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Microseismic Activity and Basement Controls on an Active Intraplate Strike-Slip Fault, Ceres–Tulbagh, South Africa

by Louis Smit, Åke Fagereng,^{*} Benjamin Braeuer, and Jacek Stankiewicz

Abstract Earthquakes far from plate boundaries are poorly understood, because of few well-studied examples and uncertainty of what controls the location of these events. In 1969 a damaging local magnitude (M_L) 6.3 strike-slip earthquake, with no surface expression, occurred in the Ceres–Tulbagh region in the South African stable continental interior. Here, we present a microseismic study of the Ceres–Tulbagh area, conducted over three months in 2012, in which 172 events recorded on at least three stations follow a Gutenberg–Richter relationship for $-1.5 < M_L < 0.5$. The events delineate a 5 km wide, subvertical zone that is microseismically active to a depth of 15 km. This fault zone is subparallel to the 1969 aftershock zone and at low angle to the regionally inferred greatest horizontal stress. We argue that the microseismically active zone is guided by inherited structures in the basement geology. This and similar structures may represent significant earthquake hazard in plate interiors.

Online Material: Tables of station and event parameters, and *P*- and *S*-wave arrival times.

Introduction

Despite being tectonically quiet, and deforming at slow strain rates, plate interiors are affected by intraplate stresses produced by forces acting on plate boundaries (Zoback and Zoback, 1981; Zoback, 1992). Although patterns of strain accumulation and elastic strain release along plate margins, in response to steady relative plate motion, lead to seismic activity with a somewhat predictable recurrence time and fault mechanism (e.g., Stein and Freymuller, 2002), intraplate earthquakes have a more complex spatial and temporal distribution (e.g., Crone *et al.*, 2003; Li *et al.*, 2009; England and Jackson, 2011).

Globally, the best-studied intraplate seismic zone is the source of the New Madrid earthquakes of 1811–1812. This earthquake sequence, of which at least three events have estimated moment magnitude (M_w) above 7.0 (Hough, 2009), occurred within a Late Proterozoic to Early Cambrian failed intracontinental rift (Thomas, 1991). Within this rift, recently active faults appear to form an irregular geometry involving dextral strike-slip faulting and an unstable, young, compressional stepover (Schweig and Ellis, 1994). The 1811–1812 earthquake sequence was not a typical mainshock–aftershock sequence, but has been interpreted as failure on multiple fault structures across a range of scales, perhaps related to failure of a series of existing faults, and a range of length scales in this

fault system (Johnston and Schweig, 1996). It is also possible that some faulting is related to deformation along rheological boundaries separating rift-related intrusions and surrounding country rock (Hildebrand *et al.*, 1992). A paradox, however, arises from the observation that current geodetically determined strain rates in the New Madrid region are indistinguishable from zero (Newman *et al.*, 1999). Thus, the fault system is not undergoing steady loading, but rather the strain build-up leading to the 1811–1812 earthquakes was transient (Calais and Stein, 2009). From paleoseismic records, the fault system appears to have seen a number of temporal clusters of large earthquakes, about 500 years apart, over the last 2000 years (Tuttle *et al.*, 2002), but there are no definite reports of evidence for earlier activity. It therefore seems that intraplate faults, at least in the New Madrid seismic zone, are transient features that can turn on and off on 1000-year time scales, maybe by movement of deformation between faults in a distributed deforming zone (Stein *et al.*, 2009).

One way to explain intraplate seismicity is to invoke weak faults within otherwise strong upper crust. Mechanisms of fault zone weakening have been explored in great detail (e.g., Holdsworth *et al.*, 2001) and include the development of throughgoing surfaces of low friction (Collettini *et al.*, 2009), effects of frictional–viscous flow facilitated by pressure–solution creep (Niemeijer and Spiers, 2006), local fluid overpressure (Hubbert and Rubey, 1959; Rice, 1992), reshear of existing frictionally weak fabric that also serves to contain

^{*}Now at the School of Earth & Ocean Sciences, Cardiff University, Park Place, Cardiff CF10 3AT, United Kingdom; FagerengA@cardiff.ac.uk.

high fluid pressures (Fagereng *et al.*, 2010), or ongoing fluid-assisted chemical reactions (Wintsch *et al.*, 1995). Existing weak faults or shear zones may therefore exist in the upper crust, allowing for localized, preferential reactivation. Study of earthquake fault mechanisms and regional stress orientations in the continental interior of the United States and Canada indicates that in these areas, including New Madrid, earthquake mechanisms are compatible with the reactivation of optimally oriented faults in the current local stress field (Hurd and Zoback, 2012).

Intraplate seismicity may also be triggered by having a weak zone in the lower crust and/or upper mantle, leading to local increase in lower crustal ductile strain rate and consequent upper crustal stress concentration, and localized seismicity overlying the lower crustal shear zone (Zoback *et al.*, 1985). Liu and Zoback (1997) suggested this mechanism for the New Madrid seismic zone, on the argument that increased heat flow in the seismic zone reflects higher thermal gradient and warm, weak, lower crust and upper mantle in this area. However, later studies failed to find significant thermal differences between the New Madrid seismic zone and its surrounding crustal volumes (McKenna *et al.*, 2007). Alternatively, weak ductile shear zones in the mid to lower crust may be present from previous tectonic episodes and not require high heat flow to explain their weakness. Such weakening in the ductile mid to lower crust may be related to, for example, compositional anisotropy (Goodwin and Tikoff, 2002), fine grain size (Rennie *et al.*, 2013), a well-developed foliation (Montesi, 2013), or existing brittle faults on which new ductile shear zones preferentially nucleate (Mancktelow and Pennacchioni, 2005). If such a weak zone is reactivated by localized ductile flow in the lower crust, it is conceivable that stress concentration in the overlying brittle upper crust leads to seismic activity. It is also theoretically possible to create stress concentrations at boundaries between rocks of significantly different strength, such that intraplate seismicity could be related to (1) stress concentrations around weak, serpentinized mafic intrusions (Campbell, 1978); (2) elevated stresses in relatively rigid wallrock surrounding weak felsic plutons (Stevenson *et al.*, 2006); or (3) stress risers within zones of otherwise slow, continuous deformation, such as more rigid lenses within weak shear zones (Fagereng, 2013).

Although it appears likely that the New Madrid earthquake sequence involved reactivation of existing rift structures (either faults, lithological boundaries, ductile shear zones, or a combination of different structural elements), there are many faults, shear zones, and lithological boundaries in the area that have not been reactivated, despite having similar orientation and inferred strength to the structures that have failed (e.g., Zoback and Zoback, 1981). Therefore, at least for New Madrid, a source of local stress concentration or fault weakening has been sought. Grollmund and Zoback (2001) suggested that deglaciation following the last glacial maximum might have led to sufficient stress change to trigger faulting. This mechanism is unlikely, however, given that the distance from New Madrid to the inferred ice margin is

too large unless a fault was already on the brink of frictional failure (Wu and Johnston, 2000). More recently, Calais *et al.* (2010) proposed localized erosion as a mechanism of reducing normal stress, and thereby also frictional resistance, on some faults that already experienced shear stress near the failure stress. Both postglacial stress change and local erosion are features of the last few thousand years and thus are able to explain the recent onset of seismicity in New Madrid. If this interpretation is true, the driving force is still present and seismic risk continues in the New Madrid seismic zone (Zoback, 2010).

The above review of potential origins for intraplate seismicity highlights a multitude of possible sources. We highlight (1) frictional reactivation of existing upper crustal faults, (2) stress concentration at lithological boundaries representing rheological contrast, and (3) ductile strain localization in existing mid to lower crustal shear zones, as three main likely origins of localized deformation within stable continental interiors. In addition, at least in the New Madrid seismic zone, a temporal trigger such as localized erosion or glacial rebound is also likely to play a role in clustering intraplate seismicity in time and space (Calais *et al.*, 2010). A weakness in the study of intraplate seismicity is the lack of study sites, and not much data have been collected outside the United States, as also highlighted by the review above. Here, we offer new insights from a seismic zone in the South African continental interior. An intraplate sinistral strike-slip earthquake with local magnitude (M_L) 6.3 occurred near the towns of Ceres and Tulbagh in the Western Cape, South Africa, on 29 September 1969 (Green and Bloch, 1971; Green and McGarr, 1972; Krüger and Scherbaum, 2014). To date this is the largest instrumentally recorded earthquake in South African history, and the towns of Ceres and Tulbagh have continued to experience regular seismicity of $M_L > 3$ after 1969 (Singh *et al.*, 2009; Saunders *et al.*, 2012). We first review the geology and tectonic history of this area, before we consider the 1969 earthquake and its aftershocks. We present new microseismic observations recorded on a local, temporary network to constrain current deformation, and finally integrate these observations to address the origin of intraplate seismicity in the Ceres–Tulbagh region.

Geology of the Ceres–Tulbagh Area

South Africa comprises a number of orogenic belts surrounding the Kaapvaal craton (Fig. 1a), among which the Namaqua–Natal belt likely records collisional tectonics during the Mesoproterozoic assembly of Rodinia (Dalziel *et al.*, 2000; Rodgers and Santosh, 2003) and the Gariep and Saldania belts formed during Pan-African, Late Neoproterozoic to Early Paleozoic, amalgamation of the supercontinent Gondwana (Frimmel and Frank, 1998; Rozendaal *et al.*, 1999). The Permo-Triassic Cape Fold belt has been related to more distant subduction, toward the north under the southern margin of Gondwana (Lock, 1980). The general interpretation is that the Cape Fold belt formed along reactivated Pan-African

Figure 1. Geological setting of the study area: (a) major tectonostratigraphic elements of South Africa and location of the study site (CFB, Cape Fold belt), and (b) structure and geology of the syntaxis of the Cape Fold belt (after de Beer, 1992). The rectangle in (b) shows the study area, which is shown in more detail in Figure 2. The color version of this figure is available only in the electronic edition.

structures (e.g., de Beer, 1992), which led to the development of two branches: a north–south-trending western branch and an east–west-trending southern branch (Fig. 1b). The epicenter of the 1969 Ceres–Tulbagh earthquake, and the current study area, lie in the syntaxis of the Cape Fold belt, where the southern and western branches meet (Figs. 1b, 2).

The area around Ceres is mountainous, and the exposed geology is Cape supergroup quartzites in high areas and recent alluvium overlying Malmesbury group shales in the valleys (Fig. 2). This surface geology reflects the structure of the underlying geology, which comprises rocks of the Namaqua metamorphic complex at depth, overlain by the Malmesbury group and Cape supergroup sedimentary sequences. The Namaqua metamorphic complex comprises granulite facies, dominantly quartzofeldspathic, rocks; however, because they do not crop out in this area, their exact composition and structure are unknown. In comparison with observations further north, where outcrops exist, the fabric of the Namaqua metamorphic complex is likely to be dominated by gently north-to-northeast-dipping foliations (Colliston *et al.*, 2014). The Malmesbury group comprises lower greenschist facies

siliciclastic sedimentary and volcanic rocks, deformed in the Pan-African Saldania orogeny, in which thrust and strike-slip faults have a predominant northwesterly strike (Belcher and Kisters, 2003). Throughout western South Africa, these steeply dipping, west-northwest to north-northwest-striking frictional–viscous strike-slip faults related to Pan-African orogens have experienced multiple events of reactivation and affect Namaqua metamorphic complex as well as Pan-African rocks (Fig. 1b) (e.g., Viola *et al.*, 2012). The Early Ordovician to Early Carboniferous Cape supergroup overlies the Malmesbury and comprises up to 10 km in thickness of siliciclastic sedimentary rocks, with quartzites by far the dominant lithology (Tham and Johnson, 2009). Rocks of the Cape supergroup were deformed in the Permian to Triassic formation of the Cape Fold belt, which involved reactivation of Pan-African structures, as well as formation of new east–west-trending folds in the southern branch, and north–south-trending folds in the western branch (Fig. 1b) (e.g., de Beer, 1992). In places, reactivated Pan-African faults propagated through the Cape supergroup sediments, whereas in other regions, the thick Cape supergroup sediments buried the existing fault systems (Fig. 1b).

The 1969 Ceres–Tulbagh Earthquake

The 1969 Ceres–Tulbagh earthquake occurred where the surface geology comprises Malmesbury group and Cape supergroup rocks (Fig. 2). There is no fault surface trace that correlates with the epicenter of the subvertical rupture plane (Green and McGarr, 1972), and the main deformation structures visible in the area are open to close folds with upright, north–south-striking axial planes (Fig. 2). However, one cannot exclude west-northwest to north-northwest-striking, subvertical faults being present in the Malmesbury and Namaqua metamorphic complex rocks buried by the Cape supergroup and recent alluvium.

A focal mechanism based on 42 stations of the Global Seismic Network indicates that the 1969 rupture plane was subvertical with a strike of $\sim 319^\circ$ (Green and McGarr, 1972). Using full waveform modeling, Krüger and Scherbaum (2014) constrained the centroid depth to less than 15 km. Aftershocks from the 1969 event occurred along a subvertical plane that strikes 299° (Fig. 2) (Green and Bloch, 1971). Although the instrumental record of the South African National Seismic Network (SANSN) defines a cluster of activity near Ceres, the geometry of this cluster of recent-to-current activity is unknown (Singh *et al.*, 2009). Therefore, to investigate current seismic activity in Ceres, a temporary array of 15 surface stations was set up to record microseismic activity for three months from February to May 2012, covering the aftershock zone of the 1969 earthquake (Table S1, available in the electronic supplement to this article). We performed a coupled velocity–hypocenter inversion to obtain the spatial distribution of seismicity, 1D velocity structure, and earthquake magnitude distribution.

Figure 2. Location and geology of the study area, as well as aftershocks from 1969 and the epicenters of the 1969 M_L 6.3 earthquake and the 1970 M_L 5.7 aftershock (Green and Bloch, 1971). The focal mechanism for the 1969 M_L 6.3 event by Krüger and Scherbaum (2014) is also shown. (Geology after Gresse and Theron, 1992.) The east–west cross section shows the inferred structures within the Cape supergroup and Malmesbury group rocks, but the structure of the underlying Namaqua–Natal belt is not known in detail.

Methodology

Each station consisted of a PE-6/B triaxial 4.5 Hz geophone, a three-channel DSS-cube data recorder with built-in Global Positioning System, and a battery pack supplying power. The instruments were buried roughly 30–50 cm deep in sand or clay and recorded continuously at 200 samples/s. Batteries were changed and data were downloaded after six weeks. One station malfunctioned after two days and worked only sporadically after battery change. Another was vandalized by baboons after 19 days, and the battery pack was missing. A new battery pack was fitted, and the station continued recording. Excluding the two failed stations, the stations recorded for 89 days on average.

From the continuous waveform data, 351 seismic events were manually identified for further processing. The arrival time of P and S phases was manually picked by viewing each station for each of these 351 events individually. P phases were picked on the vertical axis and S phases on the horizontal axes.

Picks were weighted according to the clarity of the waveforms and the confidence of the pick. To improve the signal-to-noise ratio, a band-pass filter with a minimum low and maximum high cutoff filter of 1 and 40 Hz, respectively, was applied.

Coupled hypocenter–velocity inversion was performed in VELEST (Kissling *et al.*, 1994) to produce a minimum 1D velocity model (Fig. 3) and station corrections. A total of 59 events recorded on a minimum of six stations were used for coupled inversion. A further 113 events recorded on three or more stations with a gap of less than 210° were used for single-mode relocation. The distribution of events with $181^\circ < \text{gap} < 210^\circ$ was checked against events with $\text{gap} < 180^\circ$; no irregularities were found, so the events were included for hypocenter distribution plots (Fig. 4).

Event arrival times were normalized to S picks, and maximum amplitude values were extracted for the 172 microseismic events. Local magnitudes (M_L) were calculated with the equation of Havskov and Ottemöller (2010), using local correction factors developed for the SANSN (Saunders

Figure 3. The velocity model derived from 2011 microseismicity data. The preferred models (lowest residual) are shown as solid lines, and the range of uncertainty is shown as dashed lines.

et al., 2012), calibrated to an M_L 2.4 event that was recorded by the temporary Ceres array as well as SANSN. The event was located offshore the west coast of South Africa roughly 110 km west of Ceres. (E) A full list of events is available in Table S2, and the associated P and S times are tabulated in Table S3.

Results

In the preferred 1D velocity model based on the recorded microseismicity (Fig. 3), V_P ranges from 4.9 km/s at the surface to 6.4 km/s at a depth of 12 km. V_S is 2.8 km/s at the surface and increases to 4.7 km/s at 12 km depth. The locations of the microseismic events define a subvertical fault zone roughly 5 km wide and striking roughly 301° . There is a small angular difference of $\sim 6^\circ$ between the average surface traces of the microseismic events recorded in 2012 and the 1969 aftershocks, within the uncertainty of the surface trace averages. Seismicity is clustered in space and defines two distinct clusters with different depth distribution separated by a lateral discontinuity in seismic activity (Fig. 4). Microseismic activity in the Ceres cluster ranges from the surface down to roughly 5 km depth with little seismicity in the 5–8 km depth range. Microseismicity in the Tulbagh cluster occurs from roughly 8 to 15 km, with little activity shallower than 8 km depth. This depth distribution is comparable to that of 1969 aftershocks (Green and Bloch, 1971). The 1969 mainshock appears to have occurred in the along-strike transition between the two clusters (Fig. 4).

The events recorded by the temporary array fall in the range $-2.2 < M_L < 1.6$, with a magnitude of completeness (M_c) of -1.5 (Fig. 5). The low magnitude of completion may be ascribed to the fact that bedrock is present at or near the surface, and noise levels are relatively low. The magnitude distribution follows a Gutenberg–Richter relation with a b -value of 0.89 ± 0.02 (95% confidence) for magnitudes in the range $-1.5 < M_L < 0.5$ (Fig. 5).

Discussion and Conclusions

The magnitude of completeness of $M_L -1.5$ is exceptionally low, whereas the b -value is within the normal range. As such, the results here mirror the frequency–magnitude dis-

tribution seen in other studies (e.g., Abercrombie, 1995), albeit at lower magnitudes. Although the low magnitude of completeness may be globally unique, it does not seem to reflect an unusual magnitude distribution. As with studies in South African gold mines (e.g., Richardson and Jordan, 2002; Kwiątek *et al.*, 2010), this may indicate that the Gutenberg–Richter relationship holds to magnitudes as low as $M_L -1.5$.

The recent microseismicity and the 1969 aftershocks have spatial distributions within error of each other; therefore, for this area, we find that the distribution of seismicity is independent of magnitude range and of any differences between aftershocks and continuous seismicity. It also appears that the distribution of seismicity has remained unchanged for over 40 years. This implies a long-term (at least on human time scales) localization of intraplate seismicity to this microseismic zone.

Neither the active microseismic zone nor the 1969 inferred rupture plane and observed aftershock zone show a correlation to a mapped fault surface trace. However, the strike of the microseismic zone is typical of reactivated Pan-African frictional–viscous shear zones in western South Africa (Viola *et al.*, 2012). The change in V_P from 5.4 to 5.8 km/s at the 6 km depth layer (Fig. 3) is interpreted to represent the Namaqua metamorphic complex below the Malmesbury group. For the three months of recording, active seismicity fell in two distinct spatial clusters, where depths are dominantly less than 6 km in the southeast and greater than 6 km in the northwest. If the velocity model is interpreted correctly, this means the southern cluster falls within the Malmesbury group and Cape supergroup low-grade metasedimentary rocks, whereas the northern cluster lies in the underlying Namaqua metamorphic complex. The 1969 event, at nearly 15 km depth (Krüger and Scherbaum, 2014), therefore likely occurred within the Namaqua metamorphic complex basement rocks. The maximum depth of seismicity of 15 km is interpreted as the base of the seismogenic zone in the Ceres region. This depth is typical of the frictional–viscous transition in quartzofeldspathic crust with a typical stable continental geothermal gradient of 25° – 30°C (e.g., Sibson, 1982; Scholz, 1988).

Stress measurements from borehole breakouts (Andreoli *et al.*, 1996; Viola *et al.*, 2005) and thin-shell rheological modeling (Bird *et al.*, 2006) indicate that the southwestern coastal margin of South Africa experiences a northwest–southeast-trending greatest horizontal stress. This stress direction is roughly parallel to the strike of the surface trace of the Ceres–Tulbagh microseismic zone reported here. Viola *et al.* (2012) inferred southwest African stress evolution constrained by fracture systems and suggest a north-northwest-trending greatest horizontal stress for the most recent stress field along the coast of southwest Africa. Inland, the greatest horizontal stress appears to turn to a west-northwest–east-southeast trend (Viola *et al.*, 2005; Bird *et al.*, 2006). If we accept this stress field for the Ceres region, the subvertical microseismically active zone is relatively well oriented, but not ideal, for sinistral strike-slip faulting. Its orientation is

Figure 4. Location of all microseismic events observed in this study (open circles). Clusters discussed in the text are outlined in solid (Ceres) and dashed (Tulbagh) lines. The gray-filled circles show the aftershocks from the 1969 M_w 6.3 event, and the large open circle shows the 1969 mainshock (Green and Bloch, 1971). The depth of the 1969 event is estimated just shallower than the maximum depth of 15 km from Krüger and Scherbaum (2014) but may have been shallower. Station locations for the current study are shown by open triangles. The color version of this figure is available only in the electronic edition.

Figure 5. Frequency-magnitude distribution of microseismic events. Note the very low magnitude of completion of M_L -1.5 and distribution fitting of a Gutenberg-Richter relationship with a b -value of 0.89.

subparallel to pre-existing features in the Malmesbury group and in the Namaqua-Natal metamorphic belt (Fig. 1). It therefore seems likely that the microseismic zone represents reactivation of long-lived structural elements underlying, in this area (Fig. 2), the relatively thin surface exposure of Cape supergroup rocks. This model is comparable to the hypothesis that the New Madrid seismic zone is related to reactivation of old rift structures (e.g., Johnston and Schweig, 1996), except in Ceres-Tulbagh the reactivated faults are steep strike-slip faults inherited from Pan-African transpression (Viola *et al.*, 2012). There is no evidence for increased heat

flow in the Ceres-Tulbagh region compared with surrounding crust (nor is there reason to suspect such increased heat flow), so we do not invoke higher temperature as a reason for fault zone weakening.

A question remains of why this region sees such localized seismicity, when there are numerous faults of similar orientation elsewhere in western South Africa. As in New Madrid, there may be a temporal stress reason for this, but this seems unlikely, as there are no recent changes in erosion rates or uplift, such as related to removal of ice sheets in North America. Another possibility is spatial variation in fault weakness. We know little of the structure and composition of Pan-African shear zones at depth, but high electric conductivity and magnetic susceptibility along the southern extent of the Namaqua metamorphic complex (the southern Cape conductive belt) have been interpreted as serpentinized shear zones and a reason for a long-lived weak zone here (De Beer *et al.*, 1982). Alternatively, Rennie *et al.* (2013) described a Namaqua-Natal metamorphic belt shear zone in detail and explained its relative weakness by interconnected, fine-grained phyllosilicates oriented subparallel to the shear zone margins.

The source of driving stress for the intraplate seismicity in Ceres-Tulbagh is also ambiguous. The strike-slip faulting in Ceres-Tulbagh contrasts with the normal-faulting regime that is observed in central and northeast South Africa, which can be related to topographically controlled gravitational potential energy and effects of the southward propagation of the East African rift system (Coblentz and Sandiford, 1994; Delvaux and Barth, 2010). The northwest-southeast-trending greatest horizontal stress observed on the southwest coast, however, does not correlate clearly with models of extension caused by either rift propagation (Bird *et al.*, 2006) or gravitational potential energy (Coblentz and Sandiford, 1994), and

thus another explanation is required. Possible sources of a driving stress include stresses from the oceanic spreading ridges surrounding the southern part of the continent (Viola *et al.*, 2005; Brandt and Saunders, 2011) or relative rotation of the Somali and Nubian plates, with the cold and rigid lithosphere underlying southern Africa posing resistance to deformation (Bird *et al.*, 2006).

Our observations emphasize that the Ceres–Tulbagh intraplate fault zone still experiences localized seismic activity over 40 years after the last major intraplate earthquake. The microseismic zone lies at the boundary between the high plateau of southern Africa and the surrounding oceans, and it therefore may relate to a complex stress field involving plate boundary ridge push and gravitational potential energy from the plate interior. These stresses, however, would not create a localized zone of crustal deformation, but rather lead to distributed faulting. Therefore, it seems that the critical control on the location of intraplate seismicity in this part of the African plate is the presence of a pre-existing crustal weakness within which deformation is localized. The details of the nature of this weakness are still not clear, although they appear related to a steep shear zone like those typically inherited from Pan-African orogenic belts in this part of southern Africa. As pointed out by England and Jackson (2011) in other areas, the lack of a surface rupture highlights the seismic hazard related to buried structures in intraplate regions.

Data and Resources

All the data presented in this article were collected by a local seismic network in February–May 2011 and have not been published previously. © A full list of event details is provided in the electronic supplement.

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References

- Abercrombie, R. (1995). Earthquake source scaling relationships from -1 to $5 M_L$ using seismograms recorded at 2.5-km depth, *J. Geophys. Res.* **100**, 24,015–24,036.
- Andreoli, M. A. G., M. Doucote, J. van Bever Donker, D. Brandt, and N. J. B. Andersen (1996). Neotectonics of southern Africa—A review, *Afr. Geosci. Rev.* **3**, 1–16.
- Belcher, R. W., and A. F. M. Kisters (2003). Lithostratigraphic correlations in the western branch of the Pan-African Saldania belt, South Africa: The Malmesbury group revisited, *S. Afr. J. Geol.* **106**, 327–342.
- Bird, P., Z. Ben-Avraham, G. Schubert, M. A. G. Andreoli, and G. Viola (2006). Patterns of stress and strain rate in South Africa, *J. Geophys. Res.* **111**, no. B08402, doi: [10.1029/2005JB003882](https://doi.org/10.1029/2005JB003882).
- Brandt, M. B., and I. Saunders (2011). New regional moment tensors in South Africa, *Seismol. Res. Lett.* **82**, 69–80.
- Calais, E., and S. Stein (2009). Time-variable deformation in the New Madrid seismic zone, *Science* **323**, 1442.
- Calais, E., A. M. Freed, R. Van Arsdale, and S. Stein (2010). Triggering of New Madrid seismicity by late-Pleistocene erosion, *Nature* **466**, 608–611.
- Campbell, D. L. (1978). Investigation of the stress-concentration mechanism for intraplate earthquakes, *Geophys. Res. Lett.* **5**, 477–479.
- Coblentz, D. D., and M. Sandiford (1994). Tectonic stresses in the African plate: Constraints on the ambient lithospheric stress state, *Geology* **22**, 831–834.
- Collettini, C., C. Viti, S. A. F. Smith, and R. E. Holdsworth (2009). Development of interconnected talc networks and weakening of continental low-angle normal faults, *Geology* **37**, 567–570.
- Colliston, W. P., A. E. Schoch, and J. Cole (2014). The Grenvillian Namaqua-Natal fold belt adjacent to the Kaapvaal Craton: 1. Distribution of Mesoproterozoic collisional terranes deduced from results of regional surveys and selected profiles in the western and southern parts of the fold belt, *J. African Earth Sci.* **100**, 7–19.
- Crone, A. J., P. M. De Martini, M. N. Machette, K. Okumura, and J. R. Prescott (2003). Paleoseismicity of two historically quiescent faults in Australia: Implications for fault behavior in stable continental regions, *Bull. Seismol. Soc. Am.* **93**, 1913–1934.
- Dalziel, I. W. D., S. Mosher, and L. M. Gahagan (2000). Laurentia-Kalahari collision and the assembly of Rodinia, *J. Geol.* **108**, 499–513.
- de Beer, C. H. (1992). Structural evolution of the Cape fold belt syntaxis and its influence on syntectonic sedimentation in the SW Karoo basin, in *Inversion Tectonics of the Cape Fold Belt, Karoo and Cretaceous Basins of Southern Africa*, M. J. de Wit and I. G. D. Ransome (Editors), A. A. Balkema, Rotterdam, The Netherlands, 197–206.
- De Beer, J. H., J. S. V. van Zijl, and D. I. Gough (1982). The southern Cape conductive belt (South Africa): Its composition, origin and tectonic significance, *Tectonophysics* **83**, 205–225.
- Delvaux, D., and A. Barth (2010). African stress pattern from formal inversion of focal mechanism data, *Tectonophysics* **482**, 105–128.
- England, P., and J. Jackson (2011). Uncharted seismic risk, *Nat. Geosci.* **4**, 348–349.
- Fagereng, A. (2013). On stress and strain in a continuous-discontinuous shear zone undergoing simple shear and volume loss, *J. Struct. Geol.* **50**, 44–53.
- Fagereng, A., F. Remitti, and R. H. Sibson (2010). Shear veins observed within anisotropic fabric at high angles to the maximum compressive stress, *Nat. Geosci.* **3**, 482–485.
- Frimmel, H. E., and W. Frank (1998). Neoproterozoic tectono-thermal evolution of the Gariep belt and its basement, Namibia/South Africa, *Precambrian Res.* **90**, 1–28.
- Goodwin, L. B., and B. Tikoff (2002). Competency contrast, kinematics, and the development of foliations and lineations in the crust, *J. Struct. Geol.* **24**, 1065–1085.
- Green, R. W. E., and S. Bloch (1971). The Ceres, South Africa, earthquake of September 29, 1969: I. Report on some aftershocks, *Bull. Seismol. Soc. Am.* **61**, 851–859.
- Green, R. W. E., and A. McGarr (1972). A comparison of the focal mechanism and aftershock distribution of the Ceres, South Africa earthquake of September 29, 1969, *Bull. Seismol. Soc. Am.* **62**, 869–871.
- Gresse, P. G., and J. N. Theron (1992). *The Geology of the Worcester Area*, Geological Map Sheet 3319, Council for Geoscience, Pretoria, scale 1:250,000.
- Grollmund, B., and M. D. Zoback (2001). Did deglaciation trigger intraplate seismicity in the New Madrid seismic zone, *Geology* **29**, 175–178.
- Havskov, J., and L. Ottemöller (2010). *Routine Data Processing in Earthquake Seismology*, Springer, New York, New York, 347 pp.
- Hildebrand, T. G., J. G. Rosenbaum, and R. L. Reynolds (1992). High-resolution aeromagnetic study of the New Madrid seismic zone: A preliminary report, *Seismol. Res. Lett.* **63**, 209–222.
- Holdsworth, R. E., R. A. Strachan, J. Magloughlin, and R. J. Knipe (2001). *The Nature and Tectonic Significance of Fault Zone Weakening*, Spe-

- cial Publication, Vol. 186, Geological Society of London, Bath, United Kingdom, 344 pp.
- Hough, S. E. (2009). Cataloging the 1811–1812 New Madrid, central U.S., earthquake sequence, *Seismol. Res. Lett.* **80**, 1045–1053.
- Hubbert, M. K., and W. W. Rubey (1959). Role of fluid pressure in the mechanics of overthrust faulting, *Bull. Geol. Soc. Am.* **70**, 115–206.
- Hurd, O., and M. D. Zoback (2012). Intraplate earthquakes, regional stress and fault mechanics in the central and eastern U.S. and southeastern Canada, *Tectonophysics* **581**, 182–192.
- Johnston, A. C., and E. S. Schweig (1996). The enigma of the New Madrid earthquakes of 1811–1812, *Annu. Rev. Earth Planet. Sci.* **24**, 339–384.
- Kissling, E., W. L. Ellsworth, D. Eberhart-Phillips, and U. Kradolfer (1994). Initial reference models in local earthquake tomography, *J. Geophys. Res.* **99**, 19,635–19,646.
- Krüger, F., and F. Scherbaum (2014). The 29 September 1969, Ceres, South Africa, earthquake: Full waveform moment tensor inversion for point source and kinematic source parameters, *Bull. Seismol. Soc. Am.* **104**, 576–581.
- Kwiatek, G., K. Plenkers, M. Nakatani, Y. Yabe, G. Dresen, and the JAGUARS Group (2010). Frequency–magnitude characteristics down to magnitude –4.4 for induced seismicity recorded at Mponeng Gold Mine, South Africa, *Bull. Seismol. Soc. Am.* **100**, 1165–1173.
- Li, Q., M. Liu, and S. Stein (2009). Spatiotemporal complexity of continental intraplate seismicity: Insights from geodynamic modelling and implications for seismic hazard estimation, *Bull. Seismol. Soc. Am.* **99**, 52–60.
- Liu, L., and M. D. Zoback (1997). Lithospheric strength and intraplate seismicity in the New Madrid seismic zone, *Tectonics* **16**, 585–595.
- Lock, B. E. (1980). Flat-plate subduction and the Cape fold belt of South Africa, *Geology* **8**, 35–39.
- Mancktelow, N., and G. Pennacchioni (2005). The control of precursor brittle fracture and fluid-rock interaction on the development of single and paired ductile shear zones, *J. Struct. Geol.* **27**, 645–661.
- McKenna, J., S. Stein, and C. A. Stein (2007). Is the New Madrid Seismic Zone hotter and weaker than its surroundings? *Geol. Soc. Am. Spec. Pap.* **425**, 167–175.
- Montesi, L. G. J. (2013). Fabric development as the key for forming ductile shear zones and enabling plate tectonics, *J. Struct. Geol.* **50**, 254–266.
- Newman, A., S. Stein, J. Weber, J. Engeln, A. Mao, and T. Dixon (1999). Slow deformation and lower seismic hazard at the New Madrid Seismic Zone, *Science* **284**, 619–621.
- Niemeijer, A. R., and C. J. Spiers (2006). Velocity dependence of strength and healing behaviour in phyllosilicate-bearing fault gouge, *Tectonophysics* **427**, 231–253.
- Rennie, S. F., A. Fagereng, and J. F. A. Diener (2013). Strain distribution within a km-scale, mid-crustal shear zone: The Kuckaus mylonite zone, Namibia, *J. Struct. Geol.* **56**, 57–69.
- Rice, J. R. (1992). Fault stress states, pore pressure distributions, and the weakness of the San Andreas fault, in *Fault Mechanics and Transport Properties of Rocks*, B. Evans and T. F. Wong (Editors), Academic Press, San Diego, California, 475–503.
- Richardson, E., and T. H. Jordan (2002). Seismicity in gold mines of South Africa: Implications for tectonic earthquakes, *Bull. Seismol. Soc. Am.* **92**, 1766–1782.
- Rodgers, J. J. W., and M. Santosh (2003). Supercontinents in Earth's history, *Gondwana Res.* **6**, 357–368.
- Rozendaal, A., P. G. Gresse, R. Scheepers, and J. P. Le Roux (1999). Neoproterozoic to Early Cambrian crustal evolution of the Pan African Saldania belt, South Africa, *Precambrian Res.* **97**, 303–323.
- Saunders, I., L. Ottemöller, M. B. Brandt, and C. J. Fourie (2012). Calibration of an M_L scale for South Africa using tectonic earthquake data recorded by the South African National Seismograph Network: 2006 to 2009, *J. Seismol.* **17**, 437–451.
- Scholz, C. H. (1988). The brittle-plastic transition and the depth of seismic faulting, *Geol. Rundsch. Z. Allg. Geol.* **77**, 319–328.
- Schweig, E. S., and M. A. Ellis (1994). Reconciling short recurrence intervals with minor deformation in the New Madrid Seismic Zone, *Science* **264**, 1308–1311.
- Sibson, R. H. (1982). Fault zone models, heat flow, and the depth distribution of earthquakes in the continental crust of the United States, *Bull. Seismol. Soc. Am.* **72**, 151–163.
- Singh, M., A. Kijko, and R. Durrheim (2009). Seismotectonic models for South Africa: Synthesis of geoscientific information, problems, and the way forward, *Seismol. Res. Lett.* **80**, 71–80.
- Stein, S., and J. T. Freymuller (2002). *Plate Boundary Zones: Concepts and Approaches*, American Geophysical Union, Washington, D.C., 425 pp.
- Stein, S., M. Liu, E. Calais, and Q. Li (2009). Mid-continent earthquakes as a complex system, *Seismol. Res. Lett.* **80**, 551–553.
- Stevenson, D., A. Gangopadhyay, and P. Talwani (2006). Booming plutons: Source of microearthquakes in South Carolina, *Geophys. Res. Lett.* **33**, L03316, doi: [10.1029/2005GL024679](https://doi.org/10.1029/2005GL024679).
- Tham, A. G., and M. R. Johnson (2009). The Cape supergroup, in *The Geology of South Africa*, M. R. Johnson, C. R. Anhaeusser, and R. J. Thomas (Editors), Geological Society of South Africa and the Council of Geoscience, Pretoria, South Africa, 443–460.
- Thomas, W. A. (1991). The Appalachian-Ouachita rifted margin of southeastern North America, *Bull. Geol. Soc. Am.* **103**, 415–431.
- Tuttle, M., E. S. Schweig, J. D. Sims, R. H. Lafferty, L. W. Wolf, and M. L. Haynes (2002). The earthquake potential of the New Madrid Seismic Zone, *Bull. Seismol. Soc. Am.* **92**, 2080–2089.
- Viola, G., M. A. G. Andreoli, Z. Ben-Avraham, I. Stengel, and M. Reshef (2005). Offshore mud volcanoes and onland faulting in southwestern Africa: Neotectonic implications and constraints on the regional stress field, *Earth Planet. Sci. Lett.* **231**, 147–160.
- Viola, G., A. Kounov, M. A. G. Andreoli, and J. Mattila (2012). Brittle tectonic evolution along the western margin of South Africa: More than 500 Myr of continued reactivation, *Tectonophysics* **514**, 93–114.
- Wintsch, R. P., R. Christofferson, and A. K. Kronenberg (1995). Fluid-rock reaction weakening of fault zones, *J. Geophys. Res.* **100**, 13,021–13,032.
- Wu, P., and P. Johnston (2000). Can deglaciation trigger earthquakes in North America? *Geophys. Res. Lett.* **27**, 1323–1326.
- Zoback, M. D. (2010). Climate and intraplate shocks, *Nature* **466**, 568–569.
- Zoback, M. D., and M. L. Zoback (1981). State of stress and intraplate earthquakes in the United States, *Science* **213**, 96–104.
- Zoback, M. D., W. H. Prescott, and S. W. Krueger (1985). Evidence for lower crustal ductile strain localization in southern New York, *Nature* **317**, 705–707.
- Zoback, M. L. (1992). First hand second-order patterns of stress in the lithosphere: The World Stress Map Project, *J. Geophys. Res.* **97**, 11,703–11,728.

Department of Geological Sciences
University of Cape Town
Private Bag X3
Rondebosch 7701
South Africa
(L.S., Å.F.)

Deutsches GeoForschungsZentrum
14473 Potsdam, Germany
(B.B.)

European Center for Geodynamics and Seismology
Rue Josy Welter 19
L-7256 Walferdange
Grand Duchy of Luxembourg
(J.S.)