4   **The Sagaing Fault**

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13   **Abbreviated title:** Sagaing Fault

14   The Sagaing Fault is amongst the longest and most active strike-slip faults in the world (e.g. Molnar
15   & Dayem 2010; Robinson *et al.* 2010; Searle & Morley 2011). It accommodates more than half of the
16   right-lateral motion between Sundaland and India, within the diffuse plate boundary along the eastern
17   margin of India, which occupies much of Myanmar (e.g. Vigny *et al.* 2003; Nielsen *et al.* 2004;
18   Socquet *et al.* 2006) (Fig. 1). Acting as a ridge-subduction transform (Le Dain *et al.* 1984; Guzmán-
19   Speziale & Ni 1996; Yeats *et al.* 1997), the 1500 km long Sagaing Fault links major thrust systems in
20   the north, such as the Naga, Lohit and Main Central thrust zones near the eastern Himalayan syntaxis,
21   to the Andaman Sea spreading centre in the south (e.g. Win Swe 1970; Le Dain *et al.* 1984; Guzmán-
22   Speziale & Ni 1993; Curray 2005). Acting as a sliver-bounding lithospheric strike-slip partition
23   inboard of a 3700 km long section of oblique subduction at the Sunda Trench, it links to the dextral
24   West Andaman and Sumatran faults in the south (e.g. Curray *et al.* 1979; Nielsen *et al.* 2004; Curray
25   2005; Searle & Morley 2011).

26   Topographic scarps, lineaments, earthquake clusters and gravity anomalies along the trace of the
27   Sagaing Fault have long been recognised as indicating the presence of an important N-S trending
28   structure west of the Shan Plateau (e.g. La Touché 1913; Coggin Brown & Leicester 1933; Chhibber
29   1934; Dey 1968; Aung Khin *et al.* 1970) (Fig. 2a). Large parts of the fault were first documented and
30   pressure ridges, sag ponds and right-lateral offsets interpreted as signals of a major strike-slip fault by
31   Win Swe (1970, 1972), who also introduced the name *Sagaing Fault*. Further research during the last
32   two decades has begun to reveal the fault's profound role in the tectonic evolution of south and
33   southeast Asia (e.g. Bertrand *et al.* 2001; Vigny *et al.* 2003), and the great hazard it poses to the
34   millions who live along its trace (e.g. Hurukawa & Phyo Maung Maung 2011; Wang *et al.* 2011).

Tectonic setting

Myanmar straddles the complex oblique plate boundary between the India-Australia plate and Sundaland, the SE promontory of Eurasia (e.g. Hall & Morley 2004; Nielsen *et al.* 2004; Curray 2005; Kundu & Gahalaut 2012) (Fig. 1). The Sunda trench and arc, related to the down-going Indian plate, become progressively more parallel to Indian plate motion northwards past Sumatra and into the Indo-Myanmar ranges (e.g. Nielsen *et al.* 2004; Whittaker *et al.* 2007). In the south, simple strain partitioning between the Sumatra trench and the Sumatran Fault (e.g. McCaffrey 1996) gives way to a wide zone of distributed deformation in Myanmar between the Andaman Sea and the eastern Himalayan syntaxis in the north (e.g. Vigny *et al.* 2003; Nielsen *et al.* 2004; Sahu *et al.* 2006). Contraction is accommodated in the Indo-Myanmar ranges in the west, while 18-20 mm/yr of a total of 35 mm/yr right lateral motion between India and Sundaland is partitioned onto the Sagaing Fault, with the remainder distributed across dextral structures within the Indo-Myanmar ranges and far inboard of the trench (e.g. Molnar & Tapponnier 1975; Curray *et al.* 1979; Tapponnier *et al.* 1982; Le Dain *et al.* 1984; Mitchell 1993; Vigny *et al.* 2003). The Sagaing Fault thus defines the western margin of Sundaland and the eastern margin of the Neogene Burma Plate, a fore arc sliver which occupies much of western Myanmar, is partially coupled to the Indian plate, and is bounded by active tectonic structures (Curray *et al.* 1979).

Conversely, the Sagaing Fault lies close to but does not mark the western margin of the Gondwana-derived Sibumasu continental block which underlies eastern Myanmar (Ridd 1971; Bunopas 1981; Metcalfe 1984). The western margin of that block in Myanmar is marked at the Shan Scarp by the Mogok Metamorphic Belt (e.g. Searle & Ba Than Haq 1964; Mitchell *et al.* 2007) (Fig. 2b), and the Slate Belt - a sliver of late Palaeozoic glacial marine pebbly mudstones that comprise the Karen-Tenasserim Unit of Bender (1983) or the Mergui Group of Mitchell *et al.* (2002, 2007) that may have been thrust onto the Sibumasu margin in the Late Triassic to Early Jurassic (Mitchell 1992) or translated by a system of dextral faults east of the present-day Sagaing Fault during the Late Cretaceous to Paleogene (Ridd & Watkinson 2013). It remains unclear how far such continental basement continues west of the Sagaing Fault, where it is mostly buried below thick siliciclastic sequences of the Myanmar Central Basin (e.g. Stephenson & Marshall 1984; Pivnik *et al.* 1998; Win Swe 2012).

West of the metamorphics along the margin of Sibumasu, the basement is poorly known. A West Burma block (which is not the same as the Sagaing Fault-bounded Neogene Burma Plate) has been interpreted as an island arc thrust onto the western margin of SE Asia during the Middle Jurassic (Mitchell 1981) or Late Cretaceous to Eocene (e.g. Mitchell 1992, 1993). The same block has also been viewed as a fragment of Sibumasu separated in the Triassic and re-accreted during the Early Cretaceous (e.g. Hutchinson 1989), a tectonic slice of the Indochina terrane translated during the Triassic to the western margin of Sibumasu (e.g. Barber & Crow 2009), or an Australian fragment that collided with Sibumasu in the Early Cretaceous (Metcalfe 1996) or Late Cretaceous to Eocene (e.g. Charusiri *et al.* 1993; Hutchison 1994). However, recent work on the provenance of detrital zircons within Triassic turbidites in western Myanmar suggests that West Burma was in place alongside SE Asia before the Mesozoic (Sevastjanova *et al.* 2015), or that Sibumasu extends westwards to the Western Ophiolite Belt suture (Gardiner *et al.* 2015).

The Sagaing Fault is closely associated with a number of Cenozoic structures within central Myanmar (Fig. 2a). Many of these structures remain active (e.g. Coggin Brown 1917; Wang *et al.* 2014), yet many probably pre-date the Sagaing Fault and may have played a role in the Neogene to Recent evolution of the fault (e.g. Morley 2012).

80   Immediately east of the Sagaing Fault between Mandalay and the Gulf of Mottama, the Shan Scarp
81   fault zone forms the prominent eastern topographic margin of the Shan Plateau and appears to be
82   obliquely truncated by the Sagaing Fault (Bertrand & Rangin 2003) at about 23°N. Composed of a
83   post-Middle/Upper Miocene dextral transpressional fold and thrust belt trending 140°-160°, the Shan
84   Scarp fault zone places pre-Cenozoic rocks of the Shan Plateau over Mogok Metamorphic Belt
85   gneisses at the margin of the Central Basin (Bertrand & Rangin 2003) (Fig. 2b). The southernmost
86   segment of the fault zone includes the Panlaung and Taungoo faults which bound the Phuket Slate
87   Belt and may have their origins as early Paleogene strike-slip splays of the Mae Ping Fault (e.g.
88   Garson *et al*. 1976; Morley 2004; Ridd & Watkinson 2013).

89   Lying between the Sagaing Fault and the Shan Scarp fault zone, the Mogok Metamorphic Belt
90   comprises a narrow strip of gneisses and schists which in the north curves around the Bhamo Basin,
91   and may be continuous with northern Thailand metamorphic core complexes such as Doi Suthep and
92   Doi Inthanon, and the Lhasa terrane in south Tibet (Mitchell 1993; Mitchell *et al*. 2007, 2012; Searle
93   *et al*. 2007). Like other areas close to the India-Asia suture (e.g. Karakoram/Lhasa/Qiangtang blocks),
94   the Mogok Metamorphic Belt experienced Late Cretaceous pre-collisional magmatic thickening,
95   Paleogene post-collisional thickening, followed by high temperature metamorphism and leucogranitic
96   melting (Barley *et al*. 2003; Searle *et al*. 2007). An early phase of metamorphism is sealed by ~59 Ma
97   dykes, followed by high temperature Eocene-Oligocene metamorphism from 37 Ma (possibly even 47
98   Ma) to 29 Ma, and all metamorphic fabrics are sealed by 24.5 Ma ± 0.3 Ma leucogranites (Searle *et
99   al*. 2007). Gently plunging ductile stretching lineations representing NNW-SSE orientated
100   transtension overprint older fabrics, and yield mica Ar-Ar plateaux from 27-25 Ma in the south to
101   22.7-16.6 Ma in the north (Bertrand *et al*. 1999; 2001). Apatite fission track ages range from 18.7-
102   14.6 Ma (Torres *et al*. 1997), supporting a period of Late Oligocene-Mid-Miocene cooling and
103   exhumation.

104   East of the Shan Scarp fault zone is the N-S trending dextral Kyaukkyan Fault, which passes through
105   the Inle Lake pull-apart system and curves into the NW-SE trending Mae Ping Fault in the southeast
106   (Soe Min *et al*. *this volume*). The Kyaukkyan Fault produced perhaps the largest recorded earthquake
107   in Myanmar (M_w 8.0 (Gutenberg & Richter, 1954)) centred near Pyin-Oo-Lwin in 1912 (Coggin
108   Brown 1917), though it has generated little seismicity since then. Like the presently dextral Mae Ping
109   Fault in Thailand (Lacassin *et al*. 1997), the Kyaukkyan Fault shows evidence of slip sense reversal
110   after an earlier phase of sinistral activity (Wang *et al*. 2014), which may indicate that it, like the Mae
111   Ping Fault, is a longer-lived structure than the continuously dextral Sagaing Fault.

112   Near the northern termination of the Kyaukkyan Fault east of Mogok, a series of apparently left-
113   lateral faults, including the Kyaukme and Momeik (*Nanting* in China) faults, trend ENE-WSW and
114   terminate adjacent to the Sagaing Fault near Thabeikkyin. These arcuate faults, like the parallel Nam
115   Ma Fault further east which generated the 2011 M_w 6.8 Tarlay earthquake, are part of the broad Shan
116   fault system (Soe Thura Tun *et al*. 2014), which may partially accommodate clockwise flow of
117   continental crust around the eastern Himalayan syntaxis (e.g. Royden 1996; Copley & McKenzie
118   2007). The faults' orientations are broadly related to GPS vectors in a stable Eurasia reference frame
119   (e.g. Niu *et al*. 2005). Correspondingly the most northerly members of this fault population, such as
120   the Ruili Fault, attain a more NE-SW trend and become sub-parallel to the termination splays of the
121   Sagaing Fault on the far side of the Bhamo Basin, though their sense of displacement is opposite.

122   Occupying the 200 km wide valley between the Shan Scarp in the east and the Indo-Myanmar ranges
123   in the west, the Myanmar Central Basin represents a postulated fore-arc/back-arc basin filled largely
124   with Cenozoic shallow marine and terrestrial sediments, with major depocentres bounded in the east

125 by the Sagaing Fault (e.g. Pivnik *et al*. 1998). The basin's linked *en echelon* sub-basins include (from
126 north to south): the Hukawng, Chindwin, Bhamo, Salin and Prome basins, which record NNW-SSE
127 directed extension (Pivnik et al. 1998) coeval with fabrics recording ductile transtension in the Mogok
128 Metamorphic Belt (Bertrand & Rangin 2003). The basin also contains calc-alkaline andesite-dacite
129 volcanics related to east-dipping subduction below the basin, that had likely ceased by Mid-Miocene
130 times (e.g. Lee *et al*. 2010; Searle & Morley 2011; Gardiner *et al*. 2015). Late Miocene-Recent basin
131 inversion resulted in reactivation of the extensional faults, and development of a series of NNW-SSE-
132 trending *en echelon* folds within the basin (e.g. Pivnik *et al*. 1998; Rangin *et al*. 1999). One prominent
133 area of inversion-related uplift is Pegu Yoma, hilly terrain stretching from Bago (formerly *Pegu*) in
134 the south almost to Sagaing in the north. It is bounded in the east by linear segments of the Sagaing
135 Fault. Along its western margin thrusts such as the West Pegu Yoma and Gwecho faults (Soe Thura
136 Tun & Maung Thein 2012) are part of a suite of *en echelon* thrusts and associated hangingwall
137 anticlines that occupy the low relief western half of the basin. The modern Ayeyarwaddy River passes
138 around the north and west of Pegu Yoma, and appears to have been diverted from an original route
139 along the southern Sagaing Fault (now occupied by the Sittaung River) by uplift of the hills.
140 Bounding structures along the west side of the Central Basin include dextral strike-slip, oblique and
141 thrust faults such as the Seindaung, Kabaw and East Limb faults (e.g. Hla Maung 1987; Searle &
142 Morley 2011; Wang *et al*. 2014).

Tectonic geomorphology

144 For much of its onshore length the Sagaing Fault passes through low topography west of the Shan
145 Scarp. Unlike the sinuous Shan Scarp fault zone across which the ground rises over 1 km in a short
146 distance, the Sagaing Fault is expressed by more subtle but highly linear topographic features that are
147 only sometimes associated with significant topographic relief (e.g. Dey 1968; Le Dain *et al*. 1984).
148 Major geomorphic domains of the Sagaing Fault comprise the Sittaung/Ayeyarwaddy deltas in the
149 south, the eastern scarp of Pegu Yoma, the palaeo-Ayeyarwaddy valley, the modern Ayeyarwaddy
150 valley and the northern Basin and Range topographic domain (Fig. 3).

151 Overlapping deltas of the Sittaung, Ayeyarwaddy and Bago rivers, which rise to elevations rarely
152 greater than 10 m, are crossed by networks of meandering rivers, abandoned channels and ox-bow
153 lakes. Frequent avulsions combined with intense agricultural activity mostly erase evidence of active
154 faulting. On the delta top south of Bago, several low linear pressure ridges (<40 m elevation) lie along
155 the 1-2 km wide zone between the Sagaing Fault and a parallel strand, the Pale Fault (Tsutsumi &
156 Sato 2009) (Fig. 4a). Elsewhere along the Bago segment the fault is marked by remarkably straight
157 geomorphic lineaments defined by small linear sag ponds and changes in soil moisture and vegetation
158 several tens of metres either side of the fault (Fig. 5a). In places, soil is anomalously moist (dark
159 colour, solid deep green vegetation, lines of trees), reflecting springs or poorly drained sag ponds
160 filled with lacustrine deposits; or it is anomalously well drained (light mottled colour, patchy
161 vegetation), a result of accumulated coseismic sand blows or sandy sag pond fill. In the area around
162 Bago, fault scarps up to 2.8 m high lie along the trend of several low ridges and sag ponds. Streams,
163 terrace risers, rice paddy dikes, property boundaries, and a wall show right-lateral offsets of up to 18.6
164 ± 0.5 m, accumulated during several historic earthquakes, including the May 1930 M_w 7.2 event
165 (Tsutsumi & Sato 2009).

166 The eastern scarp of Pegu Yoma along the Pyu segment is where the Sagaing Fault has its clearest and
167 most simple geomorphic expression (Fig. 5b). The scarp rises almost 450 m above the Sittaung valley
168 in the east, is highly linear and is crossed by numerous streams displaying right-lateral offset. This

169   section was first compared to the San Andreas Fault in California by Dey (1968), and the stream
170   offsets are as regular and well defined as they are along the famous Carrizo Plain section of the San
171   Andreas Fault (e.g. Wallace 1968). A bajada of broad, low amplitude stream-flow-dominated alluvial
172   fans flanks the ridge, with the traces of palaeochannels clearly radiating from the stream source and
173   fan axes offset from the upstream incised stream. Very poorly developed triangular facets mark the
174   most linear sections of the scarp. In a number of places, notably NNW of Pyu, strips of the bajada are
175   uplifted >20 m adjacent to the fault, possibly by west-dipping reverse faults (Replumaz 1999). Incised
176   streams crossing the uplift occupy rounded, well vegetated gulleys, and larger rivers occupy wide
177   meander belts, suggesting that the uplift is either very slow or not presently active. Dextral river
178   offsets are well developed across the strike-slip segment, but not across the easterly reverse faults,
179   suggesting complete strain partitioning.

180   East of Meiktila the Sagaing Fault lies in flat topography of the palaeo-Ayeyarwaddy valley. A north-
181   plunging anticline alongside the fault at 20.4°N (Fig. 4b) near the northern limit of Pegu Yoma is
182   defined by relatively subtle topographic relief (~50 m). It lies at a gentle southward change in fault
183   trend from nearly N-S to NNW-SSE. The surface fault trace is highly segmented and sag ponds exist
184   at three right-step-overs no more than 0.5 km wide. A network of wide, meandering river channels
185   draining east from the fold axis are beheaded close to the fault trace, and are marked by darker, less
186   mottled farmland than surrounding areas. All the wide channels terminate at the fold, having
187   presumably been isolated from their upstream basins by growth of the fold. Deeply incised modern
188   tributaries on the eastern fold limb are offset to the right as they cross the fault. Abandoned meander
189   bends along the top of the fold indicate progressive northward fold propagation and deflection of the
190   river around the fold tip, clearly supporting young (Holocene?) dextral transpression along the
191   Sagaing Fault.

192   The Ayeyarwaddy River flows along or parallel to the Sagaing Fault for 180 km from Tagaung to
193   Sagaing city. The sediment-laden river occupies a sprawling system up to 10 km wide of braided
194   channels along the southern reach of this section, bounded in the west by the Sagaing Fault. The river
195   is briefly deflected west of the fault trace by Pleistocene basalts north of Singu (Bertrand *et al*. 1998).
196   The river's northern reach is a remarkably linear single channel rarely more than 1 km wide
197   dominated by erosion and bounded by linear fault-controlled topographic ridges. Steep-sided slopes,
198   triangular facets, (Fig. 4c) offset streams and deltas and down-cutting into Recent fluvial sands
199   characterise this reach. Two dry streams offset ~1.7-2.0 m near Thabeikkyin (Fig. 4d) include ~0.78 m
200   of local dextral displacement acquired during the 2012 M_w 6.8 earthquake, leaving 0.9-1.2 m which
201   can be attributed to a previous event, possibly the 1946 M_w 7.7 Wuntho earthquake. The linear
202   morphology of sand bars cut by recent Sagaing Fault activity in the Ayeyarwaddy River at Sabeanago
203   (Fig. 5c) reveal rapid along-strike changes in uplift/subsidence, a characteristic of strike-slip faulting.
204   Immediately south of a downthrown western block defining the modern river channel, the fault comes
205   onshore and is bounded in the west by Thein Taung ridge, where folded Neogene sandstones are
206   uplifted. Similar pressure ridges can be seen at a variety of stages of development along the
207   Ayeyarwaddy domain. While not quite fractal, there is an element of self-similarity in the morphology
208   of different sized ridges. At the smallest scale, embryonic ridges up to 0.5 m high in agricultural soils
209   deformed by the 2012 Thabeikkyin earthquake are bounded by a segmented strike-slip system on one
210   side, and an upwards flattening reverse fault on the other (Fig. 6a). A longer-lived pressure ridge in
211   Quaternary fluvial/lacustrine sediments (Fig. 6b) is 1.5 m high, has a length:width ratio of about 3:1,
212   and is again bounded on one side (east) by a strike-slip fault. Ridges north of Singu up to 25 m high
213   are bounded by transpressional strike-slip systems, exhume gently folded Irrawaddy Group sands
214   (Fig. 6c) and are locally capped by vesicular Singu basalt. An asymmetric shutter ridge exhuming

215   highly tectonised amphibolites north of the Singu basalt plateau is >100 m high (Fig. 6d), and has
216   over-steepened slopes scarred by landslides on the side bounded by the Sagaing Fault. A slightly
217   higher but highly elongate ridge parallel to the Ayeyarwaddy river is bounded on both sides by strike-
218   slip fault strands and has an almost symmetric topographic cross-section (Fig. 6e). All of these ridges'
219   long axes are either sub-parallel to the local Sagaing Fault trend, or are orientated <5° anticlockwise.
220   Where close together they are arranged *en echelon*.

221   The most prominent pressure ridge, Sagaing Ridge (Fig. 6f), rises 330 m above the adjacent braided
222   Ayeyarwaddy River, and is flanked by alluvial fans on its western side (e.g. Myint Thein *this volume*).
223   Sagaing Ridge tapers in the north to a slightly elevated flat-topped plateau. Large antecedent streams
224   flowing right across the plateau from west to east are deeply incised into the plateau. Smaller east
225   flowing streams originating on the plateau top represent modern drainage, and are characterised by
226   steep-sided canyons, slope failures and deltas where the streams enter the Ayeyarwaddy, indicating
227   energetic down-cutting as a result of recent uplift. The highly linear western margin of Sagaing Ridge
228   is marked by poorly developed triangular facets. A smaller, parallel ridge, Minwun Ridge, lies to the
229   west of the main fault strand. In the valley between the two ridges is Yega In, a linear lake which may
230   be a pull-apart basin, and a number of elongate sag ponds filled with wet, fine grained sediments. In
231   the valley immediately north of Sagaing city, a 1 m high west-facing, partly collapsed scarp cuts
232   across an embayment in the mountain front and may represent surface rupture and subsidence during
233   a historic earthquake (Fig. 4e). The eastern margin of Sagaing Ridge is more sinuous, perhaps because
234   bounding faults are less active, more segmented or dip at a shallower angle.

235   North of the point where the Ayeyarwaddy River is captured by the Sagaing Fault valley (at about
236   23.5°N), splays of the Sagaing Fault partly define a series of prominent, steep-sided curvilinear ranges
237   separated by basins of approximately equal width (Fig. 7). Elevations range from about 107 m in the
238   basins (e.g. Indaw Lake) to over 1400 m in the ranges. Ranges are bounded on one or both sides by
239   basin-dipping faults which are in some cases marked by triangular facets and flanked by alluvial fans.
240   In many cases modern axial drainage within the basins is asymmetric, suggesting tilting of the basin
241   floor towards the active fault. Such normal faulting and topography has similarities to regions of
242   extreme continental extension, for example basin and range domains of the western North American
243   Cordillera, central Turkey, south Tibet and northern Thailand, all regions of hot crust, crustal
244   thickening, uplift, collapse and extension (e.g. Coney & Harms 1984; Armijo *et al*. 1986; Wernicke *et*
245   *al*. 1987; Rhodes *et al*. 2005; Genç & Yürür 2010; Whitney *et al*. 2013). However, the apparent
246   localisation of modern tectonic activity along only one fault strand (e.g. Maurin *et al*. 2010) and the
247   presence of higher topography in the west, north and east beyond the region of prominent basins and
248   ranges indicates that the northern Sagaing Fault area is not a typical 'basin and range' region.

Segmentation

250   The Sagaing Fault is distinguished from other major strike-slip faults of similar length, such as the
251   Sumatran Fault and the Altyn Tagh Fault, by its straightness and apparent structural continuity (e.g.
252   Vigny *et al*. 2003). Straight and continuous faults can result from the removal of complexities such as
253   step-overs and rotation of pre-existing fabrics into parallelism with the through-going fault by high
254   cumulative displacements (e.g. Wesnousky 1988; Stirling *et al*. 1996; King & Wesnousky 2007). The
255   range of high displacement estimates for the Sagaing Fault, ranging from 100-150 km (Bertrand &
256   Rangin 2003) to 460 km (Curray *et al*. 1979; Hla Maung 1987) and even 700 km (Replumaz &
257   Tapponnier 2003), support the conclusion that the Sagaing Fault has experienced considerable strain
258   smoothing. In contrast, the similar length North Anatolian Fault is highly segmented and possesses

259 several wide rupture-arresting restraining and releasing step-overs along its length (e.g. Lettis *et al*.
260 2002; Duman *et al*. 2005). Estimates of its displacement (25-85 km, e.g. Armijo *et al*. 1999) are up to
261 an order of magnitude less than those proposed for the Sagaing Fault, which are discussed in the next
262 section.

263 Significant step-overs and bends along the central 700 km section of the Sagaing Fault are poorly
264 documented, if they exist at all. Hurukawa and Phyo Maung Maung (2011) suggested that the Sagaing
265 Fault is segmented based on the occurrence of M6 events at the termini of M7 earthquake-generating
266 segments. Small sag ponds and pressure ridges along the fault are common (e.g. Dey 1968; Tsutsumi
267 & Sato 2009) (Fig. 6), though it is unclear whether even the largest pressure ridge, Sagaing Ridge, can
268 be related to extant fault complexities, or has simply been translated along the fault from a
269 subsequently breached transpressive feature. Interpretation of satellite imagery suggests that step-
270 overs >3-5 km wide do not currently exist, and Robinson *et al*. (2010) consider that the central
271 Sagaing Fault entirely lacks steps wider than 1 km.

272 Here we describe Sagaing Fault segments based on interpretation of 30 m ASTER GDEM satellite
273 data, the ESRI World Imagery compilation, which includes 2.5 m SPOT and <1 m DigitalGlobe
274 imagery, and field observation of the fault. The fault interpretation incorporates previous mapping,
275 including Tsutsumi and Sato (2009), Soe Thura Tun and Maung Thein (2012) and Wang *et al*. (2014).
276 Numerous schemes have been proposed to divide the Sagaing Fault into structurally distinct
277 segments, a task complicated by the remarkable continuity of the central 700 km from 17-23° N (e.g.
278 Vigny *et al*. 2003). Wang *et al*. (2014) divided the onshore Sagaing Fault into five main segments,
279 plus seven additional segments within four active strands of the northern horsetail array, based on
280 fault geometry, geomorphic expression and historical seismicity. Here we broadly follow the work of
281 those authors, modified based on new satellite imagery interpretation and field observations (Fig. 2a
282 & 7).

283 *Northern segments and strike-slip termination*

284 Strike-slip along the Sagaing Fault terminates in a northern sickle-shaped horsetail splay largely
285 within Kachin State, comprising a paired right and left bend system almost 500 km from north to
286 south (Fig. 7). The sickle-shaped geometry is influenced by clockwise flow of the lower crust around
287 the eastern Himalayan syntaxis (e.g. Royden 1996; Copley & McKenzie 2007; Rangin *et al*. 2013).
288 Within this splay dextral strike-slip is transferred via a basin and range-style region of linear fault-
289 bounded basins and high ranges, such as the Katha-Gangaw range, exposing mid- and lower-crustal
290 units, to compressional structures around the eastern Himalayan syntaxis. As such the northern
291 horsetail splay is presently not strictly a termination splay, rather a splaying zone of strain transfer
292 converting strike-slip to extension at the releasing bend in the region of Myitkyina, and contraction at
293 the restraining bend in the region of Putao and beyond. The development of the present-day releasing
294 segments may be a consequence of rotation of earlier straight fault segments by gravitational collapse
295 of the Tibet Plateau (Rangin *et al*. 2013).

296 The major northern splay is composed of three (Maurin *et al*. 2010) or four (Wang *et al*. 2014) main
297 splay systems. Excluding the minor splays that originate at Singu (see below) the most westerly splay
298 is composed of the section of the Sagaing segment north of the prominent Ayeyarwaddy River bend,
299 termed Tawma segment by Wang *et al*. (2014). The Tawma segment steps to the left, via a series of *en*
300 *echelon* faults, to the Ban Mauk segment in the northwest, which itself terminates in a prominent
301 releasing horsetail splay east of Taungthonton volcano (Fig. 8a). Although the Ban Mauk segment

302 occupies a deeply incised narrow valley, geomorphology suggests that the segment has a low rate of
303 Quaternary slip (Wang *et al*. 2014).

304 The Indaw and Mawlu segments make up a complex zone of faulting that terminates in the south
305 close to the tips of the Tawma and Sagaing segments. This zone forms the western margin of the
306 Katha-Gangaw belt, a >1300 m high curvilinear range of mountains composed of low-medium
307 pressure – high temperature gneisses which pass east into lower grade metamorphic rocks of the
308 Tagaung-Myitkyina belt and the Mogok Metamorphic Belt on the east side of the Bhamo Basin
309 (Mitchell *et al*. 2007; Searle *et al*. 2007). Indaw lake lies at a right step between the Indaw segment
310 and Mawlu segment (Fig. 8b). The rhomboidal step-over is 10 km long and almost 3 km wide, a 3:1
311 length:width ratio characteristic of many pull-apart basins. Satellite imagery reveals evidence of
312 recent normal faulting in the alluvial and lacustrine sediments of the basin, including sets of inward-
313 dipping NE-SW trending normal faults linking sidewall fault systems. Although the lake itself forms
314 the locus of local subsidence at the step-over, the modern depocentre and axial drainage lies near the
315 western edge of the 80 km long basin west of the fault segments. Nonetheless, up to 4 km of dextral
316 geomorphological offset (Wang *et al*. 2014) and ~18 mm/yr of dextral slip measured using GPS
317 across the Kaimang segment along strike to the north (Maurin *et al*. 2010) indicate that these
318 segments comprise perhaps the most active of the Sagaing Fault's northern splays.

319 The horsetail splay proper originates from the northern tip of the Mawlu segment, where Indawgyi
320 lake lies at the intersection of the Mawlu, Shaduzup and Kamaing segments, as well as a number of
321 other minor branching structures (Fig. 8c). Unlike the Shaduzup segment, the Kamaing segment,
322 which includes strands locally known as the Takri and Koma faults, shows geomorphic evidence of
323 recent activity and continues around the Hukawng basin towards the Assam Valley in India (Wang *et*
324 *al*. 2014). This is also the segment across which Maurin *et al*. (2010) measured all instantaneous
325 strain, with other strands of the splay inferred to be presently inactive. Like Indaw Lake, the larger
326 Indawgyi Lake has been interpreted as forming at a releasing step-over (e.g. Soe Thura Tun 1999; Hla
327 Hla Aung 2011), with normal faults concentrated in its NW and SE quadrants, though its position at
328 the intersection of several fault segments makes its structural architecture unclear. It may be related to
329 the right-step and linkage between the northern Ban Mauk horsetail splay and the southern Shaduzup
330 segment tip, or a narrower step-over between the Mawlu and Kamaing/Shaduzup segments. A third
331 option involves strands of the Mawlu segment passing below alluvial fans south of the lake, and
332 linking directly with the Kamaing segment. This would lend the lake a releasing bend geometry. Like
333 Indaw, Indawgyi has a length:width ratio of 3:1, and like the extremely shallow Inle Lake on the Shan
334 Plateau, Indawgyi's 22 m depth is anomalously shallow for a pull-apart lake – reflecting either slow
335 tectonics or rapid sedimentation. Oral reports suggest that Indawgyi lake formed in historic times:
336 several thousand *Taman* (an early tribe living in the upper Chindwin region) reportedly died when the
337 previously dry and fertile Indawgyi valley was catastrophically flooded (Grant Brown 1911). No date
338 for this event or record of an earthquake is reported, and it remains unclear whether the flooding of
339 Indawgyi was the result of massive coseismic subsidence or a landslide.

340 Forming the easternmost segment of the Sagaing Fault horsetail splay, the Mogang segment forms the
341 northern margin of the Bhamo basin, and links to the Sagaing-Namyin Fault before terminating in a
342 north-directed thrust near Putao. The Mogang segment forms the western margin of the narrow belt of
343 structures and exhumed metamorphic complexes of northernmost Sundaland in Yunnan, which
344 include the Gaoligong Shan and Chong Shan shear zones, the Diangcan Shan and Ailao Shan-Red
345 River Fault (e.g. Socquet & Pubellier 2005; Searle & Morley 2011). Despite evidence of a 10 km river
346 offset, the Mogang segment is presently seismically quiet (Wang *et al*. 2014) and does not show a
347 significant geodetic slip rate (Maurin *et al*. 2010).

348   *Sagaing segment*

349   The Sagaing segment, immediately south of the termination splay (Fig. 2a), is interpreted by Wang *et*
350   *al.* (2014) to originate close to the convergence of the northern termination splay at a left-step-over at
351   23.5°N, and to terminate south of the uplifted pressure ridges close to Sagaing city. We prefer to
352   extend the segment north to about 23.9°N on the basis of 1) scant evidence for the uplift that might be
353   expected at a major left-step; 2) the apparent continuity of the fault trace along the west bank of the
354   Ayeyarwaddy River north of Male; 3) the marked discontinuity of active segments on the east bank
355   north of Sabenago; and 4) a series of left-stepping *en echelon* faults at the tip of the segment north of
356   23.9°N (SW margin of the Indaw basin). This brings the length of the Sagaing segment to >220 km.

357   For much of the northern half of the Sagaing segment the Ayeyarwaddy River follows the Sagaing
358   Fault's N-S trace, which also forms the margin of the most westerly extension of the Shan Plateau in
359   the region of soft linkage between the Momeik, Kyaukkyan, Shan Scarp and Sagaing faults. Where
360   the Sagaing Fault forms the plateau margin, a series of small pressure ridges composed of intensely
361   deformed and locally mylonitic paragneisses, mafic rocks and marble, as well as slices of strongly
362   folded Upper Pegu and Irrawaddy clastics mark the fault trace. The ridges are bounded by two sub-
363   parallel fault strands less than 0.5 km apart, with the freshest geomorphic expression on the eastern
364   strand, part of which failed in 2012. A series of narrow sag ponds along the eastern strand reflect
365   shallow segmentation of the Sagaing Fault as it reaches the surface, and include one 40 m wide
366   incipient pull-apart near Thabeikkyin associated with >0.4 m of coseismic subsidence (Fig. 8d). Like
367   Indaw and Indawgyi lakes. the incipient pull-apart lies at a right step-over, has normal faults in its NW
368   and SE quadrants, and an element of dip-slip along the dominantly strike-slip sidewall faults. Unlike
369   the larger basins, the incipient pull-apart represents geologically infinitesimal strain, thus the faults are
370   represented by minor fractures such as R and R' shears, linking P shears and mode I fractures. At
371   higher geologic strain such features become rotated and overprinted by through-going faults, and
372   should not be interpreted on the scale of through-going fault strands (e.g. Christie-Blick & Biddle
373   1985). The fault strands pass north into the Ayeyarwaddy River close to Male, before continuing along
374   the western bank of the river adjacent to a wide region of uplift. Structural control of recent sand bars
375   in the Ayeyarwaddy River between Sabenago and Male (Fig. 5c), supports recent vertical motion
376   along the fault. Wang *et al.* (2014) identify an east facing scarp on a fluvial surface on a less
377   prominent eastern fault trace further north, also indicating recent activity.

378   North of Singu the Sagaing Fault bends and steps to the right and cuts a complex path through the
379   Singu basalt plateau (Fig. 9a), which was emplaced as a 100 m thick series of flows overlying the
380   Mio-Pliocene Irrawaddy Group (e.g. Chhibber 1934; Bender 1983). The potassic alkali trachy-basalts
381   have yielded whole rock ^{40}K-^{40}Ar ages of 0.45 Ma to 0.25 Ma, and have been displaced 2.7-6.5 km by
382   the Sagaing Fault (Bertrand *et al.* 1998). The fault is expressed through much of the Singu basalt
383   plateau as a discontinuous series of *en echelon* shears with an apparent synthetic Riedel geometry,
384   consistent with lower accumulated strain through the late Pleistocene basalts compared to older rocks
385   elsewhere along the fault zone. Near the south end of the plateau, the fault comprises a multi-strand
386   zone almost 3 km wide on the east Ayeyarwaddy River bank at Kule (Fig. 9a), and passes into an
387   apparently simple trace on the west bank of the river south of Singu town. The partly breached Kule
388   releasing step-over results in graben-like subsidence at Kyet Phyu Taung lake, and likely terminated
389   the November 2012 earthquake surface rupture (see below). This step-over, mid-way along the
390   otherwise continuous Sagaing segment, highlights the difficulty of defining meaningful structural
391   segments. While it arrested the 2012 M_w 6.8 rupture, it appears to have been breached by larger
392   magnitude earthquakes in 1839 and 1946 (Hurukawa & Phyo Maung Maung 2011; Wang *et al.* 2014).

393   West of Singu is the southernmost evidence of splaying into the northern horsetail. Two poorly
394   developed splays, including the Male Fault, curve to the NNW from about 22.5°N before rejoining the
395   main horsetail array at 23.8°N. Though there is little geomorphic evidence of recent activity, both
396   splays truncate folded Mio-Pliocene strata and are associated with small sag ponds.

397   A prominent feature of the southern part of the generally linear Sagaing segment is the presence of
398   several N-S trending pressure ridges, notably Sagaing ridge, whose tectonic geomorphology is
399   described above. Sagaing ridge is bounded by a locally segmented strike-slip fault zone in the west
400   and a system of curved reverse and strike-slip faults in the east (Fig. 9b). The ridge is composed of
401   pelitic gneisses overlain by calcsilicate and marble (De Terra 1943; Myint Thein *et al*. 1982),
402   generally of amphibolite to granulite metamorphic facies, and greenschist facies in Minwun ridge in
403   the west (Kan Saw 1973; Me Me Aung 2007; *both in Myint Thein 2012*). Deformed Miocene to
404   Recent siliciclastic sediments including Upper Pegu and Irrawaddy groups unconformably overlay the
405   metamorphics (Myint Thein *et al*. 1982). Folds and thrusts within these sediments trend along 150°
406   (Bertrand & Rangin 2003). Unlike the young uplift associated with restraining bends along the
407   Meiktila and Nay Pyi Taw segments (see below), the Sagaing ridge uplift does not occur at a local
408   restraining bend or step along the Sagaing Fault. There are a number of alternatives to explain its
409   elevation: 1) it may represent uplift at an older restraining feature subsequently breached and now
410   translated along a straight section of the fault; 2) its position at the apex of the long, gentle westward
411   curve of the onshore Sagaing Fault places it close to the entry point of a regional-scale, albeit very
412   gentle restraining bend (<5°); 3) it may be related to transpression as a result of the westward flow of
413   the Shan Plateau south of the Bhamo Basin, again centred on the apex of the broad westward curve; 4)
414   there may be an element of inselberg amplification whereby the ridge is part of a much larger
415   heterogeneous fault-bounded sliver from which non-crystalline rocks are completely planed away
416   leaving an upstanding metamorphic massif.

417   South of the Ayeyarwaddy River the Sagaing Fault is difficult to trace for about 30 km. However,
418   where the fault can be observed, for example at Mandalay airport, it has stepped to the right of its
419   position at Sagaing city, indicating that the low ground south of the city may result in part from a
420   releasing step/bend up to 4 km wide. This apparent discontinuity forms the intersection between the
421   Sagaing and Meiktila segments, as defined by Wang *et al*. (2014).

422   *Meiktila segment*

423   The northern tip of the Meiktila segment lies close to the Ayeyarwaddy River south of Sagaing city,
424   and the southern tip links to the Nay Pyi Taw basin, making a length of 220 km (Wang *et al*. 2014)
425   (Fig. 2a). From the wide Ayeyarwaddy valley in the north, the segment connects, via a system of
426   actively growing anticlines, to the eastern margin of Pegu Yoma. Gentle restraining double bends at
427   21.2°N and 21.5°N are associated with elongate zones of uplift, while other small ridges may be due
428   to shallow shear dilation in the floodplain sediments (Wang *et al*. 2014). A more pronounced change
429   in orientation along the southern half of the segment, from the nearly N-S Sagaing segment trend to
430   the NNW Nay Pyi Taw trend, is associated with a growing fold at 20.4°N (Fig. 4b), discussed above.
431   Here the surface fault trace is highly segmented and sag ponds exist at three right-step-overs no more
432   than 0.5 km wide. The southern limit of the Meiktila segment is defined by a releasing splay at the
433   northern tip of the rhomboidal Nay Pyi Taw basin. However, an apparently continuous strand of the
434   splay continues south along the Meiktila trend, meaning there is no significant structural break
435   between the Meiktila and Nay Pyi Taw segments.

436   *Nay Pyi Taw segment*

437   The Nay Pyi Taw segment comprises a western strand that includes a paired releasing-restraining
438   double bend system, and a linear cut-off fault along the east side of the Nay Pyi Taw basin which
439   bypasses the bends and is parallel to both the Meiktila and Pyu segments (Figs. 2b & 9c). The two
440   strands are ≤11 km apart. Strike-slip strain is distributed across both strands and topographic relief is
441   low (Wang *et al*. 2014). In the north, a releasing double bend forms a 40 km long rhomboidal basin
442   within which lies the newly built capital city Nay Pyi Taw. In the south, a restraining double bend
443   rejoins the main Sagaing Fault trace via a system of splays around the latitude of Taungoo. The
444   restraining double bend is associated with *en echelon* folds within Mio-Pliocene strata of the Upper
445   Pegu and Irrawaddy groups, several of which are truncated by strike-slip fault strands. Alluvial fan
446   uplift east of the southern portion of the cut-off fault is associated with west-dipping reverse faults
447   which continue south to characterise the Pyu segment (Replumaz 1999). Although the Nay Pyi Taw
448   segment is structurally more complex than other central sections of the Sagaing Fault, the apparent
449   continuity of the eastern cut-off fault means that the major releasing and restraining bends need not
450   provide a barrier to rupture propagation along and beyond the segment. From the northern tip of the
451   rhomboidal Nay Pyi Taw basin to the southernmost termination of the restraining bend, the segment is
452   almost 130 km long.

453   *Pyu Segment*

454   The entire, highly linear Pyu segment south of the Nay Pyi Taw restraining bend is orientated about
455   10° anticlockwise from the overall strike of the Sagaing Fault (Wang *et al*. 2014), resulting in
456   characteristic transpression throughout the segment. A prominent restraining step-over at 18°N (Fig
457   9d) produces an elongate zone of uplift immediately east of the main Pegu Yoma ranges, and marks
458   the southern limit of the Pyu segment. Background seismicity at the step-over is high, and it has been
459   interpreted as a low-coupled patch which terminates large seismic ruptures (Wang *et al*. 2011),
460   although the step-over itself is less that 1 km wide, smaller than the minimum empirically determined
461   likely to terminate rupture (e.g. Wesnousky 2006). The northern boundary of the Pyu segment is
462   marked by splays curving west into the Nay Pyi Taw paired bend system. Wang *et al*. (2014) identify
463   splays at 19.1°N near where the 1930 M_w 7.3 Pyu earthquake rupture was arrested, but we prefer to
464   terminate the segment at the most southerly set of splays (Fig. 2a), making the segment only 85 km
465   long and accepting that the 1930 earthquake breached the segment boundary. Strike-slip along the Pyu
466   segment is partitioned onto a highly linear fault at the foot of an east-facing scarp rising >400 m
467   above the Sittaung River basin. An elevated and deeply incised elongate sliver of the basin up to 7 km
468   wide adjacent to the fault may be bounded in the east by west-dipping reverse faults (Replumaz
469   1999).

470   *Bago segment*

471   The Bago segment passes 170 km south from the restraining bend at 18°N to the Mottama Gulf coast
472   (Wang *et al*. 2014) (Fig. 2a). Along large portions of the segment the fault trace is largely obscured
473   because of agricultural activity, recent sedimentation by the Sittaung and Bago rivers and negligible
474   topographic relief. However, geomorphic evidence of earthquakes, including those in 1930 and older,
475   such as fault scarps, subsidence, uplift, offset of streams and field boundaries is well preserved along
476   the segment particularly from Bago south almost to the coast (Tsutsumi & Sato 2009; Wang *et al*
477   2011). The Pale Fault, a 17 km long fault strand west of and parallel to the main Sagaing Fault, also
478   shows evidence of surface rupture during the 1930 earthquake (Tsutsumi & Sato 2009) and marks a
479   very gentle net right bend in the fault trace. The southernmost 60 km of the onshore fault is
480   remarkably straight and continuous, and trends exactly N-S resulting in no vertical motion. The

481 southern termination of the Bago segment has no structural significance other than being the point
482 where the fault goes offshore, and as such would not be a barrier to rupture propagation.

Southern segments and termination

484 As it passes offshore into the Gulf of Mottama south of the Bago segment, the Sagaing Fault splays
485 into three segments, two of which terminate in a broad zone of ENE-WSW trending normal faults
486 (Fig. 1). The splays are associated with zones of intense deformation, disruption of the sea floor, and
487 mobilisation of overpressured fluids (Morley 2013), the latter also observed locally onshore, for
488 example at Nay Pyi Taw (Sloan *et al. this volume*), following coseismic liquefaction at Bago
489 (Chhibber 1934) and near Singu (see below). The easternmost splay continues south of the Gulf of
490 Mottama extensional domain to transfer into the Pliocene-Recent Andaman Sea spreading centre and
491 associated transform faults via a series of ENE-WSW trending normal faults (Curray 2005; Morley *et*
492 *al*. 2011; Morley 2013). Along strike from the southern tip of the eastern splay, a NNE-trending linear
493 feature interpreted as the ocean-continent boundary offshore the Mergui Archipelago (Curray 2005)
494 may instead be an inactive strike-slip precursor to the modern Sagaing Fault within thinned
495 continental crust, and has been termed the 'South Sagaing' Fault segment by Morley (2013).

Displacement

Onset of dextral motion

498 Models for Cenozoic lateral extrusion in response to Indian indentation into Eurasia typically call for
499 substantial dextral shear in the position of the Sagaing Fault and Gaoligong Shan along the western
500 margin of a rigid Indochina block, while the Chong Shan and Ailao Shan-Red River Fault in Yunnan
501 form the sinistral eastern margin of the extruding block (e.g. Tapponnier *et al*. 1982, 1986; Briais *et*
502 *al*. 1993; Socquet & Pubellier 2005). However, despite Oligocene and older peak metamorphism and
503 partial melting in Xuelong Shan, Diancang Shan and Ailao Shan metamorphic complexes (e.g.
504 Harrison *et al*. 1992; Zhang & Scharer 1999; Leloup *et al*. 2001), retrograde sinistral shear and
505 exhumation along a discrete Ailao Shan-Red River Fault did not occur until the Miocene (Searle
506 2006). Similarly, there is little evidence for a discrete pre-Mid-Miocene Sagaing Fault, rather a long-
507 lived zone of diffuse dextral shear in Thailand and east/central Myanmar that may have been a lateral
508 extrusion-driven earliest precursor to the localised Recent strike-slip structure (e.g. Le Dain *et al*.
509 1984; Bertrand & Rangin 2003; Socquet & Pubellier 2005; Searle & Morley 2011).

510 At the extreme southern end of the Sagaing Fault, decoupling of West Burma from Sundaland by a
511 zone of diffuse dextral transtension may have its first expression in WNW-ESE directed extension in
512 the Mergui and North Sumatra Basins, which began developing during the Late Eocene to Early
513 Oligocene (Srisuriyon & Morley 2014). Contemporaneous (37-29 Ma) peak high temperature
514 metamorphism in the Mogok Metamorphic Belt close to Mandalay in the north may be associated
515 with NW-SE sinistral transpression along the Mae Ping and Three Pagodas Faults in Thailand (Morley
516 2004; Searle *et al*. 2007; Searle & Morley 2011). These events also shortly follow the 48-40 Ma final
517 phases of diffuse dextral shear and magmatism along the NNE-trending Ranong and Khlong Marui
518 faults on the eastern side of the Mergui Basin in Thailand (Watkinson *et al*. 2011), suggesting a
519 progressive westwards localisation of dextral strain.

520 The NNE-trending 'South Sagaing' Fault segment, along strike from the southern tip of the Sagaing
521 Fault, is parallel to the eastern boundary of the Sewell Rise in the Andaman Sea and bounds a linear
522 Late Oligocene-Early Miocene basin (Morley 2013). If the bounding fault is indeed dextral, this

523   segment could represent the site of the earliest localised strike-slip in the position of the Sagaing
524   Fault.

525   Synchronous with postulated activity along the 'South Sagaing' Fault segment, a diachronous belt of
526   Oligocene to Early Miocene ductile dextral shear overprinted Mogok Metamorphic Belt gneisses and
527   Gaoligong Shan metamorphics in Yunnan (Bertrand & Rangin 2003; Socquet & Pubellier, 2005;
528   Morley 2013), recorded by ductile stretching lineations and northward-younging Ar-Ar cooling ages
529   of 26.9 Ma to 15.8 Ma (Bertrand et al. 2001). Ductile extensional fabrics of the Mogok Metamorphic
530   Belt point to a period of post-peak metamorphism transtension which may have included dextral pull-
531   apart basins rooted in the now-exhumed ductile basement (Bertrand & Rangin 2003; Searle et al.
532   2007) and extension in the Central Basin (e.g. Bertrand & Rangin 2003). Younger cooling ages of 20-
533   10 Ma further north in the Gaoligong Shan and Chongshan (Wang & Burchfiel 1997; Akciz et al.
534   2008; Zhang et al. 2012) support uplift by a wave of northward migrating dextral transpression that
535   may have shortly pre-dated Mid to Late Miocene localisation of strain along onshore brittle dextral
536   structures including the Sagaing Fault (Socquet & Pubellier 2005).

537   Ductile fabrics in parts of the central Mogok Metamorphic Belt are sealed by unfoliated Oligo-
538   Miocene syenites and leucogranites (Barley et al. 2003; Mitchell et al. 2007; Searle et al. 2007).
539   Typical of these intrusions, the Sedo syenogranite, which cross-cuts earlier ductile fabrics, yields a
540   SHRIMP age of 22.6 ± 0.4Ma (Barley et al. 2003; Searle et al. 2007). Deformation of such Oligo-
541   Miocene intrusions by the brittle Sagaing Fault close to Mandalay supports onset of localised dextral
542   motion along the central part of the fault after 22 Ma (Socquet & Pubellier 2005; Searle et al. 2007)
543   and possibly as late as 15 Ma when diffuse dextral strike-slip ceased further east in Thailand (Morley
544   2004; Searle & Morley 2011).

545   In the Andaman Sea the deep Central Andaman Basin spreading centre separates older, rugged
546   bathymetric plateaux of poorly known genesis (Alcock and Sewell Rises, Fig. 1) and forms part of a
547   ~300 km wide right step-over system between the Sagaing Fault and the Sumatran Fault, inboard of
548   the Sunda Trench (e.g. Curray et al. 1979; Chamot-Rooke et al. 2001; Kamesh Raju et al. 2004;
549   Curray 2005; Morley 2013; Srisuriyon & Morley 2014). The Central Andaman Basin lies south of the
550   southern Sagaing Fault termination splay and has a soft linkage with the easternmost strand via a
551   series of ENE-WSW trending normal faults (Morley 2013). The western margin of the spreading
552   centre is similarly marked by dextral faults including the West Andaman Fault zone, which transfer
553   displacement south onto the Sumatran Fault (Kamesh Raju et al. 2004; Curray 2005).

554   The spreading centre is inferred to accommodate all residual dextral strain at the tip of the Sagaing
555   Fault, in the same way that the South China Sea has more controversially been inferred to terminate
556   Oligo-Miocene Ailao Shan-Red River Fault sinistral slip (e.g. Tapponnier et al. 1982; Briais et al.
557   1993; Wang et al. 2000; Hall 2002; Morley 2002; Searle 2006; Leloup et al. 2007; Morley 2013).
558   Based on poorly developed magnetic anomalies of presumed oceanic crust in the Central Andaman
559   Basin, seafloor spreading is widely considered to have occurred from the Pliocene (4 Ma) to Recent
560   (e.g. Kaesh Raju et al. 2004; Curray 2005). It follows that since 4 Ma, 118 km of spreading between
561   the Alcock and Sewell rises must have occurred at an average rate of 30 mm/yr, and must have been
562   accommodated by similar deformation along the Sagaing Fault (Kamesh Raju et al. 2004; Curray
563   2005). However, thick sedimentary cover within the deep axial trough of the Central Andaman Basin
564   where the youngest oceanic crust should be led Morley & Alvey (2015) to propose that oceanic
565   spreading occurred during the Late Miocene- Early Pliocene instead, and that the magnetic anomalies
566   are of insufficient quality to confirm a younger age. This means that a through-going Sagaing Fault

567 may have been developed from the Late Miocene, and that the average Pliocene-Recent slip rate of 30
568 mm/yr may be excessive.

569 In the Myanmar Central Basin positive inversion is recorded by terrestrial deposits of the Late
570 Miocene-Pliocene syn-inversion Irrawaddy Group (Bender 1983). Although the Irrawaddy Group has
571 a diachronous base and as presently understood is a poor marker for inversion onset (e.g. Sloan *et al*.
572 *this volume*), apatite fission track data support a similar Late Miocene uplift onset and a Plio-
573 Pleistocene uplift peak (Trevena *et al*. 1991). Onset of inversion in the basin has been considered to
574 be related to development of a through-going Sagaing Fault; for example by Himalayan buttressing,
575 permitted by northward motion of western Myanmar along the fault (Pivnik *et al*. 1998); or by a
576 regional kinematic reorganisation refocusing extension from the Central Basin onto the Andaman Sea,
577 as a result of Sagaing Fault connection and compounded by west-directed flow during Tibet Plateau
578 gravitational collapse (e.g. Bertrand & Rangin 2003; Rangin *et al*. 2013). Both models support
579 localisation of a through-going Sagaing Fault by the latest Miocene or start of the Pliocene (e.g.
580 Bertrand & Rangin 2003).

581 *Finite displacement and long-term slip rate*

582 Total Neogene dextral displacement across the Sagaing Fault remains poorly known, mainly due to
583 the paucity of pre-kinematic piercing points and uncertainty regarding linked tectonic structures such
584 as Andaman Sea basins (e.g. Curray 2005; Morley 2013; Morley & Alvey 2015). Upper estimates
585 including 700 km (Replumaz & Tapponnier 2003), 460 km (Curray *et al*. 1982) and 332 km (Curray
586 2005) are based on geodynamic reconstructions which assume rigid blocks and accommodation of
587 early phases of Andaman Sea oceanic spreading by the Sagaing Fault. Using a different approach,
588 Bertrand and Rangin (2003) assumed that the dextral Sagaing Fault initiated at 5 Ma, and extrapolated
589 a present-day geodetic slip rate of 20 mm/yr (Vigny *et al*. 2003) to propose a total dextral
590 displacement of 100 km. However, this estimate omits the poorly known pre-Pliocene displacement
591 reviewed above.

592 The apparent dextral offset between the Ayeyarwaddy River, where it crosses the Sagaing Fault near
593 Tagaung, and the upper Chindwin River, a proposed beheaded channel, is 425-460 km (Hla Maung
594 1987). A more conservative value using the same principle but incorporating the piercing points
595 shown in Figs. 10a and 10b is 340 km. All these values are problematic as displacement estimates
596 because the beheaded modern Chindwin River lacks the underfit valley that should be inherited from
597 a previously greater catchment (Myint Thein *et al*. 1991). It is also structurally difficult to accept such
598 large displacements near the tip of a northern strand of the Sagaing Fault which passes directly north
599 into the Chindwin basin and apparently terminates, especially when most strike-slip strain appears to
600 be focused on segments around the east of the Hukawng basin.

601 Correlation of pre-kinematic geological piercing points has proven problematic for estimating lateral
602 displacement along the Sagaing Fault. Albian limestones and serpentinites in the Tagaung-Myitkyina
603 belt, or schists and ophiolitic rocks of the Katha Gangaw belt within and to the east of the Sagaing
604 Fault northern splay, may correlate to the Mount Victoria-Kawlun belt of the Western Myanmar arc
605 system to the west, implying a post-Eocene dextral offset of >300 km (Mitchell 1977, 1981, 1993).
606 However, Myint Thein *et al*. (1991) point out that such a correlation relies on a similar geologic origin
607 and setting for the apparently offset belts, something that is yet to be proven. A lower total dextral
608 displacement estimate is based on the 203 km lateral separation of the Mayathein metamorphics – a
609 sliver of schists and gneisses on the west side of the Sagaing Fault at Tigyaing – and the remarkably
610 similar metamorphic suite exposed east of the fault at Sagaing Ridge 203 km to the south (Myint

611   Thein *et al*. 1991) (Fig. 10c). While the correlation is appealing, there are problems with this model.
612   The modern fault strand responsible for the offset likely bisects a basement sliver already uplifted by
613   transpression, so early lateral offset is excluded. It also seems unlikely that the two metamorphic areas
614   would have experienced such negligible differential vertical motion during the subsequent 203 km
615   lateral translation that they remain exposed at the same structural level.

616   Offsets of syn-tectonic piercing points have been used to calculate post-Miocene displacement and
617   slip rate along the Sagaing Fault. At Singu, syn-tectonic Pleistocene basalts possibly vented through
618   the active fault strands form a low elliptical plateau which overlies the upper part of the Plio-
619   Quaternary Irrawaddy Group and straddles the fault trace (Chhibber 1934; Bender 1983; Bertrand *et*
620   *al*. 1998). Apparent dextral offset of the south and north margins of the plateau across the fault by 2.7
621   km and 6.5 km respectively (Fig. 11 a-c) has been considered to result from Quaternary Sagaing Fault
622   displacement (Bertrand *et al*. 1998). Five whole rock ^{40}K-^{40}Ar ages from the basalts range from 0.25 ±
623   0.2 to 0.31 ± 0.02 Ma, yielding Quaternary slip rates of 10 ± 1 mm/yr to 23 ± 3 mm/yr (Bertrand *et al*.
624   1998). Although these rates are in accordance with geodetic results (discussed below), there is
625   considerable uncertainty in the position of the original margin of the plateau, and hence the amount of
626   offset. Further south, Pontian (lowermost Irrawaddy Fm.) alluvial fans containing muscovite schist,
627   ?Triassic carbonate clasts and Late Miocene Upper Pegu Group clastics at Mezaligyaung, near the
628   northern end of the Sagaing Ridge, are offset along the Sagaing Fault by about 20 km from their
629   nearest uplifted source (Myint Thein *et al*. 1991). Such a small offset could imply a very slow post-
630   Miocene slip rate of only 4 mm/yr, but it is not clear on what basis the fans have been dated, or
631   whether offset occurred immediately after deposition.

632   Aside from the very large and speculative Ayeyarwaddy/Chindwin river offset (Hla Maung 1987),
633   numerous rivers preserve variable offset where they cross the Sagaing Fault, particularly along the
634   eastern flank of Pegu Yoma (e.g. Dey 1968; Win Swe 1970; Myint Thein *et al*. 1987) where right-
635   lateral offsets of 3.7 km (Myint Thein *et al*. 1991) and 2.4 km (Wang *et al*. 2014) have been reported.
636   South of Thabeikkyin a stream is offset 1.0 km along the fault strand which re-ruptured during the
637   2012 earthquake. The stream now occupies a valley previously cut by a larger stream which has been
638   offset 1.6 km to the right and subsequently captured by headward erosion of a minor stream, lending
639   the larger stream a 0.2 km left-lateral offset where it empties into the Ayeyarwaddy River at Lower
640   Ponna. That stream now crosses an oversized delta which was formed by a much larger river, since
641   offset and captured 1.8 km to the south (Fig. 11d). Similar stream offset, up to 3 km, has been
642   reported along strike north of Thabeikkyin (Win Swe 1970).

643   There are few measurements of coseismic displacement across the Sagaing Fault. The Payagyi
644   fortress, north of Bago, was built across a strand of the fault during the 16th-century, and shows a total
645   of 5-7.5 m dextral offset of its outer wall accumulated during at least two major fault ruptures, and a
646   small contribution from the May 1930 Bago earthquake (Wang *et al*. 2011). Historical, radiometric
647   and stratigraphic constraints place construction of the fortress between 1539 and 1599, yielding a
648   range of historic slip rates from 11-18 mm/yr (Wang *et al*. 2011). Maximum coseismic dextral
649   displacement during the May 1930 earthquake is recorded south of Bago by an irrigation canal offset
650   3.2 ± 0.3 m and a stream channel offset 3.3 ± 0.2 m (Tsutsumi & Sato 2009). We measured a
651   maximum of 1.02 m dextral displacement of a footpath following the November 2012 M_w 6.8
652   Thabeikkyin earthquake (described below). Sixty-six years earlier, the same fault section failed during
653   the 1946 M_w 7.7 near complete rupture of the Sagaing segment (Engdahl & Villaseñor 2002; Wang *et*
654   *al*. 2014), which originated at a latitude of 22.35 ± 0.25°N, and may have propagated 185 km
655   northwards through Thabeikkyin and Tagaung (Hurukawa & Phyo Maung Maung 2011). Two small
656   stream offsets of 1.7-2.0 m measured south of Thabeikkyin in 2013 (described above, Fig. 4d) were

657 cut by the 2012 surface rupture, which locally contributed 0.78 m offset. The remaining ~0.9-1.2 m
658 dextral offset may be inherited 1946 coseismic displacement.

659 Assuming the 1946 event completely relaxed the segment that failed again in 2012 after a maximum
660 of 1.02 m of strain accumulation, and assuming recovery of all elastic strain during the 2012 event,
661 we calculate a 66-year average slip rate of 15.5 mm/yr, slightly lower than geodetic rates reviewed
662 below (e.g. Socquet *et al*. 2006) but consistent with the longer-term slip rates reviewed previously
663 (e.g. Bertrand *et al*. 1998; Wang *et al*. 2011). It is likely that field measurements of offset cultural
664 markers such as walls and paths significantly underestimate far-field displacement, so the calculated
665 historic slip rates should be viewed as minima.

666 *Instantaneous slip rate*

667 Models based on regional Global Positioning System (GPS) campaigns suggest that at the latitude of
668 Myanmar there is presently 35-36 mm/yr of motion between stable India and Sundaland along an
669 azimuth of 011°-014° (Socquet *et al*. 2006). Although the Sagaing Fault has been considered to be a
670 plate-bounding structure within this margin (e.g. Le Dain *et al*. 1984; Guzmàn-Speziale & Ni 1996),
671 significant off-fault strain is distributed throughout a >400 km wide shear zone straddling much of
672 Myanmar (e.g. Hla Maung 1987; Holt *et al*. 1991; Bertrand *et al*. 1998). Residual India-Sundaland
673 motion is accommodated on strike-slip structures within the Indo-Myanmar Ranges such as the
674 Thahtay Fault and Kabaw Fault, on *en echelon* folds within the Central Basin, and by highly oblique
675 slip within or elastic loading of the Andaman Trench (Le Dain *et al*. 1984; Hla Maung 1987; Pivnik *et*
676 *al*. 1998; Nielsen *et al*. 2004; Socquet *et al*. 2006; Wang *et al*. 2014) (Fig. 12).

677 Eighteen GPS stations make up three transects across the central Sagaing Fault – one 140 km long at
678 the latitude of Mandalay, and two, each 70 km long, 35 km north and south of the main transect
679 (Vigny *et al*. 2003). The GPS transects were established in the late 1990s in addition to the four
680 regional Myanmar GPS stations of the Geodynamics of South and SE Asia project (GEODYSSEA)
681 (Vigny *et al*. 2003). On the east (Sundaland) side of the Sagaing Fault, Ywenge station moves 6
682 mm/yr WSW relative to a stable Sundaland reference frame represented by Hpa-An station (Socquet
683 *et al*. 2006). On the western side of the fault, Kwehtaing Taung station moves 17 mm/yr to the NNW,
684 equating to a total of 18-20mm/yr of fault-parallel dextral slip across the Sagaing Fault (Vigny *et al*
685 2003; Socquet *et al*. 2006). Similar velocities of 18 mm/yr were recorded from a GPS transect which
686 straddled several strands of the northern Sagaing Fault north of Indawgyi lake (Maurin *et al*. 2010),
687 although all statistically meaningful motion was measured across the single Kamaing segment
688 (reported as the Koma Fault).

689 Supporting GPS and historic slip rates, seismic moment tensor analysis of hypocentral data from
690 shallow earthquakes along the Sagaing Fault indicates a right-lateral slip rate of 16.4 mm/yr north of
691 Mandalay, and a rate of 11 mm/yr in the south (Radha Krishna & Sanu 2000). It should be noted that
692 these values are influenced by the rather small quantity of data for the southern area, and the statistical
693 impact of including relatively few of the largest historical earthquakes, which are widely separated in
694 time but which dominate strain release. Despite these uncertainties, generally consistent geodetic
695 results suggest that the Sagaing Fault slips rather uniformly along its length and is the most significant
696 single structure of the plate boundary, accommodating about half of India-Sundaland right-lateral
697 motion.

698 Sagaing Fault-parallel velocities from GPS stations of the Mandalay transects define an arctangent
699 curve when plotted against distance, which, with a far-field velocity of 18 mm/yr correspond to an
700 average 15 km locking depth (Vigny *et al*. 2003), also modelled as 20 km (Maurin *et al*. 2010). The

701 position of maximum modelled shear stress across the Mandalay transect is 17 km east of the Sagaing
702 Fault trace, suggesting either the unlikely scenario of a 40° easterly dip, or a significant amount of
703 dextral strain being accommodated along the Shan Scarp fault system (Vigny *et al*. 2003), for which
704 there is little evidence of neotectonic activity. Modelled locking depths of the active fault in northern
705 Myanmar range from 6-8 km, perhaps a result of a thin elastic thickness and high heat flow
706 approaching SE Tibet (Maurin *et al*. 2010) (Fig. 12).

Seismicity

708 Myanmar is characterised by extensive seismicity with a wide range of focal depths and causative
709 mechanisms (e.g. Le Dain *et al*. 1984; Guzmán-Speziale & Ni 1993; Socquet & Pubellier 2005;
710 Kundu & Gahalaut 2012; Win Swe 2012; Sloan *et al*. *this volume*). In the west, seismicity is
711 associated with oblique shortening across the Indo-Myanmar Ranges, thrusting within the Central
712 Basin, and an east-dipping Wadati-Benioff zone with hypocentrers down to 150 km depth (e.g. Kundu
713 & Gahalaut 2012). In the east, flow around the eastern Himalayan syntaxis results in a complex
714 pattern of shallow strike-slip and extensional earthquakes (e.g. Socquet & Pubellier 2005). Against
715 this background, the Sagaing Fault is expressed by a narrow belt of shallow seismicity (<50 km),
716 dominantly strike-slip focal mechanisms, consistently steep/vertical fault-parallel nodal planes and N-
717 S trending right-lateral slip (e.g. Hurukawa & Phyo Maung Maung 2011; Kundu & Gahalaut 2012)
718 (Fig. 2c). The horizontal velocity gradient determined by GPS across the central part of the fault is
719 typical of a locked structure where deformation is accumulating elastically within the few tens of
720 kilometres either side of the active segment (Vigny *et al*. 2003). Shallow modelled locking depths
721 along the fault, as little as 6-8 km in the north, led Maurin *et al*. (2010) to suggest that some stress
722 release may be accommodated by microseismicity and aseismic creep. However, for most segments of
723 the Sagaing Fault, abundant historic records indicate that energy release by medium and large
724 earthquakes is usual.

Historic seismicity

726 Several hundred years of records document earthquake-induced damage to religious monuments built
727 along the Sagaing Fault. Such records are naturally subjective and biased towards events affecting
728 ancient cities, particularly in the Sagaing area, and neglect events along (previously) remote sections
729 of the fault in the pre-instrumental era. Collapse of pagodas in the Sagaing-Mandalay area due to
730 earthquakes is noted in the years 1429, 1469, 1485, 1588 and 1590 (Win Swe 2013). The first record
731 of a surface rupture comes from 8th June 1619 and again in 1688, on both occasions Sagaing pagodas
732 were also damaged. An earthquake in 1768 was linked to damage as far away as Bagan, while
733 earthquakes at Inn-wa in 1771, 1776 and 1830 caused local damage (Win Swe 2013). More detailed
734 observations were documented following the 1839 Amarapura earthquake (e.g. Chhibber 1934) and
735 the great 1912 Pyin-Oo-Lwin (Maymyo) earthquake, which was probably centred on the Kyaukkyan
736 Fault to the east (e.g. Coggin Brown 1917), but no detailed geological description of a Sagaing Fault
737 earthquake was made until the 2012 Thabeikkyin event (Watkinson *et al*. *Submitted*). Nonetheless,
738 recent work suggests that much of the Sagaing Fault has ruptured during the last 200 years, including
739 about half its length during the past 90 years (e.g. Hurukawa & Phyo Maung Maung 2011; Wang *et al*.
740 2014). Here we review historic seismicity of the last 200 years, which broadly follows a geographic
741 rupture pattern starting at the central Sagaing Fault in 1839, progressing north from 1907 to 1908,
742 before earthquake sequences in the 1930's ruptured both the northern and southern parts of the fault.
743 We conclude with observations from the most recent significant event, the 2012 Thabeikkyin
744 earthquake.

745   *Historic seismicity - central Sagaing Fault*

746   On 23rd March 1839 an earthquake apparently centred at Amarapura, 10 km south of Mandalay, killed
747   several hundred people, damaged every brick building in Yadanapura and Ava, formed a surface
748   rupture and caused still-visible damage to the solid brick Mingun Pagoda (Chhibber 1934; Win Swe
749   2011, 2013) (Fig. 13a). Shaking was reported in Bhamo, Yangon and 620 km away in Mawlamyine
750   (Win Swe 2013). It is possible that part or all of the combined 400 km long Meiktila and Sagaing
751   segments of the Sagaing Fault ruptured (Wang *et al*. 2014) (Fig. 2c). A subsequent earthquake of M_w
752   7.1 (Engdahl & Villaseñor 2002) caused further damage at Sagaing and Mingun on 16th July 1956
753   (Win Swe 2013). This event, which may have re-ruptured a ~60 km long segment of the Sagaing Fault
754   immediately south of a 1946 M_w 7.7 rupture (Engdahl & Villaseñor 2002; Hurukawa & Phyo Maung
755   Maung 2011), was the last M>7.0 earthquake along the Sagaing Fault, and the last time the fault
756   ruptured adjacent to Sagaing and Mandalay cities. There have been no further ruptures of any part of
757   the Meiktila segment since the speculated 1839 rupture (Wang *et al*. 2014), and the Meiktila segment
758   now comprises much of a 260 km long seismic gap from just south of Mandalay to just south of Nay
759   Pyi Taw (Hurukawa & Phyo Maung Maung 2011).

760   *Historic seismicity - northern Sagaing Fault*

761   Further north, an M_w 7.6 earthquake (Engdahl & Villaseñor 2002) on 27th January 1931 in Kachin
762   State, caused damage at Karming and may have been caused by a northern strand of the Sagaing
763   Fault, perhaps the Kamaing segment. Hurukawa and Phyo Maung Maung (2011) estimated the
764   earthquake ruptured ~180 km of the fault. Earlier earthquakes of M_w 7.0 in 1906 and M_w 7.5 in 1908
765   in the same area support a stick-slip mechanism for the northern Sagaing Fault (Maurin *et al*. 2010). A
766   further M7.0 event on 18th August 1950 is reported by Win Swe (2013) in the same area, but is
767   reviewed neither by Engdahl and Villaseñor (2002) nor by Hurukawa and Phyo Maung Maung (2011).
768   The 1946 Wuntho earthquake sequence culminated in a M_w 7.7 earthquake (Engdahl & Villaseñor
769   2002) on 12th September, which may have ruptured a 185 km long segment of the Sagaing Fault
770   through Tagaung and Thabeikkyin (Hurukawa & Phyo Maung Maung 2011). Remarkably the 1946
771   M_w 7.7 event occurred three minutes after a M_w 7.3 earthquake (Engdahl & Villaseñor 2002) which
772   ruptured at least 80 km of the Indaw segment to the north (Wang *et al*. 2014), and possibly up to 155
773   km, towards the southern tip of the 1931 Kachin rupture (Hurukawa & Phyo Maung Maung 2011).

774   A 5th January 1991 M_w 7.0 earthquake (Engdahl & Villaseñor 2002) north of Tagaung caused damage
775   as far south as Thabeikkyin, and may have re-ruptured 49 km of the 1946 slip segments, up to the
776   location of a June 1992 M6.3 aftershock near Indaw (Win Swe 2011; Hurukawa & Phyo Maung
777   Maung 2011).

778   *Historic seismicity - southern Sagaing Fault*

779   Historical documentation of southern Sagaing Fault earthquakes includes records of damage to
780   Shwemawdaw Pagoda at Bago (built 307 BC), which was damaged by earthquakes 33 times in the
781   2114 years from 197 BC to 1917 AD (Win Swe, 2011), on average once every 64 years. In 1930 the
782   pagoda collapsed once again and 500 people were killed when the southern end of the Sagaing Fault
783   failed in the first of two closely spaced earthquakes of M_w 7.2 and M_w 7.3 (Engdahl & Villaseñor
784   2002) about 120 km apart at Bago (5th May 1930) and Pyu (3rd December 1930) (Chhibber 1934).
785   Their proximity in space and time indicates that Coulomb stress changes resulting from the Bago
786   event may have triggered the Pyu event (Tsutsumi & Sato 2009), in a sequence remarkably similar to
787   the North Anatolian Fault's M_w 7.4 and M_w 7.1 Izmit and Düzce earthquakes of August and November
788   1999 (e.g. Utkucu *et al*. 2003). The Bago event ruptured between 100 km (Wang *et al*. 2011) and 131

789   km of the Sagaing Fault (Hurukawa & Phyo Maung Maung 2011) and caused ≥3 m of coseismic
790   dextral surface displacement, yielding a recurrence interval of ≥ 160 years (Tsutsumi & Sato 2009).
791   The 16th-century Payagyi fortress lies near the northern end of the Bago rupture, and displays 6 m of
792   dextral displacement (Wang *et al*. 2011). This displacement was accumulated mainly during two
793   strong earthquakes since construction and may have been further slightly offset by the 1930 Bago
794   earthquake, yielding a recurrence interval of one to two centuries, assuming a uniform slip model
795   (Wang *et al*. 2011). The subsequent Pyu event in December 1930 propagated northward from the
796   proposed northern termination of the Bago rupture, and ruptured a further 120 km of the Sagaing
797   Fault (Hurukawa & Phyo Maung Maung 2011) (Fig. 2c).

798   An earlier M~7.0 earthquake on August 8th 1929 bent railway lines at Swa, near Taungoo (Le Dain *et*
799   *al*. 1984; Win Swe 2011). Located at the northern end of the proposed 1930 Pyu rupture on the eastern
800   'shortcut' strand of the Nay Pyi Taw segment, the 1929 event could have contributed to triggering the
801   1930 earthquake series.

802   *2012 Thabeikkyin earthquake*

803   The most recent significant earthquake generated by the Sagaing Fault was the 11th November 2012
804   M_w 6.8 Thabeikkyin earthquake, which ruptured a ~45 km long part of the Sagaing segment between
805   Singu and Sabeanago (Fig. 2c). Centroid moment tensor solutions of the primary event and seven
806   largest aftershocks indicate dextral slip along north-south trending, steeply east dipping to sub-vertical
807   strands of the Sagaing Fault. Aftershocks continued for several months, including clusters of small
808   earthquakes at the northern end of the Sagaing Fault and close to Nay Pyi Taw in the south. Twenty
809   six people were killed and 231 were injured by the mainshock, which occurred at 06.30 local time on
810   a Sunday. Many buildings, including 201 houses, 25 schools, 13 health centres and 35 monasteries
811   were destroyed or severely damaged (Fig. 13b); clearly the fortuitous timing of the earthquake meant
812   many public buildings were unoccupied, significantly reducing casualty numbers (Ko Ko Gyi *et al.*
813   2012). Damage was most severe in towns lining the Ayeyarwaddy River, particularly those on the east
814   bank including Sabeanago, Thabeikkyin, and Kule. Much of the most severe damage was due to
815   gravitational slumping down river banks, particularly in towns beyond the rupture tips such as Male.

816   A well developed 45 km long surface rupture was associated with the 2012 Thabeikkyin earthquake,
817   trending almost exactly N-S (001°) on or close to the line of longitude 95.975° E. For much of its
818   length the surface rupture broadly followed the Ayeyarwaddy River, lying in a valley on the east bank,
819   separated from the river by narrow pressure ridges. The surface rupture was expressed by *en echelon*
820   sigmoidal Riedel shears (Fig. 13c) arranged into small arrays reflecting shallow upwards splaying and
821   curvature of the rupture, and larger arrays that may mimic the segmented geometry of the seismogenic
822   Sagaing Fault at depth. Maximum dextral strike-slip displacement of 1.02 m, recorded north of
823   Thabeikkyin, was accommodated across the shears themselves, across the inter-shear linkages, and
824   across a zone of horizontal flexure at least 5 m wide either side of the surface rupture. A number of
825   small releasing step-overs along the length of the surface rupture were associated with existing or
826   incipient sag ponds, locally associated with up to 1.17 m of coseismic subsidence. Liquefaction along
827   the surface rupture resulted in injection of mud and sand into dilation fractures and localised mud
828   volcanoes, notably near Singu.

829   The surface rupture terminated in the south within and around Kyet Phyu Taung lake, 4 km north of
830   Singu. The lake is part of a broad depression close to where the Ayeyarwaddy River resumes its
831   course along the Sagaing Fault and becomes braided after flowing in a narrow confined channel
832   around the Singu basalt plateau. Multiple sets of mole tracks and transpressional splaying structures

833   associated with 0.5 m of uplift and widely distributed strike-slip indicate that the rupture terminated in
834   a complex horsetail splay, perhaps as it attempted to cross the 1.5 km wide releasing step-over marked
835   by Kyet Phyu Taung lake. In the north the rupture terminated in a smaller array 160 m wide near
836   Kyauk Myae village on the west river bank.

837   **Discussion**

838   *Tectonic evolution*

839   The development of the Sagaing Fault can be summarised broadly as northwards-younging
840   localisation of discrete strike-slip fault segments from a long-lived zone of diffuse dextral shear along
841   the eastern margin of India (Fig. 14). Early dextral transpression along this margin (Fig. 14a) has been
842   driven by the northward motion and indentation of Greater India into Eurasia during the Himalayan
843   Orogeny since the Early Eocene and perhaps earlier, though the size of Greater India, the timing of
844   the continent's passage past Myanmar and the extent of strike-slip partitioning within Sundaland
845   remain unclear (e.g. Tapponnier *et al*. 1982; Treolar & Coward 1991; Replumaz & Tapponnier 2003;
846   Aitchison *et al*. 2007; van Hinsbergen *et al*. 2011; Hall 2012; Morley 2013). Emplacement of the
847   Naga Hills ophiolite, part of the Western Ophiolite Belt, represents this phase of oblique contraction
848   and occurred by the Mid-Eocene, somewhat later than Early Cretaceous emplacement of the Mt.
849   Victoria and Chin Hills ophiolites (Acharyya 2007). During the Eo-Oligocene the Mogok
850   Metamorphic Belt, lying at the core of a belt of dextral shear, experienced peak high temperature
851   metamorphism, while the Mae Ping and Three Pagodas faults of NW Thailand experienced the final
852   phases of sinistral shear (e.g. Lacassin *et al*. 1997; Searle *et al*. 2007; Morley *et al*. 2011) (Fig. 14a).

853   The Late Oligocene to Early Miocene coupling of western Myanmar and India (e.g. Curray 2005;
854   Searle & Morley 2011) led to a more direct requirement for dextral shear within central Myanmar,
855   whose earliest expression may be a diffuse zone of dextral slip initiated during the late Paleogene
856   (Fig. 14b), which generated deep-seated structural fabrics subsequently reactivated as strands of the
857   Neogene strike-slip fault system (e.g. Le Dain *et al*. 1984; Bertrand & Rangin 2003; Socquet &
858   Pubellier 2005). Long-lived faults of the Shan Plateau, such as the Kyaukkyan/Mae Ping, Panlaung
859   and Taungoo faults, may have played a part in this early shear zone.

860   Development of a more discrete region of strike-slip may have its origins in a zone of dextral
861   transtension west of the Mergui-North Sumatra Basin, at the southernmost end of the Sagaing Fault at
862   about 32 Ma (Polachan & Racey 1994; Curray 2005). The 'South Sagaing Fault' segment, which
863   bounds a Late Oligocene-Early Miocene basin, could represent the site of earliest localised dextral
864   motion (Morley 2013) (Fig. 14b).

865   Ductile stretching and diachronous cooling within the Shan Scarp and Mogok Metamorphic Belt in
866   Myanmar and the Gaoligong Shan and Chongshan in Yunnan from Late Eocene to Mid-Miocene
867   times (Wang & Burchfiel 1997; Bertrand *et al*. 2001; Bertrand & Rangin 2003; Socquet & Pubellier
868   2005; Akciz *et al*. 2008; Zhang *et al*. 2012) support a wave of northward migrating dextral shear that
869   shortly pre-dated localisation of strain along more northerly segments of the Sagaing Fault (Socquet
870   & Pubellier 2005) (Fig. 14c, d). Cessation of Late Oligocene to Early Miocene dextral strike-slip
871   basin development in northern Thailand (Morley 2004; Morley *et al*. 2011) may indicate eastwards
872   transferral of dextral slip from the Shan Plateau (resulting in stress relaxation and Early-Mid-Miocene
873   metamorphic core complex exhumation in northern Thailand) and onto central parts of the Sagaing
874   Fault or possible precursor strands along the Shan Scarp fault zone (Searle & Morley 2011; Morley &
875   Alvey 2015) (Fig. 14e). The youngest cooling ages from the exhumed metamorphic belts represent

876 final cessation of transtensional ductile stretching, and also correspond to the Mid/Late Miocene (11
877 Ma) onset of localised inversion in the Central Basin (e.g.Wang & Burchfiel 1997; Pivnik *et al.* 1998;
878 Bertrand & Rangin 2003; Akciz *et al.* 2008; Zhang *et al.* 2012) (Fig. 14f).

879 At the same time Mid-Miocene high-K calc-alkaline volcanic rocks of Mount Popa in central
880 Myanmar provide the final definitive record of eastward subduction-related magmatism beneath
881 Myanmar (Lee *et al.* 2010), which likely started in the Early Oligocene (Mitchell 1993). Pleistocene
882 high-Al basalts and absarokites at Mount Popa and Monywa and OIB-type alkali basalts at Singu
883 (Bertrand *et al.* 1998; Lee *et al.* 2010) have been previously attributed to continued post-Miocene
884 subduction (e.g. Mitchell 1993), but more recent work suggests they formed by extension-related
885 melting of subduction-enriched mantle, indicating possible Mid-Miocene (11 Ma) cessation of
886 subduction (Everett *et al.* 1990; Lee *et al.* 2010). However, the timing of subduction and possible
887 modern continuation remains contentious. An east-dipping slab below the Indo-Myanmar Ranges is
888 revealed by mantle topography and earthquake data (e.g. Mukhopadhyay & Dasgupta 1988; Li *et al.*
889 2008), and there is evidence of historic megathrust earthquakes along the western Myanmar coast
890 (Wang *et al.* 2011). Despite this, earthquakes within the east-dipping Benioff zone are dominantly
891 intra-slab strike-slip events and not interface thrusts, indicating that the slab is not presently being
892 subducted but is being translated northwards (Kundu & Gahalaut 2012). Mid-Miocene cessation of
893 subduction would have placed greater emphasis on the Sagaing Fault to accommodate lateral motion
894 between India/western Myanmar and Sundaland, and a more fully localised fault system was
895 developed (e.g. Socquet & Pubellier 2005).

896 Late Miocene to Pliocene (Morley & Alvey 2015) or Early Pliocene to Recent (Curray 2005) oceanic
897 spreading of 118 km in the Andaman Sea between the Alcock and Sewell rises was likely directly
898 connected to the Sagaing Fault and resulted in the same amount of dextral slip along a fully localised
899 structure (Kamesh Raju *et al.* 2004; Curray 2005) (Fig. 14g). An Early Pliocene to Recent period of
900 sea floor spreading directly connected to the Sagaing Fault results in a strike-slip rate of ~30 mm/yr,
901 significantly higher than post-Pliocene rates and modern geodetic rates. However, a thick sedimentary
902 package burying the spreading centre suggests that spreading occurred earlier (Morley & Alvey
903 2015), allowing more time for accumulation of 118 km of displacement at a rate more consistent with
904 observations. Assuming sea floor spreading was complete by Early Pliocene times, an additional ~70
905 km of strike-slip displacement may have been accumulated at an assumed average rate of ~18 mm/yr.
906 This displacement could have been accommodated by episodic extension and late sea floor spreading
907 in the Andaman Sea (Morley & Alvey 2015).

908 At the same time Late Miocene to Pliocene onset of widespread positive inversion in the Central
909 Basin is recorded by the poorly dated syn-inversion Irrawaddy Group and apatite fission track ages
910 (Trevena *et al.* 1991; Pivnik *et al.* 1998; Sloan *et al. this volume*). Extreme thinning/seafloor
911 spreading in the Andaman Sea and inversion in the Central Basin may be related to a regional-scale
912 kinematic reorganisation that was facilitated by localisation of the Sagaing Fault by the start of the
913 Pliocene at the latest, and probably during the Late Miocene (e.g. Bertrand & Rangin 2003). By this
914 time more than half of the total dextral shear across Myanmar may have already been accumulated
915 across a distributed array of precursor fault systems and shear zones by a rather different process
916 perhaps more similar to continuum mechanics (e.g. England & Molnar 2005).

917 Northernmost strands of the Sagaing Fault curve to the NW into structures such as the Lohit and Main
918 Central thrusts. Parts of the termination array have been rotated into a releasing geometry by crustal
919 flow from the Tibetan Plateau around the eastern Himalayan syntaxis (Rangin *et al.* 2013). Extension
920 across the rotated northern termination splays may not be of the same magnitude as extension in the

921   Andaman Sea, but geomorphic similarities with global examples of extreme continental extension
922   resulting in 'basin and range'-style topography (e.g. Coney & Harms 1984; Armijo *et al*. 1986;
923   Rhodes *et al*. 2005; Genç & Yürür 2010) suggest significant extension has occurred immediately
924   south of the eastern Himalayan syntaxis. Shallow normal fault focal mechanisms from the NEIC
925   catalogue within and around the margins of the Bhamo Basin support modern extension. In many
926   regions of basin and range extension, mid-lower crustal rocks are exposed below low-angle normal
927   faults in metamorphic core complexes (e.g. Coney & Harms 1984; Macdonald *et al*. 1993; Whitney *et*
928   *al*. 2013). It has been suggested that rocks of the Mogok Metamorphic Belt may be genetically linked
929   with metamorphic core complexes in northern Thailand (Searle *et al*. 2007), although the timing and
930   direction of unroofing of these two areas is not the same (Searle & Morley 2011). Nonetheless,
931   exhumation of other poorly known ranges of metamorphic rocks associated with the northern Sagaing
932   Fault, such as the Katha-Gangaw belt (Mitchell *et al*. 2007; Searle *et al*. 2007), should not be assumed
933   to be the result of transpression, a strike-slip driven extensional metamorphic core complex
934   mechanism must also be considered.

935   *Present day structural continuity and seismic potential*

936   Major faults form from the linkage and coalescence of smaller segments, removal of complexities
937   such as bends, step-overs, splays and rotation of pre-existing fabrics into parallelism with the through-
938   going fault. These processes are promoted at high shear strains (e.g. Spyropoulos *et al*. 1999). The
939   North Anatolian Fault (total slip: ~35 km), for example, is presently more segmented than the San
940   Andreas Fault (total slip: ~250 km) (Scholz 2002). Empirical results suggest that maximum
941   earthquake potential is proportional to the length of ruptured segments (e.g. Wells & Coppersmith
942   1994), so that faults become capable of generating larger magnitude earthquakes as their cumulative
943   displacement and hence segment length increases.

944   Additionally, like the San Andreas Fault, the Chaman Fault (the Sagaing Fault's sinistral counterpart
945   bounding the western margin of India) and the Sumatran Fault south of the Andaman Sea, the Sagaing
946   Fault is considered to be so continuous that its central 700 km has been classed as an 'earthquake fault
947   superhighway', capable of supershear ruptures (Robinson *et al*. 2010). Supershear occurs when
948   rupture speed exceeds S-wave or even P-wave speed, resulting in Mach shock cones and increased
949   ground acceleration, and is more likely on long continuous faults along which rupture can accelerate
950   without interruption from fault heterogeneities (e.g. Kostrov & Das 1988; Bernard & Baumont 2005;
951   Robinson *et al*. 2010). Consequently, important questions about the modern Sagaing Fault are: 1)
952   whether it is as continuous as its gross geomorphology suggests (e.g. Le Dain *et al*. 1984; Vigny *et al*.
953   2003); 2) whether it possesses step-overs between its segments (e.g. Wang *et al*. 2014); and 3)
954   whether these are large enough to arrest or slow down seismic ruptures (e.g. Wesnousky 2006;
955   Robinson *et al*. 2010).

956   Prominent releasing step-overs and bends, such as are occupied by lakes Indaw and Indawgyi (Figs.
957   8b & c), characterise the northern splays of the Sagaing Fault. Indaw Lake, a step-over ≤2.5 km wide,
958   has a similar size, geometry and structural setting as Lake Sapanca, Turkey, which separates the
959   Sapanca and Sakarya segments of the North Anatolian Fault. Despite its 1-2 km step-over width, the
960   Sapanca step-over failed to arrest the 17th August 1999 M_w 7.4 Izmit earthquake (Lettis *et al*. 2002),
961   in the same way that the Indaw step-over was likely breached by the first (M_w 7.3) of the Sagaing
962   Fault's twin 1946 earthquakes which probably ruptured the Indaw and Mawlu segments (Hurukawa &
963   Phyo Maung Maung 2011; Wang *et al*. 2014). Indawgyi is significantly wider than Indaw and Lake
964   Sapanca, but may be bypassed by a through-going fault below alluvial fans on its southern side.

965   Several step-overs were revealed along the 2012 Thabeikkyin surface rupture of the apparently
966   continuous central part of the Sagaing segment, including south of Thabeikkyin city (Fig. 8d), and at
967   Lower Ponna (Fig. 11d). Each of these was clearly breached by the fault rupture, and it is likely that
968   they reflect either upward curvature of fractures within unconsolidated overburden from a simple and
969   continuous basement fault, or minor/shallow segmentation of the basement fault only (Fig. 15a).
970   Helicoidal upward curvature of shears forming *en echelon* surface patterns is well known from
971   analogue models (e.g. Mandl 1988; Richard *et al.* 1995) and similarities between models and natural
972   examples have been recognised (e.g. Dooley & McClay 1997). This process may mask fault
973   simplicity at the depth of large earthquake nucleation (Graymer *et al.* 2007) or even immediately
974   below the surface sediment layer, as was observed in the 2001 Kunlun earthquake (e.g. Lin &
975   Nishikawa 2011). Such a process may also be responsible for Sagaing Fault step-overs such as Yega
976   In (Fig. 9b) and those along the Meiktila segment (Fig. 4b) as well as smaller-scale shallow
977   complexities (Fig. 15b). Nonetheless, the 2012 rupture terminated at the 1.5 km wide Kyet Phyu
978   Taung lake step-over, indicating that the lake marks genuine basement fault segmentation.

979   The Nay Pyi Taw double bend is the most significant complexity along the superficially most
980   continuous central 700 km of the Sagaing Fault. Though it comprises a 10 km wide releasing bend
981   and a segmented restraining bend (Fig. 9c), the entire system is bypassed by an apparently continuous
982   shortcut fault (Wang *et al.* 2014) that would be preferentially exploited by a high speed rupture.
983   Similarly, the prominent restraining bend between the Bago and Pyu segments (Fig. 9d) is breached at
984   the surface by almost continuous reverse faults, and is less than 1 km wide in any case.

985   Even if supershear rupture is limited by the fault heterogeneities discussed above, coseismic strain
986   transfer across step-overs can still be accommodated by stress transfer or hard linkage (Scholz 2002).
987   The efficacy of such transfer is a function of the source magnitude/displacement characteristics and
988   the width of the step-over (e.g. Lettis *et al.* 2002; Wesnousky 2006). Step-overs 1-2 km wide are
989   found to be breached by most historic earthquakes (Lettis *et al.* 2002), while those 3-4 km
990   (Wesnousky 2006) or 4-5 km wide (Lettis *et al.* 2002) arrest most historic earthquakes. It is clear that
991   the onshore Sagaing Fault possesses no step-overs wider than about 2.5 km, with the possible
992   exception of the poorly exposed region between the Sagaing and Meiktila segments (Wang *et al.*
993   2014), which may mark a releasing structure 4 km wide. Sagaing Fault earthquakes possibly arrested
994   by step-overs include the 1930 Pyu (18°N restraining bend), 1946 Wuntho (southern Indaw segment
995   step-over), 1956 Sagaing (Southern Sagaing segment step-over) (Wang *et al.* 2014), and the 2012
996   Thabeikkyin (Kyet Phyu Maung lake step-over) earthquakes.

997   Segment boundaries aside, all five of the main segments south of the northern splay possess
998   apparently uninterrupted portions >100 km long. Though not as great as the 700 km continuous
999   supershear segment originally proposed, it remains possible that that the Sagaing Fault is an
1000 'earthquake fault superhighway' as defined by Robinson *et al.* (2010). Critical unanswered questions
1001 include: can identified structural complexity, though not significant enough to terminate a rupture,
1002 reduce rupture velocity below supershear speeds?; and is structural complexity at the surface a
1003 reliable expression of fault geometry at seismogenic depths?

1004 When considering the seismic hazard of the Sagaing Fault, there clearly remain significant
1005 uncertainties about the fault's seismic cycle, segmentation, off-fault strain distribution and the
1006 contribution of aseismic creep. Wang *et al.* (2014) proposed maximum magnitudes for full single
1007 segment rupture earthquakes along the Bago, Pyu, Nay Pyi Taw, Meiktila and Sagaing segments as
1008 7.7, 7.5, 7.1-7.2, 7.8-7.9, 7.7 respectively, and for the northern segments from 7.0-7.2 (Tawma
1009 segment) to 7.9-8.0 (Mogang segment). These estimates are supported by the series of M>7.0
1010 earthquakes that ruptured over half of the entire fault during the first half of the twentieth century (e.g.

1011   Hurukawa & Phyo Maung Maung 2011; Wang *et al*. 2014). Based on analysis of historical seismicity
1012   and tectonic geomorphology Wang *et al*. (2014) additionally suggest whole segment failures resulting
1013   in ~ M_w 7.7 earthquakes should occur every 300-400 years, and that partial segment failures expressed
1014   by M_w 6.8 to M_w 7.0 earthquakes have short decadal recurrence intervals. Examples of the latter
1015   include the partial failure of the Sagaing segment at Thabeikkyin in 2012 just 66 years after the 1946
1016 M_w 7.7 failure, and the failure of the Indaw segment in both 1946 and 1991. Recurrence of 1930-type
1017   events along the Bago segment is likely to be >160 years (Tsutsumi & Sato 2009), but recurrence of
1018   any earthquake close to Bago (i.e. including both the Pyu and Bago segments) is likely to be between
1019   90 and 115 years (Wang *et al.* 2011).

1020   Two long sections of the Sagaing Fault lack clear records of historic seismicity (Hurukawa & Phyo
1021   Maung Maung 2011). A ~180 km long section in the Andaman Sea south of 16.6°N lacks historic
1022   seismicity since at least 1897, and a ~260 km long section straddling the Nay Pyi Taw and Meiktila
1023   segments similarly lacks seismicity, with the last probable rupture of the Meiktila segment being the
1024   1839 Amarapura event (Tsutsumi & Sato 2009; Wang *et al*. 2014). Complete failure of these segments
1025   would result in >M7.7 and M7.9 earthquakes respectively according to empirical magnitude-length
1026   relationships (Tsutsumi & Sato 2009). Assuming a conservative 18 mm/yr instantaneous slip rate (e.g.
1027   Maurin *et al*. 2010) for the whole Sagaing Fault, ~3.2 m of displacement has accumulated across the
1028   Nay Pyi Taw seismic gap between 1839 and 2015. Citing evidence that the 1839 rupture breached the
1029   Meiktila-Sagaing segment boundary, Wang *et al*. (2014) go further to propose a scenario whereby
1030   both segments fail together, resulting in a M_w 8.1-8.3 event with a recurrence interval of 500-1000
1031   years. However, it is not yet known what contribution, if any, there is from aseismic creep along
1032   segments well-constrained by GPS surveys (e.g. Maurin *et al*. 2010), much less the understudied
1033   Meiktila and Nay Pyi Taw segments.

Concluding remarks

1035   The localised and continuous modern Sagaing Fault can be considered as a relatively young (Late to
1036   post-Miocene) structure that accommodated about 100-118 km of localised dextral strike-slip strain
1037   since the onset of Andaman Sea spreading (e.g. Bertrand & Rangin 2003; Kamesh Raju *et al*. 2004;
1038   Curray 2005; Morley & Alvey 2015), despite its position on a more diffuse and longer-lived dextral
1039   boundary zone perhaps >200 km wide. This zone may have accommodated a total of 300-700 km
1040   offset since the Eo-Oligocene, along structures including the Shan Plateau metamorphic belts and
1041   associated upper crustal fault systems now largely eroded, the Shan Scarp/Kyaukkyan fault systems,
1042   the faults of western Thailand as well as embryonic strands of the Sagaing Fault (e.g. Mitchell 1977,
1043   1981, 1993; Curray *et al*. 1982; Replumaz & Tapponnier 2003; Curray 2005; Morley 2013). This
1044   represents the transition from Pre-Pliocene continuum mechanics, whereby the Myanmar crust above
1045   eastward subduction was deforming as a thin sheet with spaced strike-slip faults of relatively small
1046   local displacement, such as is currently happening in much of central Asia (e.g. England & Molnar
1047   2005); to a Pliocene-Recent high strain zone of nearly continuous localised faulting from the
1048   Andaman Sea towards the eastern Himalayan syntaxis. Localisation at the fault's current position,
1049   rather than along longer-lived structures such as the Kyaukkyan/Mae Ping fault system or Shan Scarp
1050   Fault is probably a result of northward propagation of fault systems at the northern end of the
1051   Andaman Sea (e.g. Morley 2013), coupled with thermally weakened lithosphere at the margin of the
1052   Myanmar Central Basin.

1053   The Sagaing Fault's structural continuity, high slip rate and tendency to produce large (possibly
1054   supershear) earthquakes means it has a high seismic hazard and poses a great risk to the cities that lie
1055   along or close to it, particularly Yangon, Bago, Nay Pyi Taw, Mandalay and Myitkyina (Robinson *et*

1056   *al*. 2010; Yeats 2012). Historical earthquakes in Myanmar have had relatively small death tolls, the
1057   worst on record being the 1930 Bago earthquake, which killed 550 people (e.g. Win Swe 2011).
1058   However, in contrast to 1930, Myanmar's population is now 51.5 million, an increase of 350% since
1059   the 1931 census (Department of Population, Ministry of Immigration and Population 2015). Twenty
1060   million people alone currently live in the nineteen districts cut by the Sagaing Fault, of whom almost
1061   half live in urban settings (Department of Population, Ministry of Immigration and Population 2015),
1062   many within a few kilometres of the Sagaing Fault and often close to major rivers, both factors that
1063   dramatically increase the likelihood of earthquake damage (Luo *et al*. 2012). In Mandalay urban
1064   population has increased from 163,527 (1948) to its present level of 1,225,546 (Andrus 1948;
1065   Department of Population, Ministry of Immigration and Population 2015), indicating a
1066   disproportionate urban growth rate likely observed in numerous cities along the fault. Urban
1067   population growth and increased building fragility led Wyss (2005) to propose a 3-60 times increase
1068   in expected casualties for repeat earthquakes in India compared to measured losses in the last century,
1069   a calculation that could be applied to the next Bago or Amarapura (Mandalay) earthquake, for
1070   example.

1071   An additional factor which increases the risk resulting from Sagaing Fault seismic hazard is
1072   corruption in the public sector. Anomalously high earthquake death tolls are associated with corrupt
1073   countries; devastating earthquakes in Haiti (2010) and Pakistan (2005), for example, were made worse
1074   by corruption in construction industries (Ambraseys & Bilham 2011). In 2014 Myanmar was ranked
1075   amongst the 20 most corrupt countries in the world, alongside Haiti and worse than Pakistan
1076   (Transparency International 2014). To mitigate the risk posed by the Sagaing Fault it is essential that
1077   work to improve transparency in the building industry, overall building quality and public awareness
1078   along the fault is carried out alongside future structural, geodetic and palaeoseismic studies.

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Figure and table captions

1514 **Fig. 1:** Tectonic setting of the Sagaing Fault. Modified after Morley *et al.* (2011); Soe Thura Tun &
1515 Maung Thein (2012); Wang *et al.* (2014); Morley & Alvey (2015).

1516 **Fig 2:** Overview maps of the onshore Sagaing Fault. a) Topography, geomorphic and structural
1517 elements of the Sagaing Fault. Segments modified after Wang *et al.* (2014). Digital elevation model
1518 basemap constructed from ASTER GDEM. ASTER GDEM is a product of METI and NASA.
1519 Yellows, browns and whites are elevations >400 m. Greens and blues are lower elevations. For
1520 location see Fig. 1. b) Geology along the Sagaing Fault. Geology modified after Myanmar
1521 Geosciences Society (2014). c) Seismicity along the Sagaing Fault. Earthquake hypocentres (red)
1522 1970-2015 with depths <50 km from USGS compilation
1523 (http://earthquake.usgs.gov/earthquakes/search/). Focal mechanisms for selected events from the
1524 CMT catalogue plotted using Mirone software. Relocated earthquake locations (blue) from Hurukawa
1525 & Phyo Maung Maung (2011) and (*) from Wang *et al.* 2014. Rupture extent for main historic
1526 earthquakes modified after Wang *et al.* (2014). Deeper colours are more recent. Dashed lines indicate
1527 uncertainty. Line thickness is set for the purpose of clarity only.

Fig. 3: Landsat ETM+ mosaic (bands 742) of the Sagaing Fault, showing the main geomorphic domains. Main trace of the Sagaing Fault is highlighted in white. Soils and thin vegetation appears grey, pink and purple. Thick vegetation appears deep green.

Fig. 4: Tectonic geomorphology of the Sagaing Fault. For locations see Fig. 3. a) Low relief and subtle pressure ridges south of Bago. b) Growing fold adjacent to the Meiktila segment. Brown lines indicate dry streams, blue lines indicate active rivers. See text for details. Basemap constructed from ASTER GDEM. c) Triangular facets marking a strand of the Sagaing Fault on the east bank of the Ayeyarwaddy River south of Sabeanago. Image taken in April 2013, most of the facets have been stripped by landslides during the November 2012 earthquake. d) Dry gulley showing a cumulative 1.75 m offset caused by 2012 and 1946 earthquakes. e) Possible palaeoseismic scarp in alluvial fans at the foot of the main Sagaing Ridge. Note stream erosion and partial collapse of the scarp.

Fig. 5: Satellite imagery of Sagaing Fault geomorphology. For locations see Fig. 3. a) Remarkably linear trace of the fault across deltaic sediments immediately north of the Gulf of Mottama. Imagery from the ERSI Global Imagery database and sources therein. b) Landsat ETM+ image of the Pyu segment (bands 451 with panchromatic sharpening) draped over ASTER DEM. Streams (blue) picked as they cross main faults. Wet soils appear blue, dry soils appear green, dense vegetation is red. c) Lineaments defined by sandbanks in the Ayeyarwaddy River between Male and Sabeanago, highlighting Sagaing Fault strands and vertical motions. Imagery from the ERSI Global Imagery database and sources therein.

Fig. 6: Examples of pressure ridges along the Sagaing segment. a) Small coseismic pressure ridge defined by Riedel shears and thrust faults, formed north of Singu during the 2012 earthquake. b) Pre-existing pressure ridge within a fault-bounded valley north of Thabeikkyin, amplified by the 2012 earthquake. c) One of a number of parallel and en-echelon ridges exposing Irrawaddy Gp. strata north of Singu. d) Steep-sided shutter ridge exposing tectonised amphibolites close to the western margin of the Shan Plateau. e) Representative of the highly linear, symmetric pressure ridges lining the Ayeyarwaddy River along the northern half of the Sagaing segment. f) View along Sagaing Ridge, showing the steeper western flank and more shallow eastern flank.

Fig 7: Overview map showing topographic, geomorphic and structural elements of the northern termination and transform system. Faults modified after Morley *et al.* (2011); Soe Thura Tun & Maung Thein (2012); Wang *et al.* (2014). Segments modified after Wang *et al.* (2014). Digital elevation model basemap constructed from ASTER GDEM. Greys, browns and pinks are elevations >300 m. Greens and blues are lower elevations. For location see Fig. 1.

Fig. 8: Details of fault complexities along the northern Sagaing Fault, mapped using ASTER GDEM, Landsat ETM+ and ERSI Global Imagery (except d, surveyed on the ground). For locations see Fig. 3. a) Ban Mauk segment and horsetail splay. b) Indaw Lake step-over basin. c) Strike-slip splays and extensional systems around Indawgyi Lake. d) Coseismic fractures revealing small step-over and zone of subsidence formed during the 2012 earthquake.

Fig. 9: Details of fault complexities along the southern Sagaing Fault, mapped using ASTER GDEM, Landsat ETM+ and ERSI Global Imagery and incorporating field observations. For locations see Fig. 3. a) Singu basalt plateau showing the southern termination of the 2012 surface rupture. b) Sagaing Ridge, showing Yega In and fault segmentation along the western margin. c) Nay Pyi Taw double bend system. d) Restraining step-over between the Pyu and Bago segments.

Fig. 10: Large-scale offset piercing points along the Sagaing Fault. a) Restoration of 340 km displacement along the Sagaing Fault to bring the upper Chindwin into continuity with the Ayeyarwaddy River. After Hla Maung (1987). b) Present-day arrangement of the Chindwin and Ayeyarwaddy, showing the assumed piercing points used for the restoration in a). c) 203 km displacement along the Sagaing Fault based on restoration of the Mayathein and Sagaing metamorphics, presently exposed on the west and east sides of the fault respectively. After Myint Thein *et al.* (1991).

Fig. 11: Smaller-scale offset piercing points along the Sagaing Fault. a) Present-day outline of the Pleistocene Singu basalt straddling the Sagaing Fault. b) Restoration of 2.7 km offset using the southern margin of the basalt (yellow points). c) Restoration of 6.5 km offset using the northern margin of the basalt (red points). After Bertrand *et al.* (1998). d) Displacement and capture of streams crossing the Sagaing Fault south of Thabeikkyin. Solid blue lines are modern rivers, dashed lines are abandoned channels. Blue arrows show flow direction. Deltas marked in brown. Faults based on field mapping of the 2012 surface rupture. See text for explanation.

Fig. 12: GPS vectors from all published surveys within Myanmar and around the eastern Himalayan syntaxis. After Maurin *et al.* (2010). Stations referred to in the text are marked as yellow triangles.

Fig. 13: Effects of earthquakes along the Sagaing Fault. a) Mingun Pagoda on the east side of Sagaing Ridge, a solid brick construction damaged during the 1839 Amarapura earthquake. b) Thein Taung pagoda, built on a pressure ridge south of Sabeanago, damaged during the 2012 Thabeikkyin earthquake. c) Typical expression of the 2012 surface rupture, showing well-developed Riedel shears and a component of down-dip extension.

Fig. 14: Schematic evolutionary maps showing the development of the Sagaing Fault since Late Eocene times, synthesised largely from ideas presented in Pivnik *et al.* (1998); Bertrand & Rangin (2003); Socquet & Pubellier (2005); Morley *et al.* (2011); Searle & Morley (2011); Morley & Alvey (2015). See text for explanation and additional references. Shows progressive westwards migration and replacement by discrete strike-slip structures of a broad zone of dextral shear extending from the southeastern Shan Plateau to the Gaoligong Shan and beyond. Major change from broadly NNW-SSE directed extension to NE-SW compression (Socquet & Rangin 2005) coincides with the localisation of the Sagaing Fault, basin inversion, cessation of subduction and cessation of metamorphic complex exhumation. No attempt is made to show finite displacement across the fault systems because of the large uncertainty involved. The modern coastline is for reference only.

Fig. 15: The disparity between surface fault traces and 3-D geometry of strike-slip faults. a) Simple, through-going fault in seismogenic basement splays upwards into 'cover' units, which may be Quaternary drift deformed by coseismic fractures, or at a different scale the uppermost layers of rock. Surface trace shows complex splay patterns defined by Riedel (R) and antithetic Riedel (R') shears, synthetic P-shears and through-going Y-shears, together with areas of uplift and subsidence both within strands and at step-overs. Adapted after Lin & Nishikawa (2011). b) Example of upward-splaying from an inferred single lower fault strand of the Sagaing Fault observed in a quarry wall. Coseismic fractures in Quaternary sands and silts north of Singu, formed during the 2012 Thabeikkyin earthquake.
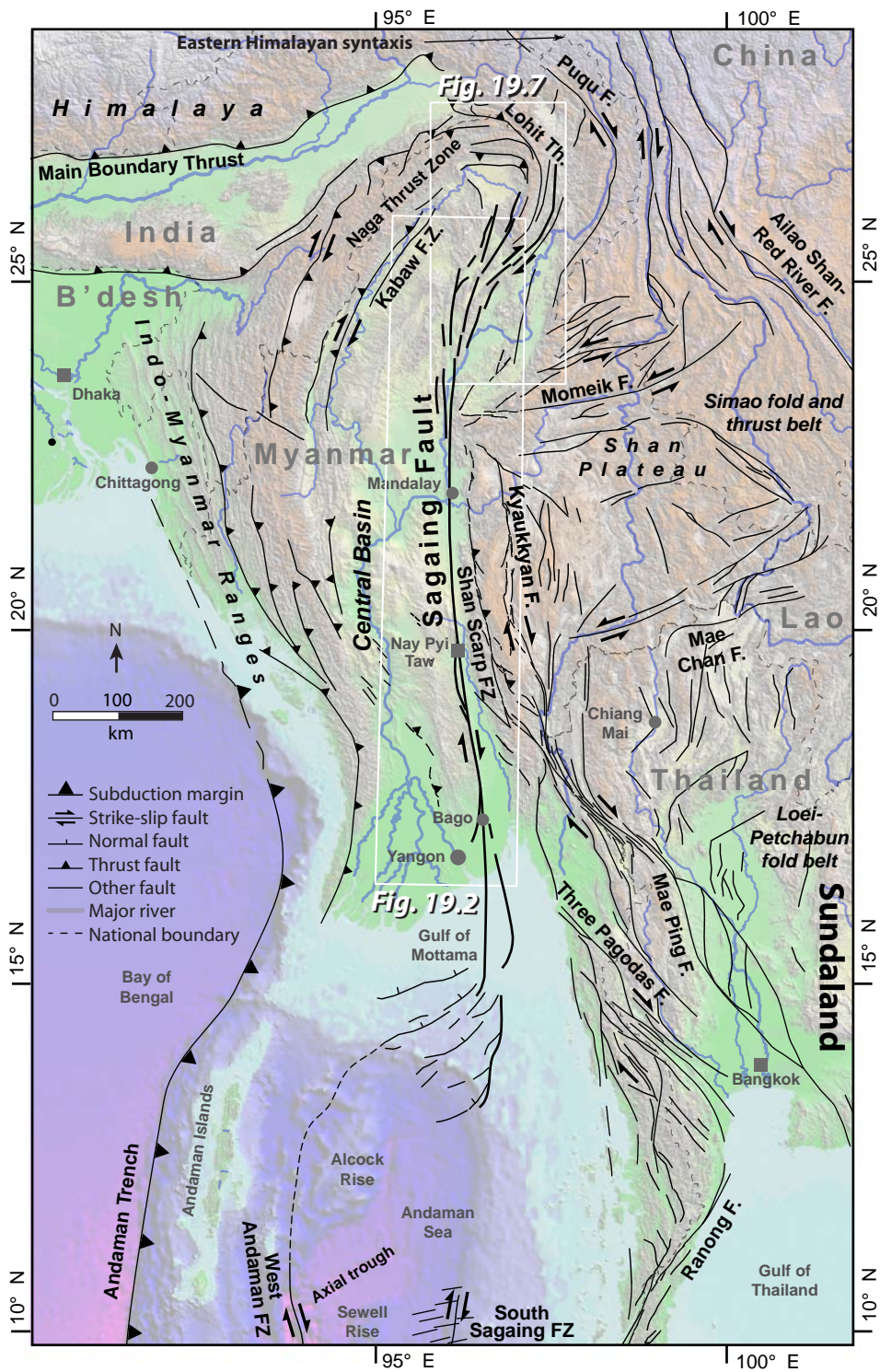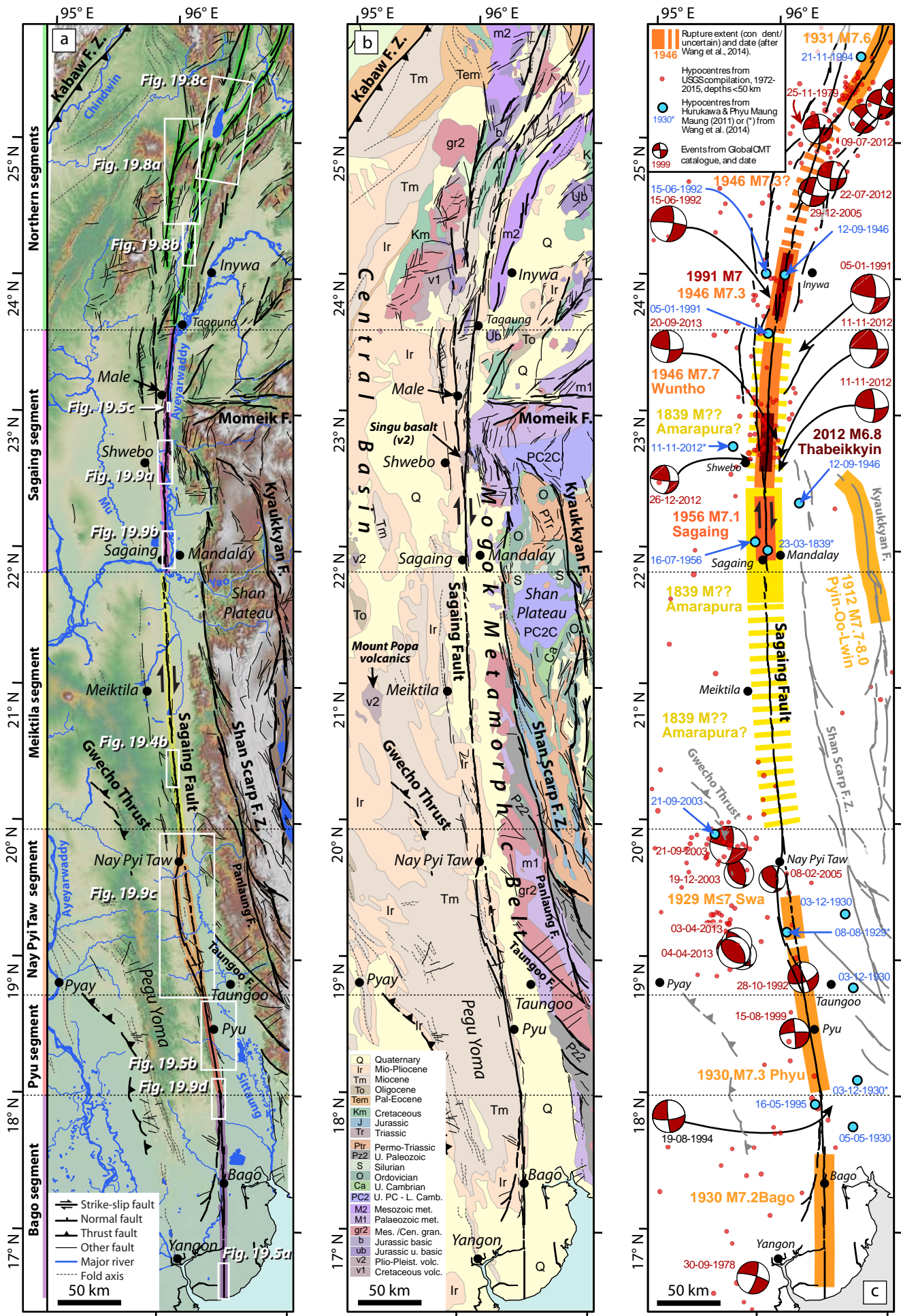
Fig. 19.1

Fig. 19.2
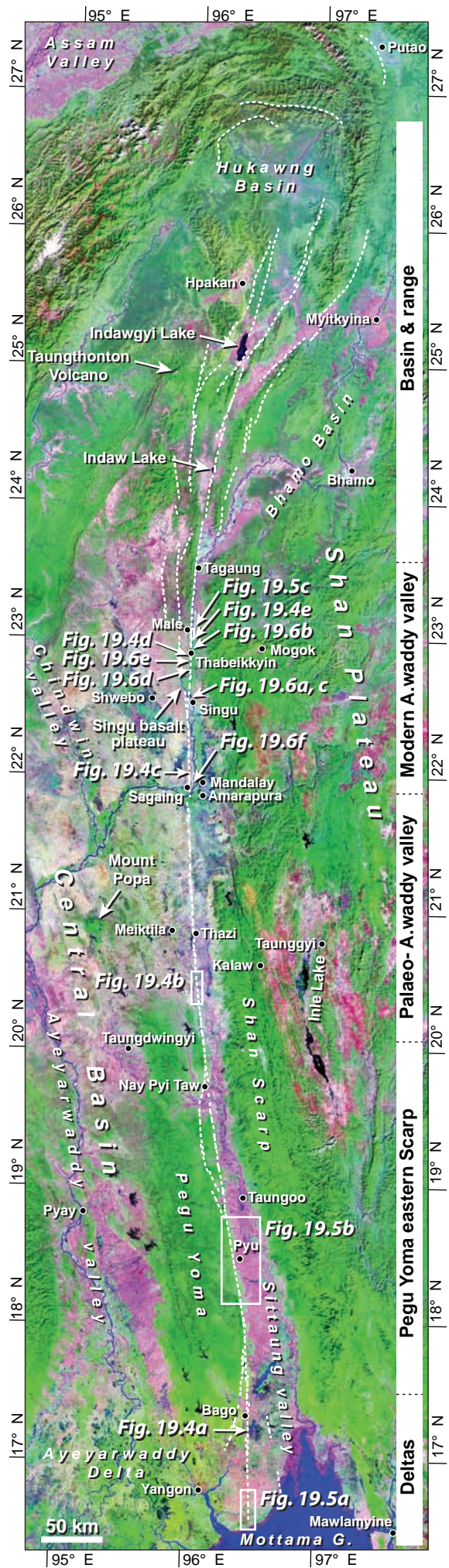
Fig. 19.3

Fig. 19.4

a

Thongwa village

Coastal terrace

Coastal terraces

Modern coast

2 km

b

96.5° E

Sittaung River

Older fans

Uplifted older fans

18.5° N

Stream A

Pyu

Stream B

Pegu Yoma

Offset stream A palaeo-fan

Sagaing Fault

Uplifted older fans

Older fans

96.5° E

5 km

c

Male

D

Ayeyarwaddy River

Sabenago sand banks

U

D

Pressure ridge

D

U

D

D

100 m

Fig. 19.5

a — View NNW

Riedel shear

Riedel shear

Riedel shear

En-echelon folds

Thrust front

0.5 m

b — View SSW

Uplift along cross-basin fault system

Broad sag-pond basin

Nov. 2012 surface rupture

c — View WNW

Eroded fragments of overlying Pleistocene basalt

Folded Irrawaddy Formation sediments

Sag pond

d — View NE

Right-lateral stream offset across fault zone

Collapse of over-steepened slopes

Active fault trace

e — View SSE

Left-lateral stream offset around uplifted delta

Fault-controlled Ayeyarwaddy valley

Uplifted delta

f — View NNW

Sagaing Ridge

Fig. 19.6

Fig. 19.7

Fig. 19.8

Fig. 9
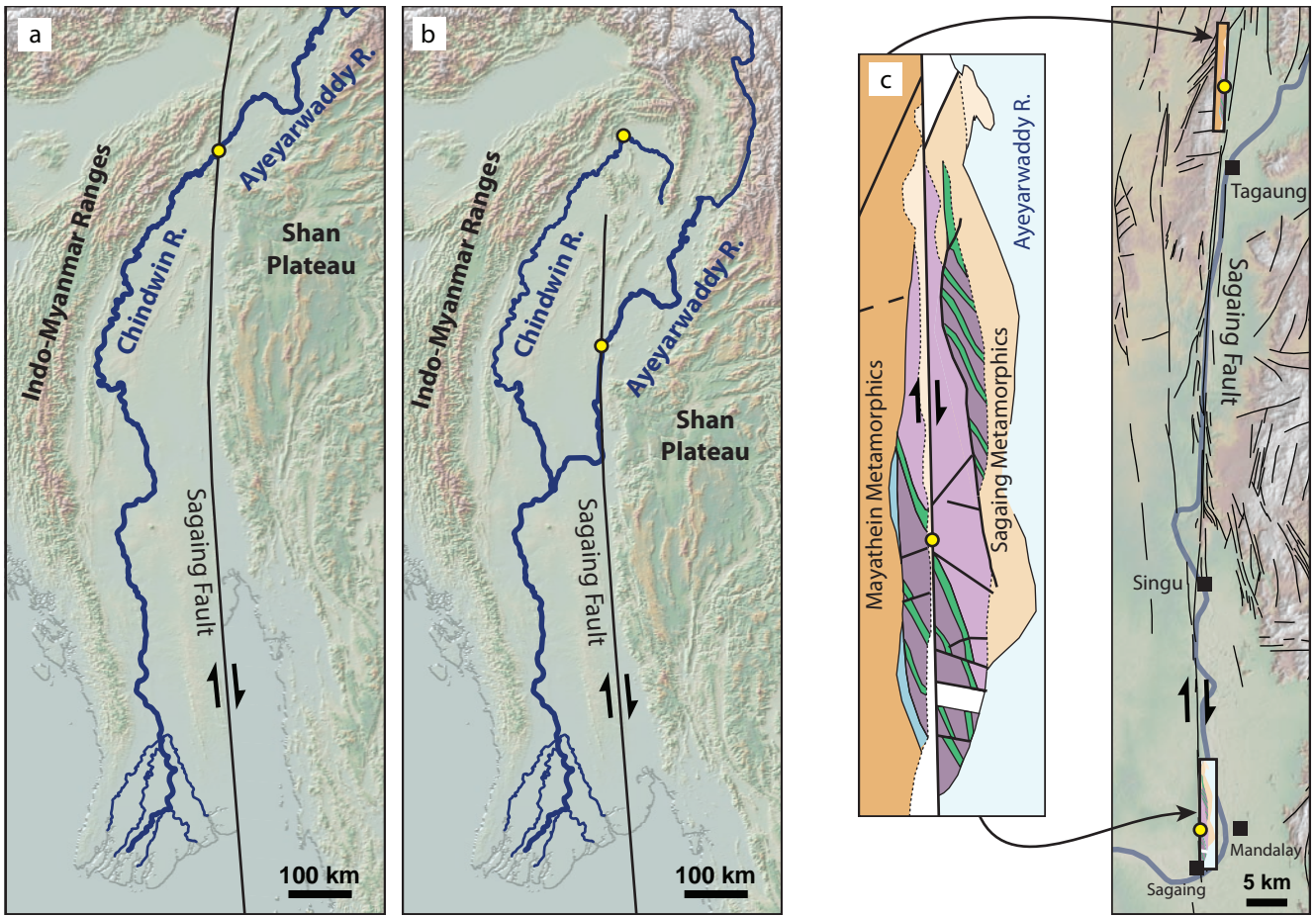
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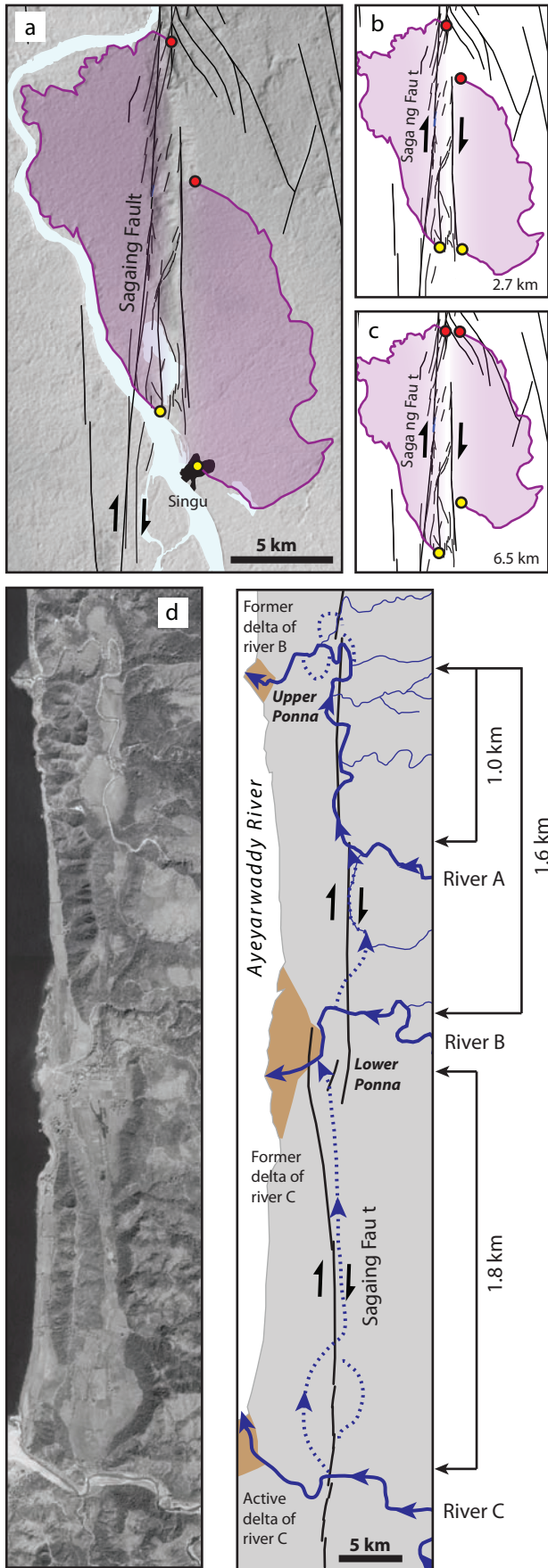
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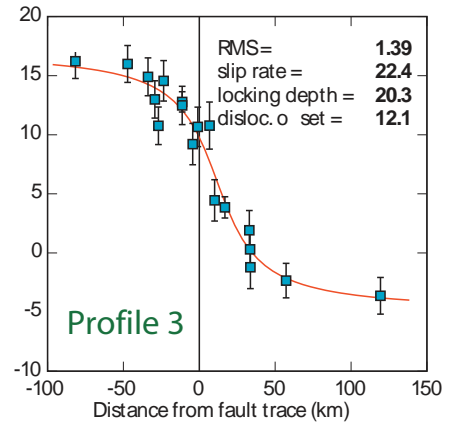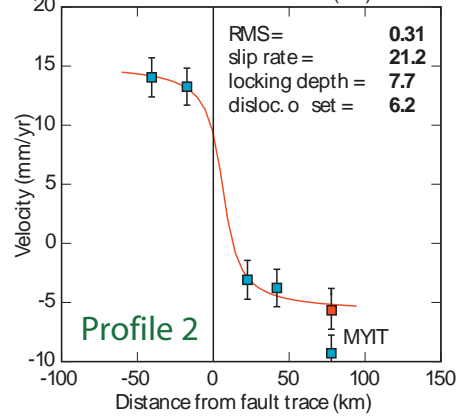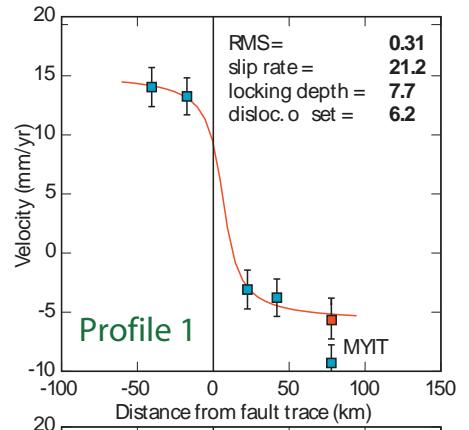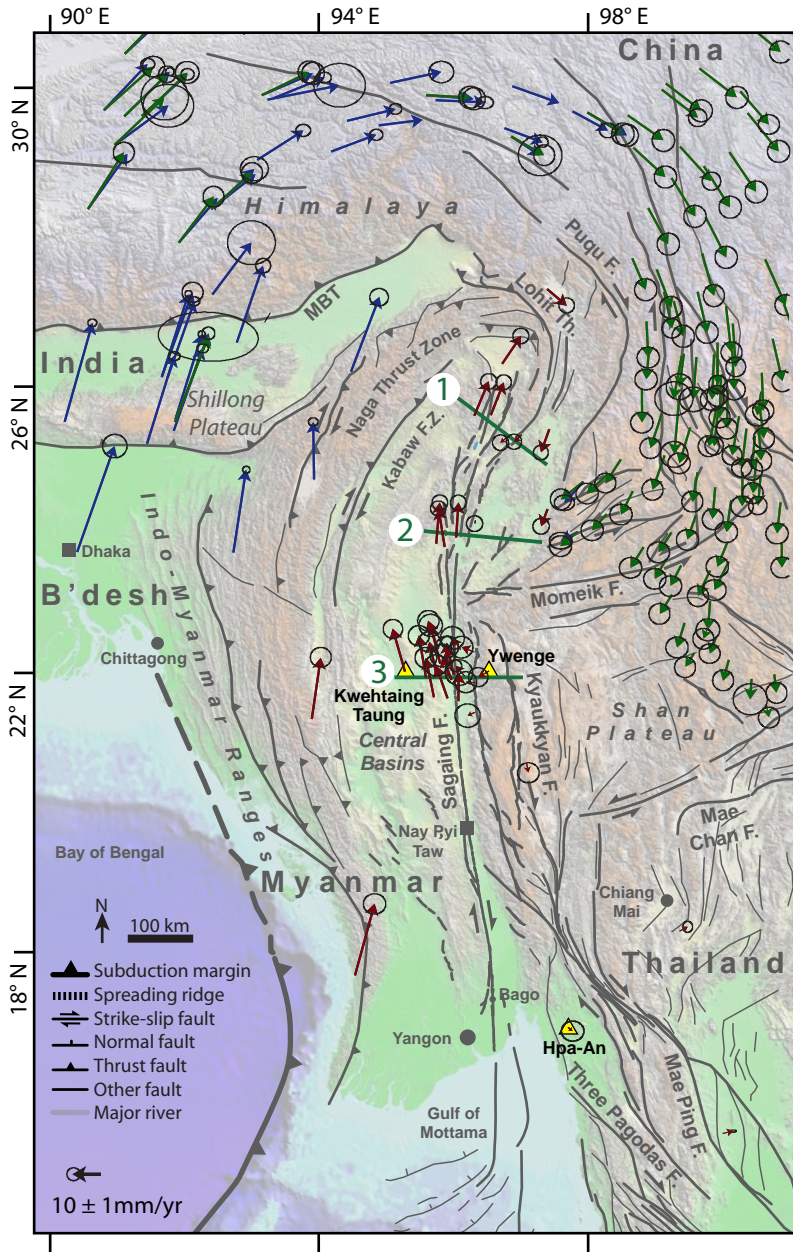
Fig. 19.12

Fig. 19.13

Fig. 19.14

a

Subsidence

Uplift

R

R

R'

Y

P

Cover

Seismogenic basement

Pre-exsting fault

b

~ 1 m

Ground surface

Quarry face

Fractures:
— Main
— Minor
-- Uncertain
··· Bedding

Quarry lake

Fig. 19.15