 faults with slip rates consistent with the current topography and neotectonic faulting record (e.g., Sandiford, 2003b; Quigley *et al.*, 2006; Clark *et al.*, 2011a). Braun *et al.* (2009) estimate that half of the current relief of the Flinders Ranges, and a non-negligible proportion of the relief in the Eastern Highlands (SESZ, Figure 2.1), might plausibly have been built since the inception of the current stress regime. In contrast, similar seismogenic strain rates calculated for the SWSZ (Leonard, 2008; Braun *et al.*, 2009) (Figure 2.1) are not consistent with the present low-relief landscape (Clark, 2010), the distribution of paleo-earthquake fault scarps (Leonard, 2008; Leonard and Clark, 2011), or GPS strain rates (Leonard *et al.*, 2007). This implies migration of the locus of seismicity with time (Leonard, 2008; Leonard and Clark, 2011), an assertion supported by evidence suggesting that the rate of seismic activity in the SWSZ has increased dramatically in only the last 50 years (Michael-Leiba, 1987; Leonard and Clark, 2011).

2.2.2 Seismogenic depth

Depth constraint for Australian earthquakes is generally poor. Leonard (2008) selected a subset of the Australian earthquake catalogue for detailed hypocentral depth analysis using only depth values whose uncertainty was either less than the depth itself, or less than 5 km. More than 75% of epicentres from the current earthquake catalogue are rejected using these criteria. The selected data correlate with the regions of highest station density in private networks, and in the ANSN (i.e., southern Australia).

The concentration of epicentres in the SWSZ (Figure 2.1) occurs in non-extended cratonic crust. The Leonard (2008) dataset shows that earthquakes in this region are typically very shallow (see also Figure 2.3), with 95% of events found to have occurred within the upper 5 km of crust, consistent with previous analyses (Doyle, 1971; Everingham and Smith, 1979; Denham, 1988; McCue, 1990). This region has produced four surface-rupturing earthquakes in the last five decades (Figure 2.1; Table 2.1). The largest of these earthquakes (1968 Meckering; Figure 2.1 – 9) is calculated to have initiated at 1.5 km depth and to have ruptured both upwards to the surface and down to ~6 km depth (Langston, 1987; Fredrich *et al.*, 1988), while the smallest (2008 Katanning; Figure 2.1 – 15) ruptured from ~0.6 km depth to the surface (Dawson *et al.*, 2008). Earthquake swarms are also particularly prevalent in the SWSZ, with well-located depths being in the upper 2–3 km (Leonard, 2002, 2003). Analysis of the characteristics of modern and prehistoric fault scarps (Clark, 2010; Leonard and Clark, 2011; Clark *et al.*, 2012) suggests that large earthquake characteristics in the SWSZ (e.g., shallow depth) are typical of what might be expected throughout non-extended cratonic crust in Australia. However, Australia's deepest known earthquakes are also from this crustal setting: the 1989 Uluru earthquake (Figure 2.3 – 1), with a calculated depth of 31 km (Michael-Leiba *et al.*, 1994), and a 1992 earthquake beneath the Arafura Sea, north of the Northern Territory coastline (Figure 2.3 – 2), which had a depth of 39 km (McCue and Michael-Leiba, 1993).

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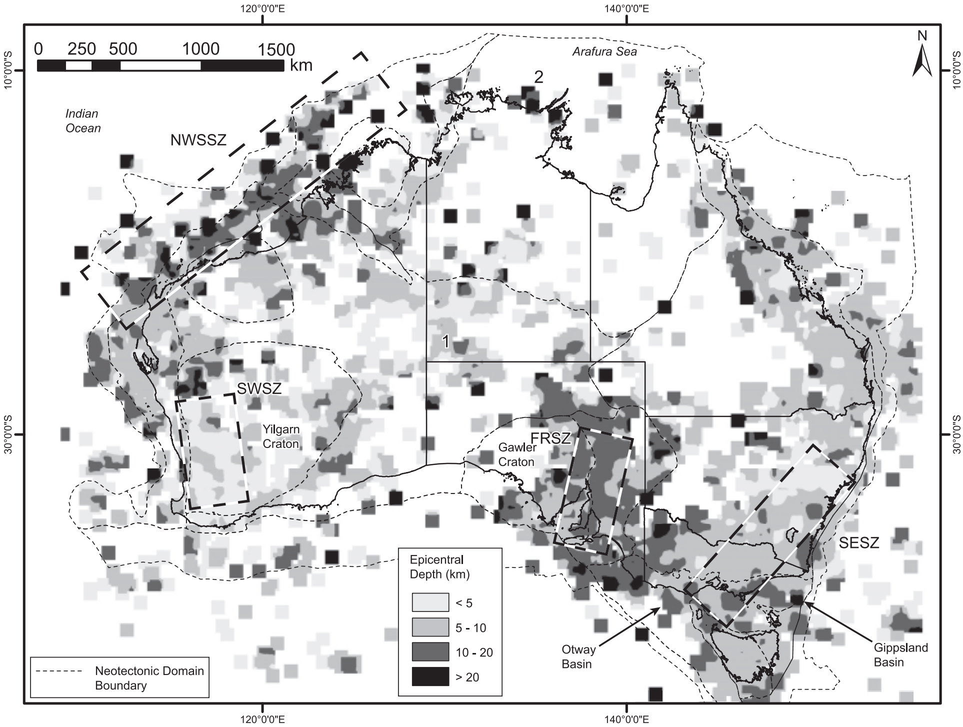

Figure 2.3 Interpolated surface of variation in earthquake hypocentral depths across the Australian continent constructed using the ARCGIS Point Statistics tool (plotted value is the mean of values within a moving 3 degree search box, output cell size 0.25 degrees). Events with a catalogued depth of greater than 50 km are not considered. The earthquake catalogue used is from Allen *et al*. (2012a), and is not filtered according to hypocentral uncertainty (cf. Leonard, 2008). Higher seismicity zones are shown (cf. Figure 2.1). Locations referred to in text are marked. Significant earthquakes mentioned in text are indicated by numbers: 1, Uluru; 2, Banda Sea. Note the greater depth of events in the FRSZ and NWSSZ compared to the SWSZ and SESZ.

Despite being one of the most seismogenic regions of Australia (Leonard, 2008; Sandiford and Egholm, 2008), the Northwest Shelf Seismic Zone (NWSSZ – Figure 2.1) is characterized by extremely poorly constrained hypocentres, with depth uncertainties typically significantly larger than the depth values (e.g., Revets *et al*., 2009; Allen *et al*., 2012a). Consequently, this region was not analysed in detail by Leonard (2008). However, earthquake hypocentres tend to be deeper in this region of extended continental crust than in the adjoining non-extended cratonic crust (Figure 2.2). Approximately 40% of events recorded in the NWSSZ occur at depths greater than 10 km.

The FRSZ in South Australia consistently produces deeper earthquakes than other parts of the continent (Figure 2.3), with over 75% of well-constrained events occurring at depths greater than 10 km (Leonard, 2008). A temporary seismograph deployment (September to December 2003) suggested spatial variation in the depth distribution, with a predominance

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of hypocentres in the 10–20 km depth range in the central Flinders Ranges, and less than 10 km depth in the southern Flinders Ranges (Cummins *et al*., 2009), consistent with previous work (Greenhalgh *et al*., 1994). The high proportion of hypocentres with depths greater than 10 km suggests that the depth to brittle–ductile transition in the Flinders Ranges is deeper than might be inferred from heat flow data (cf. Celerier *et al*., 2005; Holford *et al*., 2011).

Within the Southeast Seismic Zone (SESZ – Figure 2.3), and non-cratonic eastern Australia in general, hypocentral depth estimates are bimodal (Close and Seeber, 2007), ranging between very shallow (<5 km; 30% of data) and mid-crustal depths (10–20 km; 40% of data) (Allen *et al*., 2012). Aftershocks tend to be very shallow and numerous, perhaps accounting for the shallower mode (Gibson *et al*., 1981; Leonard, 2008). Within a sub-zone that encompasses the Pliocene to Holocene Newer Volcanic Province (e.g., Sutton *et al*., 1977; Sheard, 1995), mid-crustal earthquakes are generally deeper, with 95% of recorded events occurring in the range 9–17 km (Leonard, 2008).

The Otway and Gippsland basins of southeast Australia (Figure 2.3) form part of an aulacogen developed in non-cratonic Paleozoic crust that was extended by rifting in the late Mesozoic–early Cenozoic. Earthquake hypocentres are deep in this region compared to those in the non-extended parts of the same Paleozoic basement province that is more typical of the SESZ. More than 70% of hypocentres in the extended basins are in the 10–25 km depth range (Allen *et al*., 2012a), exemplified by the well-located 19 June 2012 M_L 5.4 Moe earthquake at a depth of 17 km (Sandiford and Gibson, 2012). It is probable that the contribution from these basins accounts for the greater depth element present in the southern southeast Australia (SEA-S) zone of Leonard (2008), as he does not distinguish between these two geological settings.

2.2.3 Attenuation and scaling relations

Several studies have shown that earthquakes in SCRs, such as Australia, are generally felt at larger distances than earthquakes in active tectonic regions (McCue, 1990; Frankel, 1994; Bakun and McGarr, 2002; Atkinson and Wald, 2007; Wald *et al*., 2011). This is because the seismic energy propagates more efficiently through cold, relatively homogeneous continental crust, which is less susceptible to anelastic and scattering effects. It is recognised from analysis of isoseismal radii (Gaull *et al*., 1990) that attenuation of seismic wave energy varies transversely across the Australian continent, with relatively low attenuation in the Archaean and Proterozoic terranes of western and central Australia and higher attenuation in the younger Phanerozoic terranes of eastern Australia (cf. Figure 2.1). Moreover, it is also observed that ground-motion amplitudes at large distances (>100 km) for an earthquake of given magnitude are lower in southeast Australia (SESZ) than in eastern North America (ENA) based on macroseismic intensities (Bakun and McGarr, 2002) and instrumental data (Allen and Atkinson, 2007). The higher attenuation observed in SESZ at distances greater than ~100 km is likely to be due to the broad crustal velocity

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gradient (Collins *et al*., 2003), which allows dispersion of Lg-wave energy into the upper mantle (e.g., Bowman and Kennett, 1991; Atkinson and Mereu, 1992). However, there appears to be no discernible difference between the attenuation of Australian and ENA ground motions in the distance range of engineering significance (i.e., <100 km; Allen and Atkinson, 2007; Allen, 2012). Furthermore, attenuation rates at these near-source distances are also comparable to those in active tectonic regions (e.g., Hanks and Johnston, 1992; Atkinson and Morrison, 2009; Campbell, 2011).

Rupture dimensions have been modelled for only two of Australia's large earthquake sequences – the 1968 Meckering earthquake and the 1988 Tennant Creek sequence (Somerville *et al*., 2010) – both of which occurred in the cratonic western and central regions of Australia respectively (cf. Figure 2.1; Table 2.1). Significant simplifications were required to model the subsurface rupture geometries of both events, which involved complex surface rupture on intersecting structural elements of varying orientation (e.g., Crone *et al*., 1992; Dentith *et al*., 2009). While this might limit the confidence that can be placed in the rupture models, earthquake scaling relations developed from the models compare favourably with empirically derived relations for intraplate dip-slip earthquakes (Johnston, 1994), and self-similar scaling relations (Leonard, 2010). Following the 2012 Ernabella earthquake, updated relations between seismic moment and surface rupture length were developed, and suggest that earthquakes in nonextended cratonic Australia produce longer surface ruptures than is estimated from published scaling relations for earthquakes larger than $M_w \sim 6{:}5$ (Clark *et al*., 2013). At present there are no estimates of the rupture dimensions of any earthquakes in the non-cratonic or extended crust regions of eastern Australia.

2.3 A long-term landscape record of large (morphogenic) earthquakes

Australia is one of the lowest, flattest, most arid and slowly eroding continents on Earth. Accordingly, large parts of Australia are favourable for the preservation of tectono-geomorphic features, such as fault scarps, for tens of thousands to millions of years (e.g., Quigley *et al*., 2010). In regions with extremely low erosion rates, such as the Nullarbor Plain (Figure 2.4), it has been claimed that a morphogenic earthquake record spanning the last 15 Ma has been preserved essentially intact (Hillis *et al*., 2008). On this basis, Australia boasts arguably the richest Late Neogene to Quaternary faulting record in all of the world's SCR crust (Sandiford, 2003b; Quigley *et al*., 2010; Clark *et al*., 2011a, 2012). Over 300 features (mainly fault scarps and folds) suspected or known to have been displaced under the current crustal stress regime have been identified and recorded (Clark *et al*., 2011a, 2012) (Figure 2.4). The majority of these features, by virtue of their length and/or vertical displacement (see Section 3.1), are likely to reflect multiple surface-rupturing events (cf. Leonard, 2010). This remarkable archive, while undoubtedly incomplete, has the potential to extend the historic record of seismicity to a timescale commensurate with the recurrence time of large earthquakes in this intraplate setting.

Variation in fault scarp length, vertical displacement, proximity to other faults, and relationship to topography permits further division of this neotectonic record according

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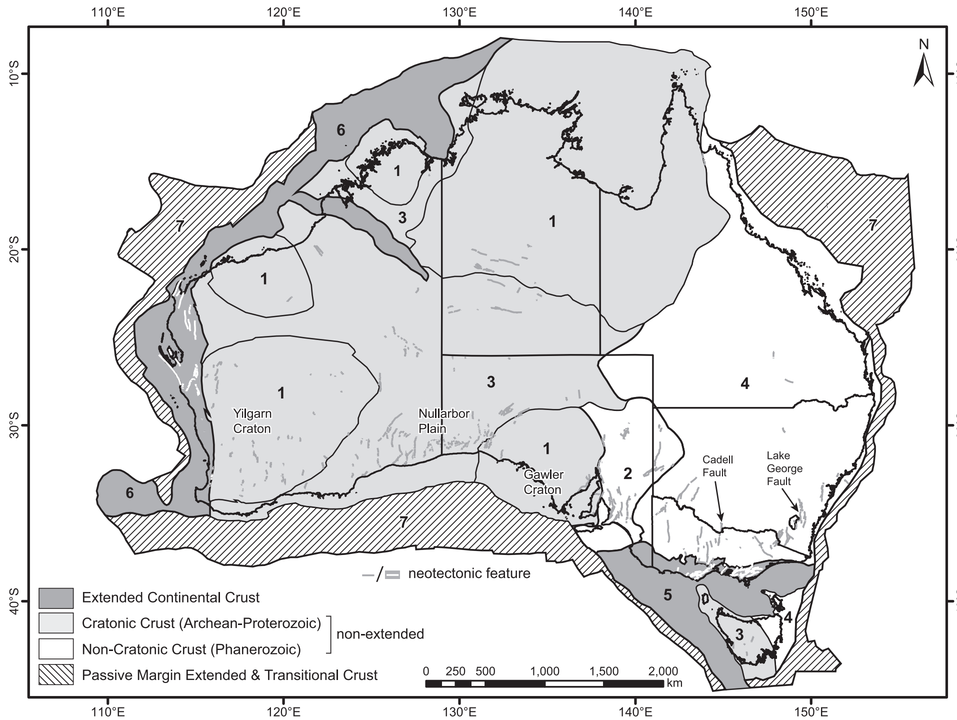

Figure 2.4 Neotectonic domains map of Australia (after Clark *et al.*, 2011a, 2012) showing features known or suspected of hosting displacement under the current stress field (grey lines). Neotectonic domains are indicated by numbers: 1, Precambrian Craton and Non-reactivated Proterozoic Crust; 2, Sprigg Orogen; 3, Reactivated Proterozoic Crust; 4, Eastern Australian Phanerozoic Accretionary Terranes; 5 Eastern Extended Continental Crust; 6, Western Extended Continental Crust; 7, Passive Margin Extended Continental and Transitional Crust. Shading and hatching as per Figure 2.1 shows the domains in their broader context of non-extended cratonic crust in the central and western parts, non-extended non-cratonic crust in the east, and extended margins around much of the continent.

to fault character. Six onshore "neotectonic domains" are recognised, with an additional offshore domain proposed by analogy with the eastern United States (Clark *et al.*, 2011a, 2012) (Figure 2.4). Each domain relates to a distinct underlying crustal type and architecture, broadly considered to represent non-extended cratonic and non-cratonic crust, as well as extended crustal settings (Figure 2.4) (cf. Johnston *et al.*, 1994). Herein domains will be referred to using the abbreviation D*x*, where *x* is the number of the domain (e.g., D1 is Domain 1).

The Australian landscape reveals a marked disparity in its crustal response to imposed tectonic stresses. For example, neotectonic faults in western and central Australia, which are founded in cratonic crust (Figure 2.4 – D1, D3), are typically spaced more than a scarp length apart, not associated with historic seismicity, and displace a low, undulating landscape by less than 10 m (e.g., Clark, 2010). In contrast, faults in non-cratonic crust of the Mount Lofty and Flinders Ranges (Figure 2.4 – D2) are relatively closely spaced, commonly associated

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with historic seismicity, and displace Neogene strata by up to a few hundred metres (e.g., Sandiford, 2003b; Quigley *et al.*, 2006). Further east, within non-cratonic Australia (Figure 2.4 – D4), faults are commonly found in *en echelon* arrangements, with fault complexes extending for several hundred kilometres along strike (Beavis, 1960, 1962; Moye *et al.*, 1963). As little as 15 m (e.g., Canavan, 1988; Robson and Webb, 2011; McPherson *et al.*, 2012b) and as much as several hundred metres of neotectonic displacement has been documented on several of these features (Beavis, 1960, 1962; Moye *et al.*, 1963; Abell, 1985), and there is no clear relationship to historic seismicity. Faults within the extended continental crust fringing the Australian continental margin (Figure 2.4 – D5, D6), a remnant of the Cretaceous breakup of the supercontinent Gondwana (e.g., Veevers, 2000), have some of the larger throws of any Australian neotectonic faults (greater than 100 m; Holdgate *et al.*, 2003). Extensional structural architecture is largely preserved (e.g., Hill *et al.*, 1994), and neotectonic faults are often spatially associated with earthquake epicentres.

2.3.1 Variation in fault scarp length and vertical displacement

The neotectonic data compiled by Clark *et al.* (2012) contain two semi-quantitative variables useful for characterising fault behaviour – length and vertical displacement. The population distributions for Australian fault scarp length and vertical displacement data are presented in Figure 2.5a, b. Fault length is defined as the along-strike distance (tail to tail) of discrete geomorphic features (most often fault scarps) that are considered to represent one or more surface-rupturing earthquake events. Vertical displacement is the vertical separation across a topographic scarp or fold. Many of the fault-length values reported by Clark *et al.* (2012) might be expected to be underestimates, as vertical displacement tapers towards the tails of ruptures, resulting in these scarp sections being less discoverable in digital elevation data: the primary tool for their identification and characterisation (Clark, 2010; Clark *et al.*, 2011a, 2012). It is also reasonable to assume that relatively short scarps are under-represented as they are less discoverable. This factor may explain the positive skew in both length and height data (Figure 2.5a, b). Furthermore, the resolution and noise content of digital elevation data from various sources might be expected to affect the precision of both length and vertical displacement measurements.

Interpolated surfaces for Australian neotectonic data demonstrate the spatial variation in these fault parameters (Figure 2.5c, d), and these spatial patterns are borne out in the statistics for each of the neotectonic domains (Figure 2.5e, f). The cratonic domains (D1, D3) are characterised by the lowest vertical displacement values (Figure 2.5d, f). In view of the extremely low erosion rates in these parts of Australia (Bierman and Caffee, 2002; Belton *et al.*, 2004; Chappell, 2006), this is indicative of very low time-averaged rates of morphogenic seismicity. Scarp lengths in D3 are up to 100 km greater than in D1 (Figure 2.5e), raising the possibility that relief in the Proterozoic mobile belts is built in fewer, larger earthquakes. Five fault scarps have been subject to detailed paleoseismological investigation in this crustal type: (1) Meckering (Clark *et al.*, 2011a), (2) Hyden (Crone

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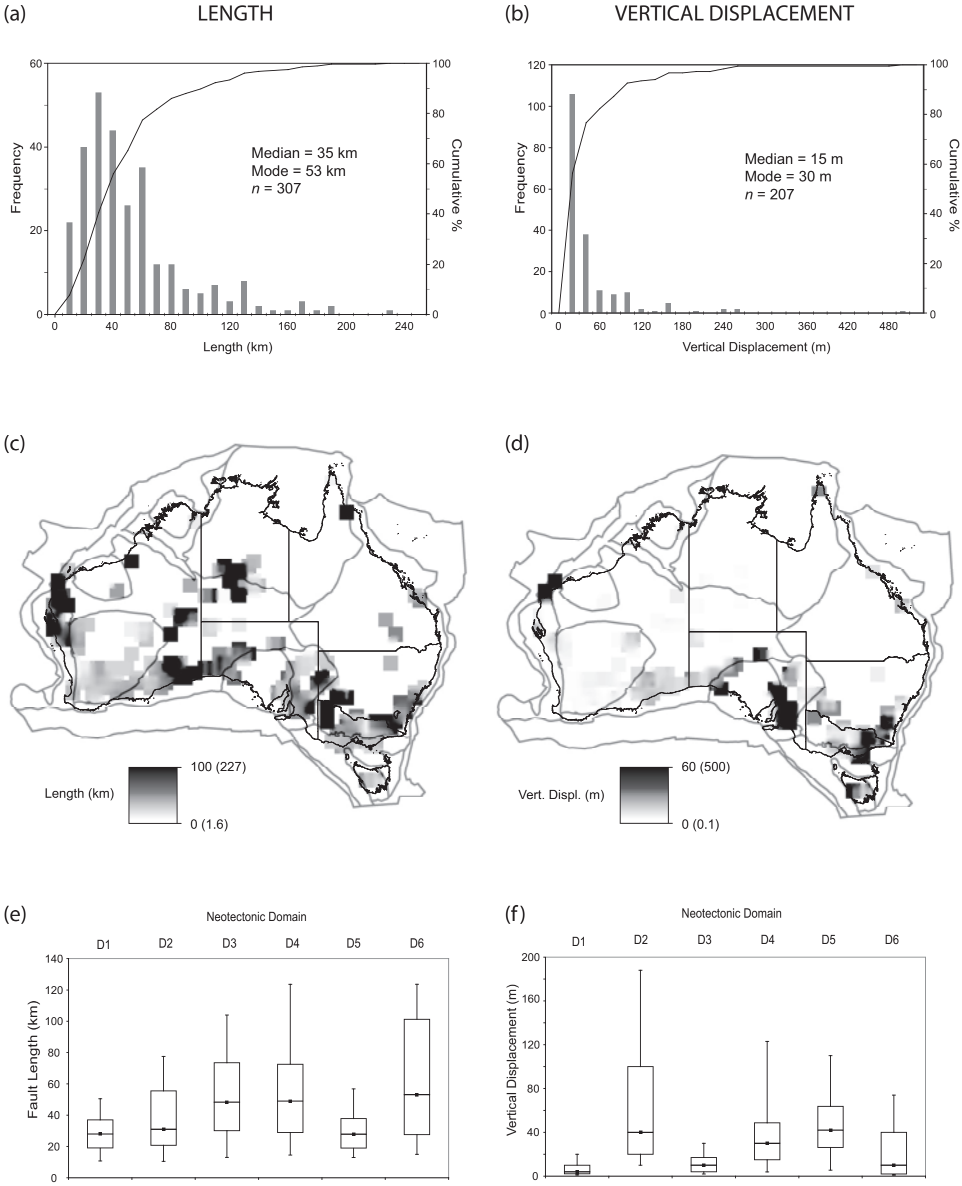

	D1	D2	D3	D4	D5	D6
min	2.7	2.8	1.6	6.9	2.2	6
max	127	172	227	186	81.5	170
med	26	31	49	49	28	53
n	69	48	71	49	44	26

	D1	D2	D3	D4	D5	D6
min	0.15	4	0.1	2.5	0.1	0.2
max	60	250	80	500	240	100
med	5	40	10	30	44	10
n	52	34	53	30	25	13

Figure 2.5 Population distributions for Australian neotectonic (a) fault length and (b) vertical displacement data, with accompanying plots showing the spatial distribution of (c) fault length and (d) vertical displacement data across the Australian continent. Interpolated surfaces in parts (a) and (b) are constructed using the same method as Figure 2.3. Grey lines indicate neotectonic domain boundaries (cf. Figure 2.4). Data values have been scaled for gridding – values in parentheses in the legend indicate the full range of underlying data. Box and whisker plots of (e) fault length and (f) vertical displacement data, binned by neotectonic domain. Boxes denote 75th and 25th percentiles, central point indicates median value, and whiskers define 90th and 10th percentiles. Data table shows minimum, maximum, median, and number of data points (*n*) for each domain.

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et al., 2003; Clark *et al*., 2008), (3) Dumbleyung and (4) Lort River (Estrada, 2009), and (5) Lake Edgar (Clark *et al*., 2011b). These features range from 30 to 40 km in length (cf. Clark, 2010) with maximum single-event vertical displacements in the order of 1.2–3.1 m. Recurrence data for morphogenic earthquake events is restricted to the most recent one to three events on each fault, and indicates inter-event intervals of ∼10–40 ka (Figure 2.6).

Non-cratonic (D2, D4) and extended (D5, D6) domains are characterised by comparatively large vertical displacements (Figure 2.5d, f), which scaling relationships (e.g., Wells and Coppersmith, 1994; Leonard, 2010; Leonard and Clark, 2011) imply must have built as the result of multiple seismic cycles. There is, therefore, less certainty as to whether scarp lengths are representative of single-event ruptures, or are the product of segmented rupture. A positively skewed fault-length data distribution (Figure 2.5a) might plausibly reflect segmented rupture behaviour. The longest fault scarp in non-cratonic eastern Australia (D4) that has been subject to paleoseismological investigation is the Cadell Fault (Figure 2.4). Evidence from abandoned fluvial and tectonic landforms (e.g., Bowler and Harford, 1966; Rutherfurd and Kenyon, 2005) is consistent with seismic rupture of the entire 80 km scarp length, potentially involving 2–4 m of uplift (Clark *et al*., 2011a; McPherson *et al*., 2012a). While the timing of individual seismic events is poorly constrained, the average recurrence interval within the 70–20 ka active period (assuming full-length rupture) may have been as little as ∼8 ka. Within D2, single-event displacement values of 1.8 m (1.5 m vertical) have been recorded on the Alma and Williamstown-Meadows faults (Clark and McPherson, 2011). The latter fault, which has a mapped length of over 100 km, is associated with a 25 km long single-event scarp.

2.3.2 The influence of crustal type and character on seismic activity rates

Given that the compressive nature of the Australian stress field (Hillis and Reynolds, 2003) results in a predominance of dip-slip faulting (Leonard *et al*., 2002), long-term qualitative seismic activity rates across the continent might be assessed in terms of neotectonic uplift of the landscape. With respect to the morphogenic earthquake record, this is a function of neotectonic fault slip rate and density.

Long-term slip rates estimated for faults in the cratonic western part of the continent (D1, D3) are typically in the order of ∼1 m/Ma (Clark *et al*., 2008, 2012; Hillis *et al*., 2008), which is equal to or less than the extant erosion rates (Bierman and Caffee, 2002; Belton *et al*., 2004; Chappell, 2006). These long-term slip rates are consistent with the low-relief landscape, and contrast with uplift rates estimated from the last several decades of seismicity in the SWSZ, which have been suggested to be in the order of 10 m/Ma (Braun *et al*., 2009).

Long-term slip rates in non-extended, non-cratonic eastern Australia (D4) are less well constrained. The slip rate on the Cadell Fault, averaged over the life of the current stress field (i.e., ∼10 Ma), has been estimated at ∼10 m/Ma (Clark *et al*., 2007). New data from the Lake George Fault (Pillans, 2012) within the Eastern Highlands (Figure 2.4) indicate a

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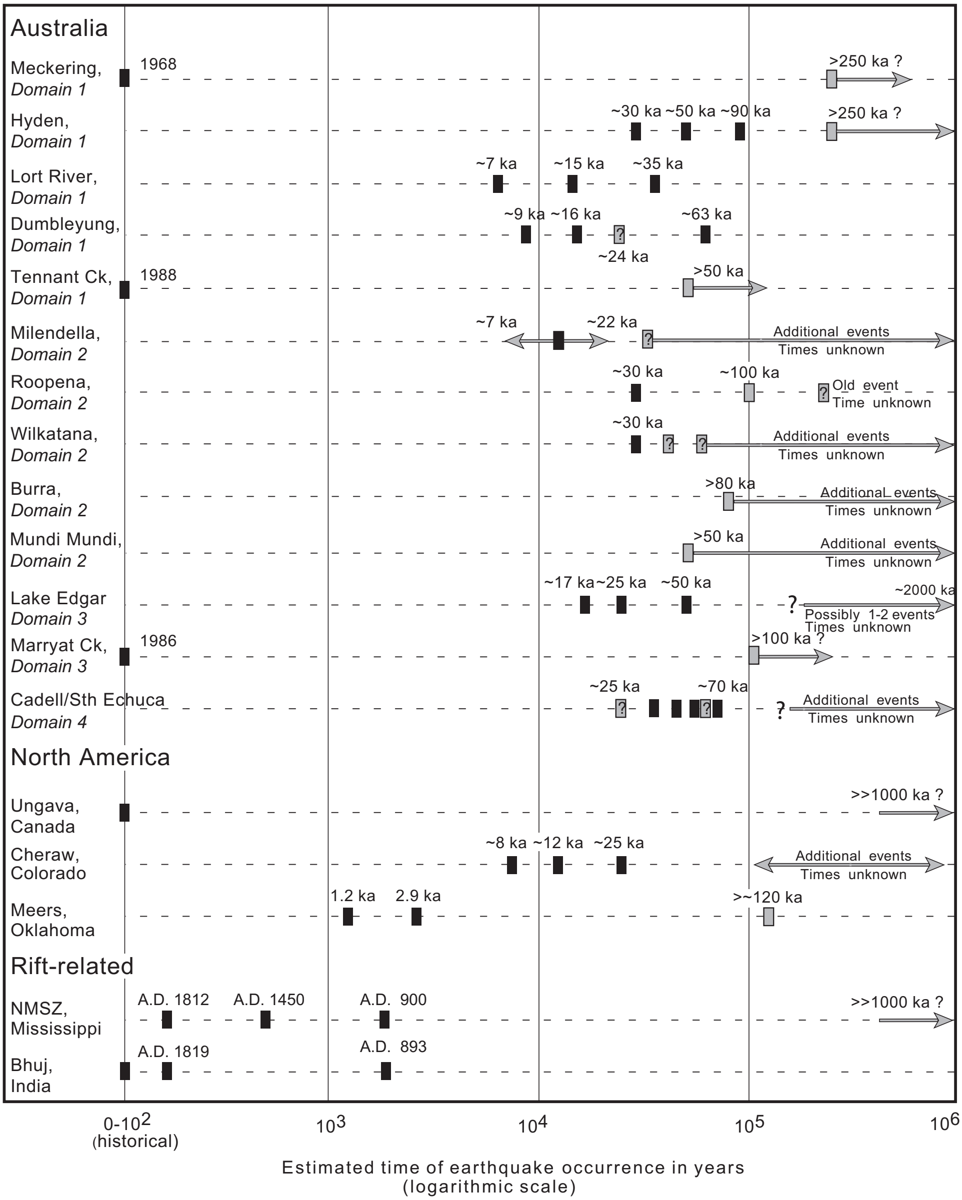

Figure 2.6 Compilation of surface-breaking earthquake recurrence data for SCR settings (Clark *et al*., 2012, modified after Crone *et al*., 2003). Lort River and Dumbleyung data from Estrada (2009), Wilkatana, Burra and Mundi Mundi data from Quigley *et al*. (2006), Lake Edgar data from Clark *et al*. (2011b) and Cadell data from Clark *et al*. (2007). Australian examples are labelled with their relevant neotectonic domain (cf. Figure 2.4).

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slip rate of ~50 m/Ma. Higher relief areas of eastern Australia are associated with erosion rates of up to 30–50 m/Ma (Weissel and Seidl, 1998; Heimsath *et al*., 2000, 2001; Wilkinson *et al*., 2005; Tomkins *et al*., 2007). Several authors have suggested that as much as 200 m of relief has been added to the Eastern Highlands over the last ~10 Ma (Sandiford, 2003b; Holdgate *et al*., 2008; Braun *et al*., 2009). Such estimates are consistent with uplift rates estimated from strain rates derived from contemporary seismicity (Braun *et al*., 2009); however, little unequivocal evidence exists to assign this uplift to active faults (cf. Holdgate *et al*., 2006). The approximate equivalence of erosion rates and fault slip rates suggests that much of the extant relief may be inherited (e.g., Bishop *et al*., 1982; Pickett and Bishop, 1992; van der Beek *et al*., 2001).

Perhaps the most spectacular examples of neotectonism in southeast Australia are the inverted Mesozoic Otway and Gippsland Basins (D5) (see Figure 2.1) (Holdgate *et al*., 2003, 2007; Sandiford, 2003a), which formed by extension of non-cratonic (D4) crust. Within the Gippsland Basin in particular, the Cretaceous basin deeps now form the topographic highs at elevations of 200–300 m above sea level. Preliminary cosmogenic radionuclide ages obtained on overlying folded alluvial sediments (Holdgate *et al*., 2007) suggest uplift rates of 60 m/Ma and greater on the major relief-forming faults (McPherson *et al*., 2009; Clark *et al*., 2011a, 2012). The relatively high fault density, combined with high slip rates and frequent contemporary seismicity (e.g., Leonard, 2008), identify these basins as among the most actively deforming parts of the continent (compare with the Flinders Ranges – D2).

Slip rates on faults underlying folds in the passive margin basins that dominate the Northwest Shelf region (D6) (NWSSZ, Figure 2.1), are poorly constrained. While locally having resulted in the uplift of Miocene marine deposits to elevations of over 100 m above sea level (e.g., van de Graaff *et al*., 1976), neotectonic uplift is typically more modest than in the eastern aulacogen and passive margin basins (e.g., D5).

Long-term (i.e., averaged over several million years) vertical slip rates are known from some of the range-bounding faults of the Flinders and Mount Lofty Ranges (D2). These typically vary between ~20 and 50 m/Ma (Bourman *et al*., 1999; Belperio *et al*., 2002; Sandiford, 2003b; Celerier *et al*., 2005; Quigley *et al*., 2006). Bedrock erosion rates have been recorded at up to 122 m/Ma, but average around 40 m/Ma (Bierman and Caffee, 2002; Chappell, 2006; Quigley *et al.*, 2007a, b), allowing that relief is being produced within the ranges. Several authors suggest that up to half of the ~800 m relief has been built in the current stress regime (Sandiford, 2003b; Quigley *et al*., 2007c; Braun *et al*., 2009). Seismic reflection data indicate that the structural architecture mapped at the surface, corresponding to the Paleozoic Adelaide Fold Belt which developed over the inversion axis of a Neoproterozoic rift basin (Jenkins and Sandiford, 1992; Paul *et al*., 1999), extends to depths of 10–12 km beneath the ranges (Flöttmann and Cockshell, 1996). However, much of the seismicity recorded in the region occurs below this depth (Leonard, 2008), potentially in cratonic crust relating to the eastern margin of the Gawler Craton (cf. Figure 2.1). Therefore, it is unclear how the surface-faulting record might be related, if at all, to most of the instrumental seismicity (cf. Braun *et al*., 2009).

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In general, greater topographic expression associated with faults and fault systems occurring in extended crust relative to non-extended crust suggests a higher rate of seismic activity in the extended setting, consistent with observations worldwide (e.g., Johnston, 1994; Cloetingh *et al*., 2008; Mooney *et al*., 2012). Using the same reasoning, non-cratonic crust might be expected to have a higher rate of seismic activity than cratonic crust (cf. Figures 2.4 and 2.5), by virtue of there being no relief generation in cratonic crust. This distinction, together with the variation in fault character between domains, should be recognised in attempts to identify analogous systems worldwide.

2.4 Patterns in earthquake occurrence

The record of contemporary seismicity in Australia suggests that earthquake epicentres are spatially and temporally clustered (Denham, 1988; Leonard, 2008; Sinadinovski and McCue, 2010). As discussed in Section 2.1, concentrations of epicentres in the historic catalogue are dominated by four "seismogenic zones" (Figure 2.1) (Hillis *et al*., 2008; Leonard, 2008; Sandiford and Egholm, 2008). However, whether the short record of contemporary seismicity is representative of time periods significantly longer than the observation window has not been statistically or empirically tested in the Australian context (cf. Kafka, 2002). The persistence of patterns in the short historic catalogue can be assessed at much longer timescales by comparison with the record of morphogenic seismicity. Evidence from the paleo-record (Crone *et al*., 2003; Clark *et al*., 2011a, 2012), which essentially captures events of $\sim M > 5.5$, suggests that large earthquake occurrence within Australia exhibits both spatial and temporal clustering (Section 2.3). Temporal patterns in large SCR earthquake occurrence may be inferred at the scale of a single fault (Crone *et al*., 1997, 2003; Clark *et al*., 2008), of groups of faults (Leonard and Clark, 2011), and at the domain scale (Holdgate *et al*., 2003; Sandiford, 2003b; Paine *et al*., 2004; Braun *et al*., 2009; Clark *et al*., 2011a, 2012).

For example, the distribution of fault scarps in cratonic southwest Western Australia (D1) (Figure 2.4), together with the seismogenic strain arguments referred to previously (cf. Leonard *et al*., 2007; Leonard, 2008; Braun *et al*., 2009), infer that seismicity in the SWSZ represents only the current locus of activity, rather than a zone of long-lived activity (cf. Sandiford and Egholm, 2008). As most fault scarps in other parts of Australia are not associated with historic seismicity (e.g., Crone *et al*., 2003; Clark, 2010), a similar rule may apply.

There is no precedent in cratonic Australia (neotectonic domains 1 and 3 of Clark *et al.,* 2012) to indicate how long seismicity will persist. However, a large proportion of contemporary seismicity in the SWSZ is thought to relate to the Calingiri, Cadoux, and Meckering earthquakes (Leonard, 2008). If aftershock activity relating to surface-rupturing earthquakes is used as a measure of the longevity of activity in a region, then the work of Stein and Liu (2009) suggests that a millennial timescale might be applicable. Liu *et al*. (2011) and Liu and Wang (2012) present evidence for migration of the locus of seismicity

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at a centennial timescale in active intraplate northern China (Kusky *et al.*, 2007; Wheeler, 2011) with no recurrence over the 2,000–3,000-year window captured by the Chinese record of historical seismicity. As crustal deformation rates in this 1,000 km $\times$ 1,000 km region of China are in the order of 1–2 mm/yr (Liu *et al.*, 2011), such data might be considered as a minimum seismicity migration rate for Australia at the neotectonic domain scale.

There is some indication that the temporal clustering behaviour emerging from single-fault studies in non-cratonic Australia may be symptomatic of a larger picture of the more-or-less continuous tectonic activity from the late Miocene to Recent being punctuated by "pulses" of activity in specific deforming regions (e.g., Quigley *et al.*, 2010; Clark *et al.*, 2012). For example, major deformation episodes are constrained to the interval 6–4 Ma in southwest Victoria (D4) (Paine *et al.*, 2004) and 2–1 Ma in the Otway Ranges (D5) (Sandiford, 2003a). An episode of deformation ceased at 1.0 Ma in the offshore Gippsland Basin (D5) although it continued onshore until ~250 ka (Holdgate *et al.*, 2003). Holdgate *et al.* (2008) present evidence from the southeast highlands that resurrects the idea of a punctuated post-Eocene Kosciuszko Uplift event (Browne, 1967) that continued into the late Pliocene and possibly the Pleistocene.

The Mount Lofty and Flinders Ranges (D2) are perhaps an exception to this rule. The FRSZ has been associated with diffuse earthquake activity throughout the historic era (Greenhalgh and Singh, 1988; Greenhalgh *et al.*, 1994). While epicentres cannot in most cases be confidently associated with neotectonic faults, strain rates and uplift rates estimated from the seismic catalogue are approximately consistent with the number of neotectonic faults and their paleoseismologically derived slip rates (e.g., Sandiford, 2003b; Braun *et al.*, 2009). Hence, it is possible that seismicity is a long-lived (millions of years) process in this crustal setting.

While the suite of neotectonic fault behaviours may vary across Australia, as implied by the neotectonic domains model, one individual fault characteristic appears to be common to most Australian intraplate faults studied – active periods comprising a finite number of events are separated by typically much longer periods of quiescence (Crone *et al.*, 1997, 2003; Clark *et al.*, 2007, 2008, 2011a, 2012; Estrada, 2009) (Figure 2.6). Data and modelling from elsewhere in the world identify similar episodic behaviour on faults with low slip rates (e.g., Wallace, 1987; Ritz *et al.*, 1995; Marco *et al.*, 1996; Friedrich *et al.*, 2003) and suggest that the time between successive clusters of events (deformation phases) is highly variable but significantly longer than the inter-event times between successive earthquakes within an active phase (Marco *et al.*, 1996; Stein *et al.*, 1997; Chéry *et al.*, 2001; Chéry and Vernant, 2006; Li *et al.*, 2009). This characteristic has been referred to as Wallace-type behaviour (see Wallace, 1987), and may be conceptualised using a model similar to that proposed by Friedrich *et al.* (2003) (Figure 2.7).

The sparse data available in Australia suggests that an active period (e.g., t_1, t_2, t_3 in Figure 2.7) in cratonic central or western Australia (D1) might comprise as few as two or three events (e.g., Hyden – Crone *et al.*, 2003; Clark *et al.*, 2008), with interseismic intervals between large events in an active period in the order of 20–40 ka (Crone *et al.*,

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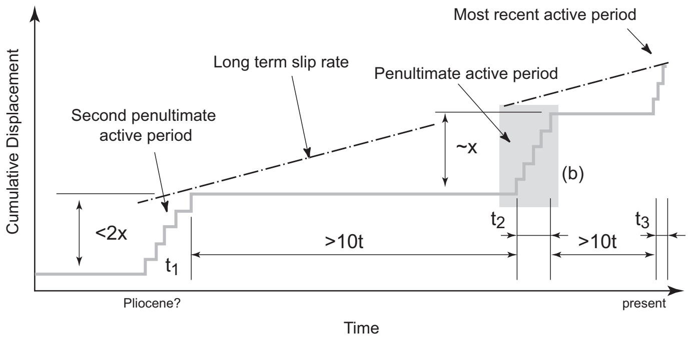

Figure 2.7 Generalised fault-slip diagram for Australian SCR faults based upon data from the Cadell Fault (Clark *et al.*, 2007; McPherson *et al.*, 2012a). Three active periods of fault growth (earthquake occurrence) are denoted by t_1, t_2, and t_3. These active periods are relatively short-lived, and each is composed of only a few ruptures (*c.* <6 events per active period). Inter-event times between successive ruptures within an active period may range up to several thousands to several tens of thousands of years. We adopt a characteristic earthquake rupture model based upon paleoseismic data from the Cadell (Clark *et al.*, 2007; McPherson *et al.*, 2012a) and Lake Edgar (Clark *et al.*, 2011b) faults. Long quiescent periods separate the active periods, and the length of the quiescent periods can range from many tens of thousands of years to greater than a million years.

1997, 2003; Clark *et al.*, 2008; Estrada, 2009). In non-cratonic eastern Australia (D4), recently acquired data on the Cadell Fault (Clark *et al.*, 2007; McPherson *et al.*, 2012a) indicate more frequent rupture, with up to six morphogenic events in the interval *c.* 70–25 ka. It is inferred that three uplift events of similar magnitude had occurred on the Cadell Fault prior to the diversion of the Murray and Goulburn Rivers at *c.* 45 ka (Bowler and Harford, 1966; Bowler, 1978), suggesting that these events are likely to have been spaced thousands of years apart, similar to the Meers and Cheraw faults in the western Central United States (Crone and Luza, 1990; Crone and Machette, 1995). This contrasts with the New Madrid seismic zone in the intraplate Central United States, where sequences of large earthquakes have occurred on average every 500 years for at least the last two seismic cycles in the current active period (Tuttle *et al.*, 2002). While Talwani and Shaeffer (2001) propose a similar recurrence for earthquakes large enough to produce liquefaction in the South Carolina Coastal Plain, it has not been determined whether their data reflect rupture on a single fault or multiple faults. Quiescent intervals can be sufficiently prolonged (hundreds of thousands to millions of years), in the western and central parts of Australia in particular, that most or all relief relating to an active period might be removed by erosion prior to the next active period (Crone *et al.*, 2003; Clark *et al.*, 2007, 2008, 2011a).

2.5 Maximum magnitude earthquake

Large earthquakes are so infrequent in SCRs such as Australia that the data distributions upon which recurrence and $M_{\max}$ estimates are based are heavily skewed towards

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magnitudes below M_W 5.0 (e.g., Leonard, 2008), and thus require significant extrapolation up to magnitudes at which the most damaging ground-shaking might be expected (Wheeler, 2009a, b). Australia is uniquely suited to assessing the validity of M_{max} estimates derived from the instrumental record of seismicity by virtue of its extraordinary neotectonic and paleo-earthquake record spanning many tens of thousands of years (e.g., Clark *et al.*, 2011a).

Where a fault has been studied paleoseismologically, single-event scarp lengths and single-event displacements can provide two independent estimates of the characteristic magnitude (cf. Schwartz and Coppersmith, 1984) for that fault, with scarp length being considered the more reliable measure (e.g., Wells and Coppersmith, 1994; Hemphill-Haley and Weldon, 1999; Leonard, 2010). Where a fault has not been studied in detail the "characteristic" rupture of the entire scarp length during each morphogenic event, subject to appropriate caveats regarding segmentation and dip, might be assumed in order to estimate future large earthquake potential (e.g., Hemphill-Haley and Weldon, 1999; Stirling *et al.*, 2002; Wheeler, 2009a).

In most regions of Australia the neotectonic record is either incomplete, or under-explored (Sandiford, 2003b; Clark *et al.*, 2011a). As such, it is not possible to assert with confidence that future large earthquakes will be restricted to known scarps/faults with documented source characteristics. It is therefore desirable to aggregate characteristic event magnitudes from a group of faults with similar source characteristics from within a region of interest (e.g., neotectonic domain or aggregation of domains [Leonard *et al.*, 2012; Clark *et al.*, 2012]) to provide an estimate of M_{max}. This can be achieved with varying levels of confidence, depending upon the estimated completeness of the paleo-record.

2.5.1 Scarp length as a proxy for paleo-earthquake magnitude

A relatively large area of 10 m resolution DEM data in the southwest corner of the Yilgarn Craton (D1) (Figure 2.4) allowed for mapping of fault scarps in unprecedented detail (Clark, 2010). Based upon an assessment of erosion and landscape modification rates, it was estimated that most scarps representing events of $M_W \geq 6.5$ that had occurred in the last ~100 ka were captured in the catalogue (Leonard and Clark, 2011). The scaling relations of Leonard (2010) were used to develop a paleo-seismicity catalogue comprising 65 events (Leonard and Clark, 2006, 2011), which was subsequently combined with the catalogue derived from instrumental seismicity. The data were found to exhibit typical truncated Gutenberg–Richter recurrence characteristics with a slope (b) of 0.9–1.0 between M_W 6.5 and 6.9. A rapid roll off in recurrence occurred above M_W 6.9, with an asymptotic value of M_W 7.25 ± 0.1 considered to be the M_{max} for non-extended cratonic crust typified by the Yilgarn Craton (Leonard and Clark, 2006, 2011). A less well constrained M_{max} of M_W 7.65 ± 0.1 was determined for scarps representing extended crust in the rift basins flanking the western margin of the continent (D6).

In most regions of Australia the neotectonic catalogue is far from complete, the preservation time of seismogenic features in the landscape is uncertain, and single-event scarp

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Table 2.2 *Ninetieth (90th) percentile fault scarp length values for each neotectonic domain of Clark et al. (2012) and corresponding characteristic earthquake magnitude (M_W) values determined using the scaling relation of Leonard (2010) assuming a 45° dipping fault and a seismogenic depth of 15 km*

Domain	Length [90%] (km)	Magnitude (M_W)
1	51	7.3*
2	78	7.4
3	104	7.6
4	124	7.6
5	57	7.3
6	124	7.6
All data	101	7.6

*The cratonic non-extended relation of Clark *et al.* (2013) obtains a value for D1 of M_w 6.9.

length and slip data are largely absent. Without these data, key assumptions of the curve-fitting approach to M_{max} estimation described above are not satisfied. However, scarp length variation across the continent (Figure 2.5) does imply variation in characteristic earthquake magnitude (and M_{max}).

Across Australia there is a strong positive skew in the length data distribution, with 90% of scarps less than 101 km in length (Figure 2.5a; Table 2.2). Assuming a generic fault dip of 45° and a seismogenic depth of 15 km (Collins *et al.*, 2003), the 90th percentile value for length corresponds to a value of M_W 7.6 (cf. Leonard, 2010) (Table 2.2). Without paleoseismic data on the longest scarps it is difficult to assess the validity of this estimate as a value for M_{max}. The longest scarp with paleoseismological evidence consistent with entire length rupture is the 80 km long ($\sim M_W$ 7.4) Cadell Fault scarp in eastern Australia (D4) (Clark *et al.*, 2007, 2011a; McPherson *et al.*, 2012a); however, rupture segmentation is plausible for intraplate faults of several tens of kilometres in length or greater (e.g., Machette *et al.*, 1991). The modal scarp length value of 53 km for all neotectonic features (Figure 2.5a) corresponds to an earthquake of magnitude $\sim M_W$ 7.3.

Fault scarp length data for all individual domains (Figure 2.5e) show a similar positive skew to the aggregated population data distribution (Figure 2.5a), which highlights the same uncertainties with respect to possible fault segmentation. The 90% values for fault length provide magnitude estimates in the range of M_W 7.3–7.6 (Table 2.2). The values for D1 and D3 are within error of the M_{max} estimated by Leonard and Clark (2011). Within D2 the only known single-event scarp is ~28 km long (M_W 6.8–6.9) and relates to the greater than 100 km long Williamstown-Meadows Fault (Clark and McPherson, 2011).

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However, paleoseismological evidence exists elsewhere in D2 for very large single-event displacements (>7 m – Quigley *et al.*, 2006; Reid, 2007; Clark *et al.*, 2011a, 2012), consistent with modelled event magnitudes of M_W 7.3–7.5 (Somerville *et al.*, 2008). Very long 90th percentile values for D3 and D4 require validation in terms of equivalence to M_{max} as these scarps also tend to have accumulated significant neotectonic throw (tens of metres), implying multiple morphogenic events. The dominance of folding as opposed to discrete faulting in D5 has thus far prevented estimation of single-event rupture lengths and displacements (Clark *et al.*, 2011a).

As expected, overall indications are that the historic catalogue of seismicity significantly underestimates the large earthquake potential (and, by proxy, M_{max}) in most regions of Australia. To a first order at least, the sub-division of the continent into domains on the basis of geology and tectonic history (Johnston, 1994; Clark *et al.*, 2011a, 2012) provides useful insights into variations in faulting character that can facilitate the interpretation of the neotectonic and historic catalogues. The use of fault-length data from the neotectonic catalogue provides reasonable preliminary estimates of M_{max} when applied in conjunction with appropriate scaling relations (e.g., Leonard, 2010; Clark *et al.*, 2013). Analysis of the current neotectonic catalogue for Australia suggests that a range of M_{max} values of M_W $\sim$ 7.0–7.6 could reasonably encompass all geological and tectonic settings continent-wide (Clark *et al.*, 2011a, 2012; Leonard, 2012; Burbidge, 2012).

2.6 Implications for SCR analogue studies: factors important in earthquake localisation

Analogues between the Australian neotectonic domains (Clark *et al.*, 2011a, 2012) and SCR crust elsewhere in the world (cf. Johnston, 1994) are readily apparent. For example, poly-phase deformation of a compressional nature is a common feature in the post-rift evolution of many extended passive margins and rifts (D5/D6 analogues) (van Arsdale, 2000; Balasubrahmanyan, 2006; Cloetingh *et al.*, 2008). Archean cratonic nuclei fringed by Paleoproterozoic mobile belts (D1 analogues) make up a large portion of the geology of Peninsular India (Kroner and Cordani, 2003) and North America (Hoffman, 1989). Meso- and Neoproterozoic mobile belts involved in the accretion of the supercontinent Rodinia (D3 analogues) are found worldwide (e.g., Collins and Pisarevsky, 2005; Cawood and Buchan, 2007), as are Phanerozoic accretionary terranes associated with the amalgamation of the supercontinent Gondwana (D4 analogues) (e.g., Hoffman, 1989). The temporal clustering of morphogenic earthquake events seen in studies on Australian faults (Quigley *et al.*, 2010; Clark *et al.*, 2011a, 2012) also appears to be mirrored in the Central and Eastern United States (Crone and Luza, 1990; Crone *et al.*, 1997; Cox *et al.*, 2006). Inferences made regarding mechanisms responsible for localising intraplate seismicity in Australia might then be assessed in terms of their crustal and lithospheric setting, and tested on analogous crust elsewhere in the world.

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2.6.1 Mechanical and thermal influences

It has been proposed that intraplate regions with higher seismic potential have pre-existing zones of weakness (Sykes, 1978; Talwani and Rajendran, 1991; Stuart *et al*., 1997; Kenner and Segall, 2000; Dentith and Featherstone, 2003), intersecting faults (Talwani, 1988, 1999), elevated heat flow (Liu and Zoback, 1997; Celerier *et al*., 2005; Hillis *et al*., 2008; Holford *et al*., 2011), crustal anomalies (Campbell, 1978; Kenner and Segall, 2000; Gangopadhyay and Talwani, 2003; Pandey *et al*., 2008; Assumpção and Sacek, 2013) or can be identified by crustal boundaries inferred from potential field data (Langenheim and Hildenbrand, 1997; Lamontagne *et al*., 2003; van Lanen and Mooney, 2007; Dentith *et al*., 2009). It is likely that the variety of models reflects the range of mechanisms that are operating, and although all of the above mechanisms have demonstrated local applicability in the Australian context, counter-examples are abundant. This is particularly the case where models relying on thermal mechanisms for strain localisation have been proposed.

Sandiford and Egholm (2008) argue that enhanced seismicity along some parts of the Australian continental margin is a consequence of thermal weakening due to steady-state heat flow across the lithospheric thickness steps between oceanic and continental crust (e.g., SESZ, Figure 2.1) and extended continental and non-extended cratonic crust (e.g., SWSZ, Figure 2.1) (Fishwick *et al*., 2008; Kennett *et al*., 2013). While the hypothesis is intuitively appealing, in Australia and along the eastern seaboard of North America (Wheeler and Frankel, 2000), several lines of evidence suggest that the contribution of this mechanism to the continental seismicity budget over geological timescales is minor.

For example, east of the large lithospheric thickness step on the western boundary of the SWSZ (i.e., across the Darling Fault), fault scarps are randomly distributed (Clark, 2010) and the topography predicted if instrumental seismic moment release rates are extrapolated to million-year timescales (Braun *et al*., 2009) is absent. Given extremely low rates of bedrock erosion (Bierman and Caffee, 2002; Belton *et al*., 2004; Chappell, 2006), this finding implies that the locus of seismicity in the SWSZ is transitory, rather than responding to steady-state heat flow at the margin. Furthermore, very little seismicity, paleo- or instrumental, can be correlated with the dramatic transition from cratonic to non-cratonic lithospheric thickness in eastern Australia (i.e., the Tasman Line, Figures 2.1 and 2.3), and significant instrumental seismicity proximal to the eastern seaboard (the SESZ) is not associated with a large lithospheric thickness step (Fishwick *et al*., 2008).

More generally, Holford *et al*. (2011) propose that the thermal properties of the crust and upper mantle exert a regional-scale (100–1000 km) modulating control on which parts of the Australian lithosphere undergo (seismogenic) failure and which parts experience relatively less deformation (cf. Celerier *et al*., 2005; Sandiford and Quigley, 2009). Specifically, these authors invoke relatively high heat flow to explain localisation of seismic moment release and deformation in the Flinders Ranges Seismic Zone (FRSZ, Figure 2.1) compared to the flanking Murray Basin and Nullarbor Plain. The correlation is imperfect; the heat flow anomaly does not extend to the Gippsland and Otway Basins (Densley *et al*., 2000;

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Holdgate *et al.*, 2003; Sandiford, 2003a) (Figure 2.1), which are manifestly amongst the fastest deforming regions on the Australian continent (Sandiford, 2003a, b; Clark *et al.*, 2011a, 2012). Actively inverting basins of the Northwest Shelf (NWSSZ, Figure 2.1) are also not associated with significant heat flow anomalies (cf. He and Middleton, 2002). Furthermore, the heat flow anomaly is most pronounced in the Cooper/Eromanga Basin, a region sparse in seismicity (Figure 2.1) and devoid of any known neotectonic features or tectonic uplift.

At the sub-regional scale, a range of factors have been proposed to explain localisation of historic seismicity. However, it seems clear that, on timescales of thousands to tens of thousands of years, seismic potential is determined by factors at a scale much larger than a single fault or region, and might instead relate more strongly to continental-scale lithospheric and crustal architecture, and the age of that architecture, as first proposed by Johnston *et al.* (1994).

2.6.2 Structural architectural influences

In the context of structural architecture, intraplate seismicity worldwide is considered to be concentrated at rifted margins (Stein *et al.*, 1989; Wheeler, 1995, 1996; Sandiford and Egholm, 2008; Cloetingh *et al.*, 2008; Etheridge *et al.*, 1991; Talwani and Schaeffer, 2001), interior rifts [aulacogens] (Johnston, 1994; Wheeler, 1995; Gangopadhyay and Talwani, 2003; Schulte and Mooney, 2005; Sinha and Mohanty, 2012) and at the margins of cratons (e.g., Lenardic *et al.*, 2000; Mazzotti, 2007; Sloan *et al.*, 2011; Craig *et al.*, 2011; Mooney *et al.*, 2012). A consequence of the relatively cold geotherm characterising cratonic areas is that there is no decoupling between crustal and mantle deformation (Braun *et al.*, 2009). This implies that, over geological timescales, deformation of the upper crust must be spatially uniform (e.g., Clark, 2010), resulting in little strain localisation, and hence minimal topography. Predictably, the historic record of seismicity in Australia is a poor guide as to where localisation of seismic activity occurs over geological timescales, with significant concentrations of seismicity occurring within cratons (e.g., SWSZ, Figure 2.1), and far from extended crust or craton boundaries (e.g., parts of the SESZ; see also Adams *et al.* [1992] and Rajendran *et al.* [1996]).

The major tenet of the Johnston *et al.* (1994) SCR model – that extended crust is more seismically active than non-extended crust, and that within the non-extended class non-cratonic crust is more active than cratonic crust – appears to hold true for the Australian neotectonic record if the uplift rate implied by vertical neotectonic fault displacement is taken as a proxy for seismic activity (Figure 2.4d, f). Australian neotectonic data (Clark *et al.*, 2011a, 2012) permit further sub-division of the extended and non-extended crustal classes proposed by Johnston *et al.* (1994). In extended crust domains (e.g., D5, D6), the age of major rifting appears to be important in terms of the record of neotectonic activity. Paleozoic intracratonic rifts (e.g., the Fitzroy Trough [Drummond *et al.*, 1991]; cf. Figure 2.1) and passive margin components (e.g., Perth Basin [Crostella and Backhouse,

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2000; Norvick, 2004]) preserve less evidence for neotectonic strain localisation than those rifted in the Mesozoic (e.g., Gippsland and Otway basins). As such, the long-term seismic potential of aulacogens impinging along the eastern margin of the North American continent (e.g., Ottawa Rift, Saguenay graben, Reelfoot Rift, Southern Oklahoma aulacogen; Wheeler and Frankel [2000]) might be assessed based upon the proportion of Mesozoic extensional reactivation of these largely Paleozoic structures. The distinction becomes even clearer when one considers most Precambrian rifts (included in D1 and D3). The major exception to this rule is the FRSZ (D2), which formed as a Precambrian rift, but was subsequently extensively reactivated under compression in the Paleozoic (Flöttmann and James, 1997). The Appalachian Orogen (Wheeler, 1996), including the St. Lawrence seismic zone (Vlahovic *et al*., 2003) at its northern margin, occupies a similar crustal setting between Precambrian mobile belts (D3 equivalent) and Phanerozoic accretionary terranes (D4 equivalent).

Considerable uncertainty remains as to the long-term seismic potential of non-extended, non-cratonic crust (D4), which typically comprises Phanerozoic accretionary terranes (e.g., Hoffman, 1989; Glen, 2005). Within the Australian continent, much of this crustal type is associated with little positive relief, and so might be thought of as having low seismic potential (e.g., Sandiford, 2003b). However, the eastern part of D4 is punctuated by the Eastern Highlands. While it is widely considered that most of the relief of the highlands relates to the opening of the Tasman Sea in the Cretaceous (e.g., Norvick and Smith, 2001), debate remains as to how much relief has been added from the Neogene to Recent (Holdgate *et al*., 2008, 2011; VandenBerg, 2010). The few known faults are poorly documented, but appear to have similar characteristics to those documented from the relief-poor western parts of the domain (compare the Lake George Fault [Abell, 1981, 1985] to the Cadell Fault [Clark *et al*., 2007; McPherson *et al*., 2012a]). Braun *et al*. (2009) hypothesise that deformation in eastern Australia is determined by the lithospheric strength (or rigidity) and hence controlled by lithospheric structures, in contrast to the western part of the continent where deformation is localised by crustal structures, because the underlying lithospheric mantle is almost ubiquitously strong.

A fundamental implication of the neotectonic domains model presented by Clark *et al*. (2011a, 2012), building upon the pioneering work of Johnston *et al*. (1994), is that intraplate fault characteristics (and by inference, seismicity) are not universal in their applicability in analogue studies. Careful choice of subject regions, structures, or faults within analogous crust of similar stress field character is required to extrapolate meaningfully to imperfectly characterised areas.

2.7 Conclusions

Australia has a short recorded history of seismicity, spanning only a couple of centuries (Leonard, 2008). As a consequence, there is significant uncertainty as to whether patterns evident in the contemporary seismic record are representative of the longer term. This uncertainty can, in part, be overcome by validation against Australia's rich record of

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morphogenic earthquakes (Quigley *et al*., 2010; Clark *et al*., 2012). Long-term patterns in large earthquake occurrence, both temporal and spatial, can be deduced from the landscape record and used to inform contemporary earthquake hazard science. Seismicity source parameters such as large earthquake recurrence and magnitude vary across the Australian continent, and can be interpreted in a framework of large-scale neotectonic domains defined on the basis of geology and crustal setting (Clark *et al*., 2012). Temporal and spatial clustering of earthquakes is apparent at the scale of a single fault, and at the 1,000 km scale of a domain. The utility of the domains approach, which ties seismicity characteristics to crustal architecture and geology, is that behaviours might be extrapolated from well-characterised regions to poorly known analogous regions, both within Australia and worldwide.

Acknowledgements

This paper is published with the permission of the CEO of Geoscience Australia. The authors would like to thank Marcelo Assumpção and Pradeep Talwani for providing review comments that improved the manuscript. Mark Leonard and David Burbidge are also thanked for their constructive reviews.

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