Chapter 2

Active tectonics of Myanmar and the Andaman Sea

R. A. SLOAN1,2*, J. R. ELLIOTT1,2, M. P. SEARLE1 & C. K. MORLEY4,5

1COMET+ Department of Earth Sciences, Oxford University, South Parks Road, Oxford OX1 3AN, UK
2Department of Geological Sciences, University of Cape Town, Cape Town, Republic of South Africa
3Present address: COMET, School of Earth & Environment, University of Leeds, Leeds LS2 9JT, UK
4PPT Exploration and Production, Vibhavadi-Rangsit Road, soi 11, Bangkok, Thailand 10900
5Present address: Department of Geology, Chiang Mai University, Thailand
*Correspondence: alastair.sloan@uct.ac.za

Burma (Myanmar) is situated along the obliquely colliding eastern margin of the Indian Plate (Fig. 2.1). It connects the ongoing India–Asia collision along the Himalayan Range in the north to the newly formed oceanic crust in the Andaman Sea in the south, via the active plate margin along the dextral Sagaing Fault and subduction along the Burma Seismic Zone. The active tectonics of Myanmar are controlled by the combination of: (1) the continuing northwards penetration of India into Asia, ongoing since the initial collision and closing of the Tethyan ocean c. 50 Ma; (2) the active dextral shear along the right-lateral Sagaing strike-slip fault, and region to the west across the Indo-Burma Ranges or Indo-Myanmar Ranges (IMR); (3) the active eastwards-dipping Burma Seismic Zone that indicates subduction of a downgoing plate to depths of >150 km; (4) the clockwise rotation around the Eastern Himalayan Syntaxis and a series of arcuate strike-slip faults (e.g. Nanting and Wanding faults) in the northern Shan Plateau; and (5) the active extensional and strike-slip tectonics in the back-arc Andaman Sea. In this paper we review the active tectonics of this region and discuss the interactions and driving mechanisms of deformation in and around Myanmar.

The northwards motion of India relative to Eurasia resulted in the closing of the intervening Neotethyan Ocean at c. 50 Ma (Green et al. 2008) and formation of the Himalayan mountain range. In this time, India–Eurasia convergence has slowed from c. 15 cm a\(^{-1}\) to c. 4 cm a\(^{-1}\) (Molnar & Stock 2009; Copley et al. 2010). GPS velocities show that convergence along a NNE azimuth continues to this day north of the Himalaya, with a significant shift to more easterly directed shear in NE and eastern Tibet (Gan et al. 2007). GPS relative motions and the traces of active faults show a major curvature around the Eastern Himalayan syntaxis region that involves clockwise crustal rotations in Yunnan and northern Myanmar. Since the initiation of the dextral Sagaing Fault, the eastern plate margin of India has shown an oblique convergence direction with transpressively uplifted mountain ranges and transtensional sedimentary basins to the west of the fault (e.g. Pivnik et al. 1998; Bertrand & Rangin 2003).

The Sagaing Fault is the major active structure in Myanmar, accommodating right-lateral motion at current rates of c. 18–22 mm a\(^{-1}\) (Vigny et al. 2003; Maurin et al. 2010). In the north, the fault splays into several strands that bound transpressively uplifted mountain ranges composed of the Katha-Gangaw and Tagaung-Myitkina metamorphic belts (Searle et al. 2007). In the south the Sagaing Fault extends into the Andaman Sea, where it eventually connects to the present-day back-arc spreading centre (Curray 2005; Morley 2012).

The presence of ongoing subduction of an Indian slab beneath Myanmar is controversial. Seismological evidence has been cited both in support of (Stork et al. 2008; Hurukawa et al. 2012) and against (Kundu & Gahalaut 2012; Rangin et al. 2013) ongoing subduction at the present day. This debate has important implications for our understanding of the seismic hazard and tectonics of the region. The petrology and geochemistry of andesites, dacites and rhyolites erupted from three Phocene-active calc-alkaline volcanoes above the Burma Seismic Zone suggests that they were derived from melting above a subducting slab of oceanic lithosphere (Stephenson & Marshall 1984).

With respect to the progression of deformation through time, a number of tectonic features (such as the Tripura Fold Belt, uplift of the Shillong Plateau and the onset of the Sagaing Fault) appear to have initiated quite late in the history of the Indian–Eurasian collision (within the last c. 20 Ma; e.g. Clark & Bilham 2008; Maurin & Rangin 2009). Additionally, some major faults (such as the Wanding, Menglian and Menxing faults) appear to have reversed their sense of slip during this time (e.g. Lacassin et al. 1998). Understanding this reorganization is an important step in understanding the dynamics that drives the deformation and development of active structures in this complex system.

Central Burma Basin and the Sagaing Fault

Central Burma Basin

The Central Burma Basin (CBB) represents a relatively aseismic sliver of material set within an extremely tectonically complex region (Figs 2.1 and 2.2). The CBB lies between the Sagaing Fault and the Indo-Burman Ranges. The CBB is affected by the Indian Plate (which is moving rapidly northwards relative to stable Eurasia) and the southeastern margin of the Tibetan Plateau, where material is moving southeastwards relative to stable Eurasia, driven by the extrusion of Tibetan material between the Eastern Syntaxis and the Sichuan Basin. Consequently the CBB lies at the centre of a region undergoing intense right-lateral shear. Much of this strain is being accommodated by the north–south-striking Sagaing Fault (18–22 mm a\(^{-1}\)). This fault has dominated the historical and instrumental records of seismicity in Myanmar, and we
describe it in more detail below. The majority of other seismicity recorded beneath Myanmar occurs within the subducted Indian Plate, at depths of up to 160 km (e.g. Stork et al. 2008). These earthquakes form the Burma Seismic Zone, which is discussed in detail in the section ‘Seismicity within the downgoing Indian plate’. First we briefly discuss the tectonics and seismicity of the lowlands of Myanmar west of the Sagaing Fault.

Southern Central Burma Basin

South of 23° N the CBB underwent dextral transpressional deformation during the Late Miocene–Recent, creating a series of en echelon folds and some thrusts which form the hilly Pegu Yoma region and the larger, more widely spaced structures of the western Central Burma Basin (e.g. Pivnik et al. 1998; Bertrand & Rangin 2003) (Fig. 2.3). These latter structures include the major fold-and-thrust hydrocarbon traps. Overall the structural style corresponds well with the classic strike-slip structural styles described for California (e.g. Harding & Lowell 1979; Sylvester 1988). Anticlines are commonly related to thrust fault development, but are also strongly compartmentalized by normal and oblique-slip faults that trend at high angles to the fault axis, a style more typical of strike-slip deformation (e.g. Harding & Lowell 1979) than compressional fold-and-thrust belts. Further east, structures are generally larger and simpler and become intense and complex passing eastwards towards the Sagaing Fault in the Pegu Yoma area. Perhaps as much as 18 km thickness of Cenozoic sedimentary rocks is present in parts of the Central Basin (Pivnik et al. 1998). Consequently, deformation of such thick sequences does not permit the development of simple fault systems traversing the entire sedimentary sequence. Multiple detachments and different levels of folding and thrusting have developed within the basin. The shortening may also be driven by the gravitational energy contrast between the Central Burma Basin and the Shan Plateau in the east, caused by the extrusion of Tibetan material around the Eastern Syntaxis (see ‘Eastern Syntaxis to the Shan Plateau’ section below).

A number of shallow events reported by the GCMT catalogue (Ekström et al. 2005) suggest that these structures may still be active. GCMT centroid locations have been shown to have large uncertainties (up to c. 30 km, Elliott et al. 2010; up to c. 60 km, Weston et al. 2011), but at least two moderate earthquakes are likely to have occurred on structures other than the main mapped Sagaing Fault. The $M_w$ 6.6 2003 Taungdwingyi earthquake most likely occurred on a north–south-trending right-lateral strike-slip fault, parallel to the Sagaing Fault, to the west of the Pegu Yoma region (Fig. 2.2). Damage from this earthquake is reported in the town of Taungdwingyi (Thein et al. 2009), so is unlikely to have occurred to the east of the Pegu Yoma range. This earthquake suggests that some of the right-lateral shear is distributed on structures other than the Sagaing Fault (also see Maurin & Rangin 2009). There is also some evidence for active transpressive shortening with the 2007 $M_w$ 5.6 NW–SE-striking thrust earthquake that occurred on the western flank of the Pegu Yoma structure.

In the SW of Myanmar, the historical record includes a cluster of earthquakes in the period 1843–58 around the city of Pyay, 50–150 km west of the Sagaing Fault (Fig. 2.2; Le Dain et al. 1984). Eyewitness reports stated that the 1858 earthquake caused flow in the Irrawaddy to reverse in this region (Oldham 1883), and Wang et al. (2014) suggest that this may be because the earthquake uplifted an anticline that crosses the river in this area. Again this earthquake emphasizes the possibility of large events occurring away from the Sagaing Fault, in this case on a blind thrust beneath the anticline.
Northern Central Burma Basin

In northwestern Myanmar the Chindwin Basin shows evidence of thrusting between the Late Eocene and Early Miocene, as well as later thrusting with Eocene rocks thrust over the Irrawaddy Formation (Pivnik et al. 1998). The major thrusts in the Chindwin Basin trend NE–SW, closely parallel to the trend of the eastern margin of the Indo-Burman Ranges.
Assuming a NE–SW maximum horizontal stress direction for stresses associated with dextral strike-slip faulting on the north–south-striking Sagaing Fault, we would expect to see compressional structures trending NW–SE. The NE–SW-trending major thrusts in the Chindwin Basin are therefore at about the worst orientation for thrust reactivation associated with strike-slip deformation. These structures also lie at a high angle to the NNW–SSE-trending thrusts and faults in the Central Basin further south. It is therefore probable that the deformation in the Chindwin Basin is much less associated with strike-slip deformation than the Central Basin south of 23° N, and is more related to thrusting and folding in the Indo-Burman Ranges.

**Sagaing Fault**

The Sagaing Fault is over 1500 km long and runs from south of the Eastern Himalayan syntaxis in the north to the NE Andaman Sea in the south. It exhibits large-scale splay geometries in its northern and southern segments (Fig. 2.3). Onshore for about 600 km, the Sagaing Fault is a strikingly narrow, linear feature (Figs 2.2 and 2.3) and is clearly visible on satellite images. The fault zone curves gently into a more northerly oriented trace northwards. Locally along the linear 600 km long segment there are splays, small bends and small jogs in the fault trace, but these are comparatively minor features. There are no large-scale bends, giving rise to clear pull-apart or releasing bend geometries, possibly due to smoothing of the fault trace due to large displacement, as demonstrated by analogue modelling of stepping geometries in strike-slip faults (e.g. Dooley & McClay 1997).

The northern part of the Sagaing Fault displays strongly splaying segments with associated uplifted ranges (Fig. 2.3). The displacement on the fault zone appears to be dissipated in part into the Himalayan structures but also in part into local zones of uplift, folding and thrusting around the splaying fault region, which covers a distance from north to south of >400 km. A small number of NW–SE-striking thrust-faulting earthquakes are recorded by the Global Centroid Moment Tensor (GCMT) catalogue which likely reflects this distributed shortening, though these earthquakes have not been studied in detail. The southern margin of the Naga Hills may also play a role in absorbing this shortening. GPS data suggest that there is c. 12 mm a−1 of shortening in the ranges between the Hukawng Valley and Upper Kachin State (Fig. 2.2). Wang et al. (2014) did not find convincing evidence of recent rupture on the southern boundary of the Naga Hills, and there is little recent instrumental evidence for significant thrusting on this boundary. The International Seismological Centre (ISC) reports two major earthquakes in this area in 1906 and 1908 (M 7.0 and 7.5, respectively). It is not known whether these events represent strike-slip failure on the northernmost section of the Sagaing Fault, or reflect shortening at its termination. Building of the town of Naypyitaw (Fig. 2.3) and of various large construction projects, such as dams west of the Yangon–Mandalay Expressway, have created a whole series of new exposures in predominantly Late Miocene–Pleistocene sediments that reveal details of the structural style and timing of deformation adjacent to the Sagaing Fault. For a considerable distance along the highway, from about 17° 42′ 00″ N, the surface trace of the Sagaing Fault lies west of the highway. The trace is seen clearly as a prominent linear feature on satellite images, and commonly on the ground as a pressure ridge. Within the pressure ridge faulted sandstones and chaotic blocks, including sandstone blocks showing extensive cataclasite, can be found (Fig. 2.4). In some places recent sedimentation has covered the trace of the fault (e.g. south of Naypyitaw). The construction projects of Naypyitaw have cut into rolling hills, and revealed the deformation style around the Sagaing Fault as it affects the Irrawaddy, Obogon and Kyaukkok formations. Steep-splaying, north–south-trending strike-slip faults have offset and juxtaposed shale and sandstone-dominated units, and caused tight folding in places (Fig. 2.5). One outcrop revealed two angular unconformities that showed similar stages of deformation to that described by Pivnik et al. (1998) for the Salin sub-basin, which comprise an angular unconformity between the Late Miocene Obogon Formation and the Pliocene Irrawaddy Formation (Fig. 2.6). A second angular unconformity occurs within the fluvial sand-dominated Irrawaddy Formation (Fig. 2.6). In another outcrop in Naypyitaw, evidence for transmission of overpressured fluids was found where, adjacent to a small sub-vertical strike-slip fault, a sandstone rich in calcite cement has been intruded or injected into a sandstone bed (Fig. 2.7).

**Geodetic and geological slip rates**

GPS campaigns of sections of the Sagaing Fault have been conducted by Vigny et al. (2003) and Maurin et al. (2010). Maurin et al. (2010) occupied 12 GPS sites at the northern end of the fault in 2005 and 2008, and showed that slip is predominantly localized along a single fault trace (the Koma Fault) (Fig. 2.2). They also reoccupied earlier sites from Vigny et al. (2003). They found a largely consistent slip rate along the 400 km of fault north of Mandalay in the range of 18–22 mm a−1 (Fig. 2.2). Maurin et al. (2010) also noted that the locking depth of the Sagaing Fault, inferred from two-dimensional elastic modelling of GPS profiles, appears to decrease from c. 20 km in its central part to c. 6–8 km in the north. They acknowledge that the data scatter in central Myanmar means that a number of different models are possible, although it is clear that a relatively large locking depth is required.

Few other estimates of the slip rate of the Sagaing Fault have been made. Measurements on the southern segment of the Sagaing Fault from the offset of a sixteenth-century fortress wall, yielded a rate of 11–18 mm a−1 (Wang et al. 2011). Over a somewhat longer term, a slip rate estimate based upon offsets of the Singu basaltic flows (22.7° N) dated at 0.25–0.31 Ma (40K–39Ar) yields 10–23 mm a−1 of dextral motion (Bertrand et al. 1998). Based on these few measurements, the slip rate appears to be largely constant through time over the past few hundred thousand years and along-strike for almost 1000 km. The overall relative India-Sunda motion is 35 mm a−1 in a N10° direction at the latitude of Myanmar (Scoquet et al. 2006) and the whole country is caught between this relative right-lateral shear. The Sagaing Fault appears to accommodate just over half of this motion, and consequently represents the dominant active tectonic feature of the area.

**Instrumental and historical seismicity**

A number of large earthquakes occurred on the Sagaing Fault in the first half of the twentieth century (International Seismological Centre 2011) (Fig. 2.2). A pair of large earthquakes (each M w 7.3) occurred, seven months apart, at the southern onshore end of the fault in 1930 (International Seismological Centre 2011). Analysis of the tectonic geomorphology and historical accounts from the 120 km long fault zone (Tsutsumi & Sato 2009) indicate faulting through the deltaic lowland with fault scarps, pressure ridges and significant offsets of >3 m.

At the other end of the Sagaing Fault, a magnitude 7.6 earthquake ruptured the northernmost part of the fault in December 1931 (Fig. 2.2). Next, two earthquakes struck north of Mandalay in September 1946 on adjacent fault segments, three minutes apart (M s 7.5 and M s 7.8; Hurukawa & Maung Maung 2011). In this same area, three M w > 6 earthquakes occurred in 1991,
Finally, in 1956 the next segment failed in a $M_\text{w} 7.0$ event (International Seismological Centre 2011). In the pre-instrumental period there are descriptions of a very destructive earthquake in March 1839 (Le Dain et al. 1984), just south of Mandalay. The temporal clustering of large earthquakes on adjacent fault segments is commonly observed on large strike-slip faults (e.g. the North Anatolian Fault, Stein et al. 1997) and is thought to be caused by coseismic and postseismic stress changes associated with the first earthquake bringing adjacent portions of the fault closer to failure.

While the southern and northern parts of the Sagaing Fault appear to have failed in large earthquakes during the instrumental period, it is noticeable that there is a significant seismic gap on the Sagaing Fault. The >200 km long section running between the second and third largest cities in Myanmar, Mandalay and the new capital Naypyitaw has no clear instrumental or historical record of rupture. Wang et al. (2014) suggest that a large earthquake in 1839 may have ruptured all or part of this section, although this has not been confirmed. The lack of a record of large earthquakes on this segment suggests that it is either slipping aseismically, or that it fails in large earthquakes and has been accumulating elastic strain for some time. Wang et al. (2014) estimate that if this segment were to fail in a single event then it would be capable of producing a $M_\text{w} 7.8$–$7.9$ earthquake. The repeat time for such an event would be c. 500 years, which is longer than the reliable historical record in this region.

**Timing and total displacement**

Most of the folding and thrusting of the Central Basin between about 17° N and 22° N occurred during the deposition of the Irrawaddy Formation, which is of Plio-Pleistocene age according to Pivnik et al. (1998). This in turn suggests that activity of the Sagaing Fault occurred mainly during the Plio-Pleistocene which, by extrapolation of modern displacement rates, requires displacements no greater than 100 km (Bertrand & Rangin 2003). However, the age of the base of the Irrawaddy
Formation appears to be strongly diachronous (Plio-Pleistocene in some places further east in the basin, as suggested from the biostratigraphy in wells) and in some places as old as Middle Miocene (Chavasseau et al. 2006). Much more precision is therefore needed about which parts of the Irrawaddy Formation section are used to date structural events onshore.

Offshore to the south in Thailand waters along the projected trace of the Sagaing Fault there is a strike-slip fault zone of Early Miocene age, whose activity is clearly sealed by Middle Miocene sediments. Passing northwards, this fault zone could easily as be linked with the Shan Scarp. Bertrand & Rangin (2003) suggested that this Late Oligocene compressive structure may also have had a strike-slip component. Offshore the Sagaing Fault runs into the Gulf of Martaban, a region that during the Late Oligocene–Early Miocene was affected by extensional deformation. Subsequently, in some basins in the

---

**Fig. 2.5.** Road cut in NW suburbs of Naypyidaw, illustrating strike-slip faulting and folding affecting Miocene Kyaukkok and Obogon formations. Shale-prone sequences are abruptly juxtaposed by faulting with folded sandstones. Location: 19.773024° N, 96.048629° E.

**Fig. 2.6.** (a, b) Road cut in NW suburbs of Naypyidaw (same road, in a cutting west of Fig. 2.5), illustrating two unconformities related to transpressional deformation associated with the Sagaing Fault. Irrawaddy Formation (Pliocene or possibly Late Miocene) overlies steeply dipping Obogon Formation (Miocene) unconformably. An internal unconformity also exists within the Irrawaddy Formation. Location: 19.77087° N, 96.042968° E.
western part of the Gulf during the Middle–Late Miocene, considerable inversion occurred (perhaps related to subduction processes). The Sagaing Fault Zone, which is composed of two main strands (east and west) and a smaller central fault, enters the margin at the transition region between more east–west-trending rift structures to the west and the more north–south-oriented structures to the east. It is therefore difficult to determine whether the Sagaing Fault initiated the north–south-trending basins, or is just superimposed upon them. However, there appear to be three different stages of development present in the northern Gulf of Martaban: (1) Late Oligocene–Early Miocene synrift extension independent of the Sagaing Fault; (2) probable Late and possibly Middle Miocene transtensional expansion of section towards early-stage East and West Sagaing Fault strands; and (3) a Late Miocene or base Pliocene change in structural style that resulted in local uplift, folding and inversion along parts of the splaying fault zone. Regional subsidence continued during this time, and local folds are covered by Plio–Pleistocene deposition. The offshore data suggest there is a Miocene history to the Sagaing Fault Zone, at least along its southern portion. There was, however, a significant change in behaviour of the fault zone towards a more transtensional style around the end of the Miocene and the beginning of the Pliocene. Further offshore, however, the fault zone shows no strong indications of transpression and exhibits a major horse-tail-splaying style in a very thick (>6 km) Pliocene depocentre as the fault transfers displacement to structures in the central Andaman Sea.

The timing of displacement on the Sagaing Fault has proven controversial for three main reasons: (1) estimates of the amount of displacement (the greater the displacement, the older the fault needs to be); (2) extrapolation of recent displacement rates to the past; and (3) the timing of the change from extension to inversion in the Central Basin. In one model, over 400 km of displacement has been proposed for the Sagaing Fault (Mitchell 1993; Mitchell et al. 2012). Conversely, Bertrand & Rangin (2003) proposed that the cessation of subduction along the Andaman Trench around the Middle–Late Miocene boundary resulted in a change from oblique extension to strike-slip-dominated deformation in the Central Basin. The result was the formation of the Sagaing Fault and the extensive thrusting, folding, inversion and strike-slip deformation of the Central Basin. These authors suggest that total displacement on the Sagaing Fault is closer to 100 km.

Looking at the outcrop patterns in northern Myanmar, it is possible to correlate potential marker units across the fault (Fig. 2.8) and restore them. In particular, ophiolite belts and putative fragments of the Mogok Metamorphic belt can be correlated and restored (Fig. 2.8), which suggests a displacement in excess of 400 km on the Sagaing Fault (e.g. Mitchell 1993; Mitchell et al. 2012). There are several issues with this interpretation. (1) The current displacement rate on the fault is about 1.8 cm a\(^{-1}\) (Vigny et al. 2003; Maurin et al. 2010), hence considerably higher displacement rates (>4 cm a\(^{-1}\)) are required earlier in the fault history to achieve 400 km offset if the fault was initiated in the Late Miocene. With slower strain rates and 400 km offset, the history of fault activity must date back to the Early Miocene or Oligocene. (2) The details of how 400 km of Late Miocene–Recent displacement could be accommodated by folds and thrusts in the Himalayas are not well established. (3) The correlations of the offset ophiolites are not based on well-established piercing points; instead, they are based on subjective correlation of lithological units. In the simple restoration (i.e. progressively working westwards on different fault strands and restoring dextral offset) the Jade Mines Belt lies south of the join between the ophiolites (points 1a and 1b, Fig. 2.8), and hence still requires a second ophiolite trend to be present. Consequently the simple restoration does not resolve the ophiolites into a single trend. In addition, if the current geochemical and dating work is accepted, there are at least four ophiolites with different characteristics present in Myanmar and the Andaman Islands (Gardiner et al. 2015); a simple correlation of ophiolites as a single trend therefore cannot be considered reliable. The offset of the Neotethys suture through the Tagaung-Myitkyna Belt by the Sagaing Fault (Mitchell 1993; Mitchell et al. 2012) is an attractive interpretation which simplifies some of the ophiolite relationships in the north.

Another candidate for an offset marker may be found in the major rivers of this region. It has been suggested that the Irrawaddy River once flowed along the course of the present-day Chindwin River, and that movement on the Sagaing Fault led the Irrawaddy to abandon the course of the Upper Chindwin River and instead to join the central section of the Chindwin...
This implies c. 360 km of total cumulative displacement; this offset is, however, also speculative. The Irrawaddy could have been diverted onto a more easterly path due to uplift associated with the northern termination of the Sagaing Fault, rather than by simple right lateral displacement. A detailed fluvial sediment provenance study in the upper reaches of the Chindwin River or thermochronological study of the ranges associated with the termination of the Sagaing Fault could potentially resolve this issue.

In later sections we note that rapid right-lateral strike-slip faulting in the Indo-Burman Ranges appears to either be very young or to have recently accelerated. The onset of faulting in the Shillong Plateau (described in ‘Shillong Plateau’ section below) may also have allowed right-lateral shear to be taken up through the rotation of the Brahmaputra Valley. It is therefore possible that the Sagaing Fault did take up a larger component of the India–Sunda shear in the past, and that a proportion of this shear has now shifted to structures in the west. This would still require at least 10 Ma to accumulate 360 km of cumulative displacement, indicating a Middle Miocene or earlier initiation of the fault.

(Maung 1987) (Fig. 2.2). This implies c. 360 km of total cumulative displacement; this offset is, however, also speculative. The Irrawaddy could have been diverted onto a more easterly path due to uplift associated with the northern termination of the Sagaing Fault, rather than by simple right lateral displacement. A detailed fluvial sediment provenance study in the upper reaches of the Chindwin River or thermochronological study of the ranges associated with the termination of the Sagaing Fault could potentially resolve this issue.

In later sections we note that rapid right-lateral strike-slip faulting in the Indo-Burman Ranges appears to either be very young or to have recently accelerated. The onset of faulting in the Shillong Plateau (described in ‘Shillong Plateau’ section below) may also have allowed right-lateral shear to be taken up through the rotation of the Brahmaputra Valley. It is therefore possible that the Sagaing Fault did take up a larger component of the India–Sunda shear in the past, and that a proportion of this shear has now shifted to structures in the west. This would still require at least 10 Ma to accumulate 360 km of cumulative displacement, indicating a Middle Miocene or earlier initiation of the fault.

Eastern Syntaxis to the Shan Plateau

Kinematics

The SE margin of the Tibetan Plateau is marked by a complex network of faults. Researchers have found that the style of deformation varies dramatically both temporally and spatially. This variation has made it a proving ground for various models of distributed continental deformation (Searle & Morley 2011; Searle et al. 2011). The interpretation of the dynamics of this region remains a topic of active debate. Earthquake focal mechanisms, mapped active structures and GPS vectors are summarized in Figure 2.9 to illustrate the current kinematics of the zone.

In the north, strike-slip faults such as the right-lateral Gaoligong Fault in the west and the left-lateral Xianshuihe Fault in the east curve around the Eastern Syntaxis, translating material from the interior of the Tibetan Plateau SE between Myanmar and the Sichuan Basin. These fault systems were mapped by Wu (1991) and Wang & Burchfiel (1997). Major earthquakes have occurred on the Xianshui-he Fault including the

![Fig. 2.8. (a) Simplified geological map modified from Mitchell et al. (2012). Possible pinning points for restoration are numbered. (b) Restoration of the map in (a) with about 400 km displacement, showing how the Tagaung–Myitkina Belt ophiolites restore as a continuation of the Neotethys, Western Belt ophiolites. In this scenario the location of the Jade Mines Belt ophiolites remains problematic.](image-url)
South of 28° N the Xianshui-he splits into a number of separate c. north–south-trending strands. The southern end of this fault system ends at the NW–SE-striking Red River Fault, and a number of short right-lateral strike-slip faults parallel to the Red River Fault (Wang et al. 1998). One of these short right-lateral faults (the Qujiang Fault) was the source of the $M_w$ 7.3 1970 Tonghai earthquake (Zhou et al. 1983). The Gaoligong Fault represents a long-lived shear zone that has significant but unquantified geological displacements. It has been suggested that this fault once formed the main boundary between the Indian Plate and Southeast Asia (Wang & Burchfiel 1997). Despite this, there has been little instrumentally recorded seismicity. Right-lateral offsets of stream courses suggest the fault is still active (Wu 1991). The 1979 Lung-Ling earthquake sequence occurred where the Gaoligong Fault meets the Nanding Fault in the south, and one event in this sequence has been attributed to right-lateral failure on the southern section of the Gaoligong Fault (Holt et al. 1991). Between the Gaoligong and Xianshui-he strike-slip systems, earthquake focal mechanisms indicate extensional faulting. In the north, east–west-aligned normal faulting earthquakes indicate north–south-aligned extension. The associated faults are not well expressed in the landscape, perhaps masked by rapid erosion. In the south (<27° N) there are north–south-directed graben with associating normal faulting earthquakes and east–west-aligned extension (Copley 2008).

South of 25° N the tectonic regime changes again. In the west deformation is dominated by NE–SW curved left-lateral strike-slip faults such as the Wanding, Nanting and Mengliang faults. In the east there is a series of NW–SE right-lateral strike-slip faults, most prominently the Red River Fault. The left-lateral faults in the west are generally thought to be accommodating north–south-aligned right-lateral shear through clockwise rotations about a vertical axis. This interpretation is consistent with both geological observations (Wang & Burchfiel 1997) and the GPS velocity field (Shen et al. 2005; Copley 2008). Clockwise rotation would result in north–south shortening and east–west extension.

There are a number of unusual hair-pin river bends associated with the NE–SW left-lateral strike-slip faults, including the Wanding, Menglian, Menxing and Nam Ma faults (Lacassin et al. 1998), the latter most recently rupturing in the 2011 $M_w$ 6.8 earthquake (Feng et al. 2013). These faults are currently left-lateral strike-slip faults, associated with 5–24 km cumulative displacements. Once these recent offsets have been restored, 50–60 km right-lateral displacements are then needed to return the rivers to their original straight course (Lacassin et al. 1998) (Fig. 2.10). This implies that these deeply incised rivers represent extremely persistent geomorphological markers that record a long-lived and complicated history of deformation, including an inversion of the sense of slip. The timing of this inversion is not well constrained. Lacassin et al. (1998) suggest that it probably occurred during 20–5 Ma, primarily by comparison with similar inversions on other nearby faults such as the Red River Fault. Just south of the Myanmar border in Northern Thailand lies the ENE–WSW-trending Mae Chan Fault. This fault connects with the north–south-trending fault system to the Oligocene–Miocene Fang Basin. In order to open the basin, the Mae Chan Fault must have been active with left-lateral motion (Morley 2007). Significant right-lateral motion would have closed and inverted the basin. Consequently, passing south it seems probable that the ENE–WSW-trending faults have experienced different stress histories from those in the north, and the observations made by Lacassin et al. (1998) do not necessarily apply to all the faults in the system, particularly those in the south.

In the east the Red River Shear Zone (RRSZ) is also thought to have undergone a slip inversion. This fault, and the associated Ailao Shan Shear Zone, has been identified as a major Cenozoic left-lateral shear zone, interpreted as accommodating the large-scale (c. 1000 km) extrusion of Tibetan and SE Asian material out of the path of the Indian indenter (e.g. Tappin et al. 1986, 1990; Leloup et al. 1993, 1995, 2001). Searle (2006) and Searle et al. (2010) questioned the geological features used in these reconstructions, and suggested that none of them provide precise pinning points to determine offsets accurately. These authors further showed that all the metamorphism, and most of the fabrics in the RRSZ in the Ailao Shan, Diancang Shan and the Day Nui Con Voi Massif in Vietnam, occurred prior to strike-slip shearing. There have been no large earthquakes on the RRSZ for at least 300 years, but geomorphological evidence, such as stream and
river offsets, and displaced Quaternary river terrace gravels, indicate that it is currently active and has a right-lateral sense of motion (Allen et al. 1984). Maximum right-lateral offset estimates range from 5.5 km (Allen et al. 1984) to c. 40 km (with 16 km of offset since the Pliocene) (Schoenbohm et al. 2006a).

**Fig. 2.10.** (a) Active fault map and hill-shaded DEM of Indo-China, highlighting the series of left-lateral strike-slip faults between the major right-lateral Saigon and Red River faults. Inset boxes indicate location of river offsets. Faults are from Taylor & Yin (2009). (b) Landsat false colour mosaic (RGB 752) of the Nam Ma and Mengxing Faults. The Nam Ma Fault has a current 11 km left-lateral displacement in the river, consistent with recent earthquake focal mechanisms, but overall has c. 31 km of offset in a dextral sense (Lacassin et al. 1998). (c) Landsat false colour mosaic (RGB 752) of the Wanding Fault with a current 10 km left-lateral offset of the Salween River, but overall a right-lateral offset of 33–54 km (Lacassin et al. 1998).
The timing and amount of deformation on the RRSZ are controversial. Leucogranites associated with the shear zone are thought by some to be syntectonic, created due to shear-heating associated with left-lateral motion on the fault (e.g. Leloup et al. 2007). Searle (2006) and Searle et al. (2010) argued that the magmatism was predominantly pre-strike-slip faulting. Monazite

Searle (2006) and Searle et al. (2010) argued that the reactivation as a right-lateral shear zone did not occur during the Late Oligocene. Subsequent dextral motion on a north–south-trending fault: the Panluang, Paungluang or Shan Scarp Fault (e.g. Garson et al. 1976; Bertrand et al. 2007). The Three Pagodas Fault is known as the Moulmein Fault in Myanmar. Important NW–SE-trending strands of the Mae Ping Fault Zone in Myanmar are the Papun and Toungoo (or Patun) faults. These faults form an extensive network, and are not just single discrete major faults. In the part of Myanmar to the west of Thailand, these faults generally trend NW–SE but tend to spl interconnect or merge with other fault segments. Strike-slip duplex geometries are common (Morley 2004; Morley et al. 2007). Further north the NNW–SSE trends become dominant. On the western side of the Shan Plateau is a major NNW–SSE-trending fault: the Panlaung, Paungluang or Shan Scarp Fault (e.g. Garson et al. 1976; Bertrand et al. 2001; Morley 2004). It separates the Slate Belt to the west from the Paungluang-Mawchi zone to the east (e.g. Mitchell et al. 2004). The Shan Scarp Fault links up with strands of the Mae Ping Fault Zone. Another NNW–SSE fault, branching off the Mae Ping Fault, is the Kyaukkyan Fault. This segmented fault zone is over 600 km long, has been recently active (Mw 7.7 earthquake in 1912), and has the potential for a large-magnitude earthquake. The total offset of the segments (Wang et al. 2014). Ridd & Watkinson (2013) speculated that large dextral motion on a north–south-trending Panluang-Ranong Fault Zone occurred in the Late Cretaceous–Palaeogene. Better established is that the largest displacement on the Three Pagodas and Mae Ping faults occurred during the Eocene and Early Oligocene (prior to 30 Ma), and was left-lateral (Lacassin et al. 1997). There is good evidence from basins at releasing bend geometries that subsequently, during Late Oligocene–Recent times, there was highly episodic right-lateral motion (Morley 2002, 2004; Morley & Racey 2011). The most continuous and largest dextral displacements seem to have occurred on the western side of the Shan Plateau. There, Bertrand et al. (2001) identified major right-lateral transtensional unroofing around the Shan Scarp during the Late Oligocene–Early Miocene. Subsequent dextral activity appears to have migrated to the adjacent Sayaing Fault to the west, and to a much lesser extent, the Kyaukkyan Fault to the east.

**Dynamics**

The dynamics of deformation within regions with distributed faulting are often explained using models that describe the lithosphere as a fluid, in which deformation is controlled by gradients in gravitational potential energy (England & McKenzie 1982). This framework, developed for Tibet, can successfully reproduce the style of deformation and long-wavelength GPS-derived velocity fields in many actively deforming areas when appropriate boundary conditions are applied (e.g. Ford et al. 2010).

The gradual topographic slope extending from the plateau to the lowlands of Thailand in the south contrasts sharply with the steep range front separating the plateau from India in the west (the Himalaya) and the Sichuan Basin (Longmen Shan) in the east. Workers have therefore sought physical explanations for this variation in plateau margin morphology. Clark & Royden (2000) influentially suggested that deformation could be described by the gravitationally driven dextral flow of lower crustal Tibetan material, within a low-viscosity channel between rigid upper crust and upper mantle layers. If the viscosity of the lower crust within India and the Sichuan Basin is taken to be relatively high, it would then resist and laterally divert Tibetan lower crustal flow, resulting in the passive uplift of the upper crust and the development of a steep plateau margin. The gentle gradient of the SE margin of the plateau between India and the Sichuan Basin can then be explained if the lower crust has a lower viscosity, allowing Tibetan lower crustal material to flow outwards from the plateau, escaping hundreds of kilometres to the south (Clark & Royden 2000; Clark et al. 2005a). The lower-crustal-flow model is distinct from the Miocene mid-crustal channel-flow model in the Himalaya (Searle et al. 2011). In the Himalayan channel-flow model, the low-viscosity channel of Indian Plate metamorphic rocks, migmatites and leucogranites is bounded by bittile upper crust above, and underthrust rigid Indian lower crust below. Consequently, rather than shortening occurring entirely within a lower crustal layer, hidden beneath a rigid upper crust, the Himalayan channel flow model results in the exhumation of material which formed part of the channel at the range front. Copley & McKenzie (2007) and Copley (2008) showed that the horizontal GPS velocities (Shen et al. 2005) and topography in the region between the Sichuan Basin and the central Myanmar lowlands are consistent with a slightly modified lower-crustal flow hypothesis. They also demonstrated that these observations can be explained by a model (resembling that of England & McKenzie 1982) in which deformation takes the form of pure shear, with material flowing out over a stress-free lower boundary. In contrast to the lower-crustal-flow model, this model requires the entire vertical thickness of the lithosphere to be shortening or extending at the same rate. Copley & McKenzie (2007) suggested that the crust of the Indian Plate and Sichuan Basin was much stronger; they therefore modelled deformation at these margins as a gravity current of weaker Tibetan material, flowing out over a rigid deformable base. With a suitable choice of rheology, this then results in the steeper range fronts observed. This difference in boundary condition (a rigid deformable base representing strong Indian and Sichuan material, or a stress-free lower boundary resulting in pure shear in southeastern Tibet) can produce the contrasting topographic characteristics of these margins.

At the southeastern margin of Tibet, both models represent the Sichuan Basin and central Myanmar lowlands as relatively high-viscosity regions, acting as ‘gate posts’ bounding the southwards flow of Tibetan material. This causes southwards velocity to decrease towards these regions, imposing an n-shaped velocity profile on material flowing out of the central plateau. Both models involve the flow of material in the direction of topographic slope, with the rate of flow being controlled by the steepness of the slope. This property predicts both north–south extension in the north where the plateau begins to slope towards the SE and east–west extension after Tibetan material has passed between the Sichuan Basin and the Eastern Synclisis, allowing it to flow outwards to the east and west. The westward
flow may drive bending of the Sagaing Fault, and uplift and compression in the Central Burma Basin and Indo-Burman Ranges in the west (Rangin et al. 2013). Consequently, both models account for the style of faulting found in this region. The two models (lower crustal flow and pure shear) represent end-members of a range of possible solutions, with differing amounts of coupling between the lower and upper crust. They differ principally in the presence or absence of shortening within the upper crust of the Tibetan margins (Copley & McKenzie 2007).

Tomographic studies of the region reveal that there is a low-shear-wave-velocity layer within the mid-lower crust of the plateau (e.g. Liu et al. 2014). There is also evidence for low electrical resistivity at these depths (Bai et al. 2010; Zhao et al. 2012). The orientation of azimuthal anisotropy also varies between the crust and upper mantle in SE Tibet, suggesting that these two layers may be decoupled (e.g. Yao et al. 2010). These observations all support the existence of a relatively weak layer in the mid-lower crust, perhaps containing partial melt, although the extent to which it is interconnected and able to act as a channel remains unclear. The fact that areas of low velocity in the lower crust appear to be truncated by major strike-slip faults, such as the Xianshui-he Fault, suggests that deformation in the lower crust is in some way connected to brittle deformation in the upper crust (Yao et al. 2010; Liu et al. 2014).

**Past deformation**

A variety of low-temperature thermochronometry studies report the onset of rapid river incision in SE Tibet between the Sichuan Basin and the Eastern Syntaxis during the Miocene. The timing is generally found to be c. 13–9 Ma (e.g. Clark et al. 2005b; Ouimet et al. 2010); however, older ages have been reported at some localities (e.g. 15–22 Ma near Daoheng, Tian et al. 2014). The more recent increase in incision is typically interpreted as the onset or acceleration of uplift in this area; however, climatic variation, or river catchment reorganization, could also play a role. It has been suggested that the earlier dates reflect the true onset of regional uplift, while the younger increase in exhumation may reflect the intensification of the monsoon around this time (Tian et al. 2014). The uplift of this region, driven by the gravitationally driven lateral escape of Tibetan material from the India–Eurasia collision, may have caused the tectonic reorganization described above. A number of authors have speculated on how this change in the distribution of gravitational energy may have affected deformation in this region, including Schoenbohm et al. (2006b) who linked the southeastern propagation of uplift to the waning influence of the left-lateral Red River–Ailao Shan Shear Zone. These suggestions certainly seem plausible, and as more constraints on the timing of deformation become available it may be possible to quantitatively explore how the system has evolved.

Lithospheric thickness maps, derived from surface wave tomography, indicate that the thick lithosphere underlying the Tibetan Plateau extends to c. 25–27° N in this area (Priestley & McKenzie 2013) (Fig. 2.11). Interestingly, this coincides with an abrupt change in seismic anisotropy from c. north–south to the north of 26° N to c. east–west in the south (e.g. Flesch et al. 2005; Sol et al. 2007). It should be remembered that although the lithospheric thickness map produced by Priestley & McKenzie (2013) is calculated on a 2° by 2° grid, it is derived from surface wave tomography with c. 250 km lateral resolution, which can produce smearing of sharp boundaries in lithospheric thickness. North of 26° N directions of seismic anisotropy parallel GPS vectors and fault orientations, whereas south of 26° N deformation is dominated by the NE–SW left-lateral strike-slip faults, cutting across the seismic anisotropy directions at a steep angle. In addition, the low-shear-velocity mid-crustal layer is present in the north but not in the south (Agius & Lebedev 2014). This abrupt change in lithospheric thickness, seismic anisotropy and style of faulting may not be a coincidence. One possibility is that north of 26° N long-term deformation has been characterized by the large-scale bulk southwards flow of Tibetan material, driven by gravitational stresses associated with gradients in the topography of the surface and the lithosphere–asthenosphere boundary. The anisotropy in the north would therefore reflect crystallographic alignment associated...
with this bulk flow, whereas to the south it may reflect other processes which are potentially (given that they do not agree well with the orientations of GPS vectors or active faults) no longer active. The presence of thick lithosphere beneath areas thought to have been uplifted in the last c. 10 Ma also suggests that thickening in this area has not just occurred within the mid-lower crust, as the strictest interpretation of the channel flow hypothesis would imply.

Northern India, Bangladesh and the Indo-Burman Ranges

Shillong Plateau

The Shillong Plateau stands over a kilometre above the surrounding sediments of the Ganges–Brahmaputra Delta (Fig. 2.12). The plateau and its surroundings are the most seismically active part of the Indian Shield. Small-to-moderate earthquakes are commonly detected in this region, and great historical earthquakes are thought to have occurred on faults bounding the plateau. The great Assam earthquake of 1897, which caused damage across northeastern India (including Dhaka), has been associated with a blind thrust fault on the northern edge of the Plateau (Bilham & England 2001). A reanalysis of seismograms suggests that this earthquake had a surface-wave magnitude of $M_s$ 8.0 ± 0.1 (Ambraseys 2000). Bilham & England (2001) used trigonometric surveys from the Indian Survey (Bond 1899; Wilson 1939) to infer that the earthquake occurred on a south-dipping reverse fault on the northern side of the Shillong Plateau. They named this previously unidentified structure the Oldham Fault.

The inferred coseismic model is remarkable for a number of reasons. The fault slip (25 ± 5 m) exceeds that normally observed in intraplate continental earthquakes, as does the maximum depth of rupture (>35 km, compared to 15–20 km in most actively deforming regions). The unusual length, amount of slip and associated high stress drop are probably the consequence of the unusually large seismogenic thickness (extending throughout the crust), which is consistent with the geologically inferred maximum rupture depth and the depth of modern earthquakes observed within this area and across the northern Indian Shield (Mitra et al. 2005; Lasitha 2007; Sloan et al. 2011; Kumar et al. 2015). These characteristics allow the occurrence of unusually large intraplate events ($M_s$ 8+). Similar rupture characteristics have been observed in northwestern India in the 2001 Bhuj earthquake and inferred from prehistoric surface ruptures on the Tapti Fault (Copley et al. 2011, 2014). The fact that the entire crust appears to be seismogenic in these regions is commonly attributed to unusual lower-crustal compositions, perhaps due to the absence of small amounts of hydrogen ions dissolved in the crystal lattice of nominally anhydrous minerals, which can greatly increase the strength of rocks (e.g. Jackson et al. 2008). Such earthquakes also have a spatial correlation with areas of thick stable lithosphere (Sloan et al. 2011).

The Shillong Plateau is one of the few compressional structures bounded by faults capable of lower crustal seismicity that are moving at geotectonically appreciable rates. Recent lithospheric thickness maps obtained from surface-wave tomography (Priestley & McKenzie 2013) suggest that both the Bhuj and the Shillong Plateau earthquakes occurred near the edge of the region of thick lithosphere, perhaps explaining why these areas are deforming comparatively quickly. This rationale was also used to explain the location of an upper mantle earthquake beneath the Arafura Sea (Sloan & Jackson 2012). The presence of an unusually thick seismogenic layer, and the potential for extremely large intraplate earthquakes south of the Shillong Plateau, has important implications for seismic hazard.

Most previous workers placed the 1897 earthquake on a north-dipping thrust fault on the southern side of the Shillong Plateau. In contrast to the north, the southern side of the plateau is bounded by the well-known active Dauki Fault, which has a very clear geomorphic expression. Despite the geodetic evidence presented by Bilham & England (2001), some authors have continued to question their interpretation, favouring the Dauki Fault or an alternative south-dipping fault projecting to the surface further north (e.g. Rajendran et al. 2004; Morino...
et al. 2011). None of the proposed alternatives can explain the geodetic constraints (Bilham & England 2001; England & Bilham 2015). Additionally, the observation of 11 km of down-to-the-west coseismic deformation on the Chedrang Fault, near the NW of the plateau (Oldham 1899), is consistent with rupture in the region of the proposed Oldham Fault if this slip was driven by coseismic elastic stresses acting on a frictionless plate (Bilham & England 2001). It is not clear how such displacement could be driven by failure of the other proposed fault planes. It may be surprising that such a large earthquake did not produce clear surface ruptures, but it is not unprecedented (e.g. the $M_w$ 7.6 Bhuj earthquake, Copley et al. 2011 and the $M_w$ 7.8 Gorka earthquake, Elliott et al. 2016) and the geodetic data show that the rupture stopped 9 km below the surface. Clark & Bilham (2008) observed that river slopes steepened above the inferred fault, but acknowledged that the total cumulative slip on this fault appears to be minor relative to the Dauki Fault in the south. A sharp increase in incision is also observed immediately south of the surface projection of the Oldham Fault further confirming that this structure exists and is active (England & Bilham 2015).

The receiver function analysis of Mitra et al. (2005) revealed that there is a sharp step in Moho topography associated with the southern margin of the Shillong Plateau. The Moho is c. 6 km shallower beneath the plateau than beneath the Indian Shield to the south, which lies at a depth of c. 42 km. A similar deepening of the Moho is also seen to the north (Mitra et al. 2005; Borah et al. 2016). The abruptness of this step, over a distance of c. 50 km, is further evidence that the lower crust in this region is relatively strong. If it were not capable of supporting large stresses over geological time, the lower crust would have undergone flow under gravity, smoothing out the topographic step.

A recent GPS survey of NE India has shown that the horizontal shortening rate across the Shillong Plateau is 3–6 mm a$^{-1}$, increasing to the east (Vernant et al. 2014). This increase is mirrored by a decrease in shortening rate across the Himalaya (Vernant et al. 2014). Geological uplift rates have been estimated at 0.7–1.4 mm a$^{-1}$ in the eastern part of the fault (Clark & Bilham 2008), where the GPS constraints suggest a shortening rate of 5–7 mm a$^{-1}$ (Vernant et al. 2014). These values are irreconcilable, unless (1) the Dauki Fault has an exceptionally low dip, in which case it would intersect the Oldham Fault in a physically unrealistic way (Bilham & England 2001), and contradict the recent receiver function analysis that suggests its dip to be c. 30$^\circ$ (Singh et al. 2016); (2) the geodetic rates are increased relative to the long-term shortening rate due to long-lasting post-seismic deformation associated with the 1897 earthquake; or (3) the shortening rate has increased dramatically over the plateau’s history, perhaps because of the evolving stress state as the Plateau approached the Himalayan Arc (Najman et al. 2016). Clark & Bilham (2008) have proposed a structural model in which the structure mapped as the Dauki Fault is in fact a fold axis, propagating above a north-dipping master fault at depth. This model still results in a contradiction between long-term geological uplift rates and the most recent GPS measurements (Vernant et al. 2014).

The timing of the onset of surface uplift has only recently been revealed. A marine transgression at the Oligocene–Miocene boundary has been linked to the onset of flexure, associated with uplift on the Dauki Fault (Alam et al. 2003), but the associated sediments do not thicken towards the plateau (Johnson & Alam 1991). The fluvial Tipam Group, deposited between 3.5 and c. 2 Ma, does thicken towards the Shillong Plateau (Johnson & Alam 1991; Najman et al. 2012, 2016) and likely indicates the onset of surface uplift (Najman et al. 2016).

Thermochronological data have been interpreted as indicating an increase in exhumation at 15–8 Ma (Biswas et al. 2007; Clark & Bilham 2008). This coincides approximately with a dramatic slowing of exhumation within the Bhutan Himalaya to the north, during 10–5 Ma (McQuarrie et al. 2014) indicating a major southward shift in the accommodation of India–Eurasia convergence. The contrast between the onset of exhumation and the beginning of surface uplift suggests that erosion of Cenozoic sediments initially kept pace with rock uplift, and surface uplift only occurred once the Precambrian basement was exposed (Biswas et al. 2007). Together these indicate a major southwards shift in the accommodation of India–Eurasia convergence.

Triputra Fold Belt

The Triputra Fold Belt (TFB, sometimes referred to as the Outer Indo-Burman Ranges) is a fold-and-thrust belt consisting of a series of arcuate anticlines running for 400 km between the coast and the eastern Shillong Plateau. Seismic reflection studies show that these anticlines are underlain by imbricate thrust faults, soling into a low-angle detachment. Seismic and borehole data suggest that this detachment is an under-compacted over-pressured shale layer within the sediment pile (e.g. Lohmann 1995; Sidker & Alam 2003). Available seismic data suggest that this detachment occurs at c. 4 s two-way travel-time, corresponding to c. 5 km depth (Sidker & Alam 2003).

Most of the instrumentally recorded seismicity below the TFB occurs within the down-going Indian Plate (Chen & Molnar 1990; Copley & McKenzie 2007; Stork et al. 2008; Kumar et al. 2015) (Figs 2.12 and 2.13). There is some evidence for earthquakes occurring within the overlying sediments. The 2003 $M_w$ 5.6 Burkhol earthquake is reported to have produced surface ruptures (Steckler et al. 2008), suggesting a shallow depth. The 1999 $M_w$ 5.2 Maheshkhali earthquake was also associated with extensional fracturing on the crest of an anticline (Ansary et al. 2000), potentially indicating the rupture of a shallow thrust beneath the structure (Steckler et al. 2008). Finally, a 1997 $M_w$ 5.2 earthquake that occurred near the India–Myanmar border had a well-determined 5 km centroid depth (Copley & McKenzie 2007; Stork et al. 2008). This earthquake appears to be a low-angle north–south-striking thrust, and could have occurred on a subhorizontal overpressured detachment within the sediment pile, similar to those imaged in seismic reflection profiles. A small number of other low-angle thrusts are listed in the GCMT catalogue, but these earthquakes lack good depth control. All of these events are thrusts, taking up shortening perpendicular to the local trend of the Indo-Burman Ranges. They have very different mechanisms from events within the underlying plate, which tend to have steep focal planes and north–south to NNE–SSW P-axes (Chen & Molnar 1990). The orientations of these focal planes are inconsistent with the orientation of the anticlines in the TFB (Chen & Molnar 1990; Baruah et al. 2013). This suggests that these anticlines are either currently inactive, or that there is a pervasive detachment within the area (Chen & Molnar 1990). A small number of earthquake mechanisms within the shallow sedimentary pile are consistent with the orientation of the anticlines, and is further evidence that these structures are active and that a detachment exists between these levels. The P-axes within the basement match the shortening direction expected from the northwards movement of the Indian Plate. The east–west shortening within the shallow sedimentary layers may be related to the gravitational collapse of the high topography of the Indo-Burman Ranges to the east (Copley & McKenzie 2007).

Folding in the western part of the range appears to be very recent, developed in the last 2 Ma, and propagating westwards at c. 10 cm a$^{-1}$ (Maurin & Rangin 2009; Najman et al. 2012). Shortening reconstructions based on the seismic profiles show
that 11 km of shortening has been taken up by thin-skinned deformation within the TFB. This corresponds to 5.5 mm a\(^{-1}\) if deformation is taken to have begun at 2 Ma (Maurin & Rangin 2009). Recent GPS results suggest there is c. 10 mm a\(^{-1}\) of range-perpendicular shortening in the northern TFB. This deformation occurs at a high angle to the direction of India–Sunda relative plate motion (Gahalaut et al. 2013) (Figs 2.12 and 2.13). The recent onset and rapid propagation of folding requires an explanation. If deformation reflects gravitationally driven movement on a weak (effectively stress-free) low-angle detachment (e.g. Rangin et al. 2013), then disequilibrium compaction from rapid sediment burial may have resulted in high overpressures (Zahid & Uddin 2005), which facilitated movement on the detachment. The onset of folding might mark the time when the build-up of overpressure was sufficient to permit sliding on the detachment.

There also appears to be evidence in the GPS data of c. 10 mm a\(^{-1}\) of right-lateral shear in the direction of plate motion accommodated across this area. Maurin & Rangin (2009) identified the Chittagong Coastal, Kaladan, Lelon and Kabaw faults as the main structures that could be accommodating the oblique dextral motion (Fig. 2.14), although other, as yet unidentified, faults are almost certainly present.

While there have been very few earthquakes within the shallow sedimentary pile in the instrumental record, it seems likely that some small-to-moderate earthquakes do occur in the upper 5–10 km of the sedimentary layer, and it is appropriate to consider what earthquake hazard is associated with these shallow structures. The casualties and significant structural damage caused by the relatively small \(M_w 5.6\) 2003 Barkhol earthquake emphasizes that earthquakes within the wedge overlying the plate interface still pose a significant risk in a country with an exceptionally vulnerable population.

The possible presence of a pervasive detachment within the sediment pile has important implications for seismic hazard. A recent assessment of seismic hazard in this area used the lengths of the anticlines as proxies for fault length, and then used empirical fault-length–magnitude relationships to infer the possible size of earthquakes in the region (Wang et al. 2014). Many of these anticlines are 40–100 km long; if this scaling holds true, these are considered to be associated with potential earthquakes with magnitudes of up to \(M_w 7.7\).
Fig. 2.14. Regional tectonic setting of the Andaman Sea Region modified from Morley (2017). See text for explanation of labels A–E. The locations of Figures 2.15– 2.17 are indicated.
Comparison with earthquakes in similar tectonic settings suggests that such large earthquakes would cause devastation over large areas of NE India and Bangladesh. These $M_w > 7$ potential magnitudes may overestimate the earthquake hazard associated with these particular structures (although earthquakes of this magnitude do occur in the area).

In the Zagros Mountains of southern Iran we see a similar fold style associated with a thick sedimentary succession, with low-angle detachments in salt and shale layers (Nissen et al. 2011; Elliott et al. 2015). In the Zagros, weak sedimentary layers at 10–15 km depth appear to limit the downdip extent of rupture. This then restricts the size of most earthquakes in the region to $M_w < 6$. In a few specific locations, where highly asymmetric anticlines expose lower Mesozoic/Palaeozoic strata, faults cross this downdip barrier (perhaps because the detachment layer is absent). In these areas faults cut through both cover and basement down to c. 20 km and earthquakes of magnitude up to $M_w 6.7$ occur (Nissen et al. 2011).

In the Zagros Mountains anticlines commonly have lengths of more than 70–110 km, but there is no evidence that this leads to $M_w$ c. 8 earthquakes. Ramsey et al. (2008) analysed modern and relict drainage patterns to infer that the long anticlines are actually composed of many small segments that have coalesced. Uplift patterns derived from InSAR studies indicate that there is no simple direct relationship between long-term anticlinal uplift and the pattern of uplift associated with individual earthquakes (e.g. Nissen et al. 2007, 2010; Roustaie et al. 2010). Nissen et al. (2011) suggest that the presence of weak layers within the sedimentary succession plays an important role in limiting the size of earthquake rupture.

A potential counter-analogy is the 1999 $M_w$ 7.6 Chi-Chi earthquake in Taiwan (Rubinstein & Hutton 1994). Taiwan lacks the concentric pattern of short-wavelength relatively symmetric anticlines found in the Zagros and the TFB; however, a pervasive very shallowly dipping detachment at c. 6 km has been identified on seismic reflection data which did not prevent the occurrence of a large shallow earthquake. The hypocentre, aftershock depths and coseismic GPS data all suggest that the Chi-Chi earthquake initiated at c. 12 km depth, below the shallow detachment, and either ruptured through the detachment (Kao & Chen 2000) or involved failure with a more complex geometry within the shallow detachment itself (e.g. Yue et al. 2005). Regardless of which interpretation is correct, the example of the Chi-Chi earthquake emphasizes that very large shallow earthquakes are possible in areas with a pervasive near-horizontal detachment if the detachment itself is capable of seismogenic failure, or if rupture propagates through the detachment.

Large earthquakes do occur beneath the TFB. The 1918 Srimangal earthquake is reported to have been $M_w$ c. 7.1–7.5. Macroseismic reports suggest that this earthquake occurred close to the Rashidpur Anticline (Stuart 1920) (Fig. 2.12 inset), and Wang et al. (2014) suggest that the event ruptured a fault associated with the anticline. If correct, this attribution would call into question the analysis above. It seems unlikely that this earthquake did occur on a structure corresponding to the Rashidpur Anticline. Macroseismic studies produced at the time (Stuart 1920) and a detailed analysis of historical records (Ambroseys & Douglas 2004) both show isoseismals elongated in an east–west direction, whereas the Rashidpur Anticline axis strikes north–south. The largest earthquake within the TFB in the recent instrumental period ($M_w 5.9$, 8th May 1997) is a strike-slip earthquake that occurred at 30 km depth and had a WSW–ENE nodal plane (Mitra et al. 2005).

Stuart (1920) also reports on a levelling line from Silchar to Comilla, which was reoccupied after the 1918 earthquake. Comparison with earlier occupation showed no sign of uplift in the epicentral area, even where it crossed the Rashidpur Anticline. Instead they found a c. 45 km wide zone of up to 9 inches (c. 23 cm) of subsidence from the eastern side of the Rashidpur Anticline to Shaitaganj and Sahaji Bazar. Occasional subsidence extended even further from the epicentral area. It seems likely that the 1918 Srimangal earthquake actually occurred within the Indian Basement, below the Tripura Fold Belt. If so, it probably had a strike-slip mechanism similar to those modelled by Mitra et al. (2005), which were found to have centroid depths of 27–45 km (Fig. 2.12). As noted above, the unusually large seismogenic thickness in this area allows such large earthquakes to occur on intraplate faults.

The seismic hazard posed by shallow structures in the Tripura Fold Belt is currently unclear. Further analysis of the structural relationship between the folds and underlying faults may help to resolve this question. It is, however, clear that very large earthquakes may occur within the Indian Plate beneath the shallow low-angle detachment. This is partly a consequence of the very high seismogenic thickness of this area, which increases the potential rupture area and is often associated with relatively high earthquake stress drops. While deeper earthquakes have lower peak shaking intensities, a $M_w$ 7+ earthquake beneath the upper sedimentary layers would still cause devastation over a wide area. It is unfortunate that the principal seismic hazard in this area may be posed by deformation occurring below a shallow detachment, severely limiting our ability to assess seismic hazard through surface observations.

The arcuate shape of the Tripura Fold Belt has been explained as a response to the impedance provided by Shillong Plateau (e.g. Wang et al. 2014). Alternatively, the curved shape of the range may have arisen as a natural consequence of gravitationally driven flow (Copley & McKenzie 2007; Copley 2012; who also suggest the curvature of the entire Indo-Burman Range may arise through this process). If this process occurred over a stress-free boundary, such as a very-low-viscosity overpressured shale layer, then it would account for the gentle topographic slope across this range and the recent rapid westwards propagation of deformation described by Maurin & Rangin (2009). A similar mechanism has recently been proposed for the morphology of the Sulaiman Ranges in the west in Pakistan (Reynolds et al. 2015). The northern part of the IBR, the Naga Hills, lacks the broad fold-and-thrust belt and low-angle long-wavelength topographic slope. More work is needed to explore the effect of lateral variations in structure and material properties in the development of this and other mountain ranges.

**Indo-Burman Ranges (IBR)**

Little instrumental seismicity has been recorded within the IBR, except for relatively deep earthquakes within the Indian Plate (e.g. Copley & McKenzie 2007; Stork et al. 2008; Kyi Khin et al. 2017) (Figs 2.12 and 2.13). A recent example was the $M_w 6.7$ 2016 Tamenglong earthquake, which took place 55 km below the IBR (Gahalaut et al. 2013; Parameswaran & Rajendran 2016). Wang et al. (2014) suggests that there may be ongoing shortening immediately to the east of the range on an east-dipping structure, although there is no seismic or geomorphic evidence for active shortening in the higher part of the range. Instead, active deformation appears to occur on a number of dextral strike-slip faults within the high Indo-Burman Ranges (Wang et al. 2014). This suggests that a significant part of the oblique motion between Indian and the Sunda plates may be taken up in this area (Maurin & Rangin 2009; Rangin et al. 2013). In the south, the 160 km long Thahtay Chaung Fault has produced c. 11 km right-lateral offsets on a number of river canyons. These offsets reduce to 5 km in the north, before becoming indistinct. No direct constraints on slip-rate are available, but if the drainage system is assumed to have developed at around 5 Ma then this would provide a slip rate of around 2 mm a$^{-1}$. 

Downloaded from http://mem.lyellcollection.org/ by guest on March 17, 2021
In the north, the 170 km long Churachandpur-Mao Fault follows the western edge of the Imphal Valley (Wang et al. 2014). This fault has produced right-lateral stream offsets of up to 3 km, and again becomes more indistinct to the north and south. The sharp change in topography associated with the fault also suggests that it is, or has been, associated with vertical movement. GPS constraints in this area suggest that up to 16 mm a⁻¹ of right lateral motion occurs between Imphal and India. It is not clear how much of this localized on the Churachandpur-Mao Fault rather than in oblique slip on the interface between India and the Indo-Burman Ranges or other cryptic faults within the ranges. The most detailed fault map of the region available was based principally on the analysis of SRTM 90 m topography (Wang et al. 2014), and distributed strike-slip faulting could be difficult to observe. Gahalaut et al. (2013) suggested that all 16 mm a⁻¹ of motion may be taken up by this fault, and observed a very sharp c. 6 mm a⁻¹ discontinuity in the fault-parallel GPS velocities across the structure. The sharpness of the discontinuity suggests that the fault may be slipping aseismically at shallow depths. It seems most reasonable to assume that c. 6 mm a⁻¹ is associated with this fault and the remainder is taken up on other structures, as suggested by Maurin & Rangin (2009). The observed stream offsets could be explained with just 0.5 Ma of slip at this rate, suggesting that this fault may be very young or has accelerated recently. Unlike the Tripura Fold Belt, there is little sign of systematic range-perpendicular shortening between GPS sites in the high Indo-Burman Ranges (Gahalaut et al. 2013) (Fig. 2.13).

Seismicity within the downgoing Indian Plate

A number of moderately large earthquakes within the downgoing Indian Plate have been modelled to determine their focal mechanism and centroid depth (Chen & Molnar 1990; Mitra et al. 2005; Copley 2008; Kumar et al. 2015) (Fig. 2.13). Earthquakes within the Indian Plate have been relocated by Stork et al. (2008) and Hurukawa et al. (2012). Stork et al. (2008) focused on a smaller number of events, and attempted to use depth phase identification to give better depth resolution in the absence of local seismometers. All these studies found that most seismicity beneath the Indo-Burman Ranges takes place within the subducting plate, and that the plate steepens sharply beneath the eastern edge of the IBR. This steepening is significantly more pronounced in the north. Earthquake focal mechanisms suggest that the recorded events occurred on steeply dipping planes within the downgoing plate, not on the plate interface (Hurukawa et al. 2012; Kundu & Gahalaut 2012). Interestingly, within the western part of the central Myanmar lowlands, both Stork et al. (2008) and Hurukawa et al. (2012) found evidence for lower crustal or upper mantle earthquakes above the slab. If these depths are correct, it would suggest that this region is relatively strong, and would explain why it has remained relatively low, potentially able to transfer large stresses as a relatively rigid body. Russo (2012) interpreted the locations obtained by Stork et al. (2008) as revealing a number of vertical tears through the slab. This interpretation most likely arises from the relatively small number of events studied by Stork et al. (2008). There is no evidence for this in the larger dataset of Hurukawa et al. (2012), who favour a continuous smoothly curving slab.

Rangin et al. (2013) reviewed the evidence against an active megathrust, and noted that the earthquakes within the downgoing plate have predominantly T-axes pointing down the dip of the slab (indicating downdip extension) and P-axes oriented along the strike of the slab (approximately parallel to Indian Plate motion with respect to the Sunda Plate) (Fig. 2.13). There are, however, a number of active subduction zones that have similarly oriented earthquakes within their slabs, for example the Hellenic subduction zone in the Aegean where ongoing subduction of the Nubian Plate is undisputed (Shaw & Jackson 2010). Downdip extension is a common feature of subducted slabs that have not reached the 660 km discontinuity, and reflects extension driven by the weight of the cold, dense slab below (Isacks & Molnar 1971). The north–south orientation of P-axes may reflect shortening associated with the northwards motion of the Indian Plate, or alternatively may be related to the curvature of the downgoing plate (as suggested in the strongly curved Nubian slab beneath the Aegean subduction zone; Shaw & Jackson 2010). The amount of seismicity in the slab is elevated beneath the easternmost IBR and the Chin- dwin Basin, where the slab bends sharply as it steepens. This seismicity may be at the result of local bending stresses within the slab (Fig. 2.13).

Continent–ocean transition

One question of particular interest relates to the nature of the crust beneath the sediments of the Bengal Basin. The boundary between the oceanic crust of the Indian Ocean, subducting beneath the Andaman Islands to the south, and the continental material to the north is difficult to pinpoint. A hinge zone (dotted grey line in Fig. 2.12) is commonly identified as the Eocene palaeoshelf edge, marking the boundary of the thick sediment pile and associated with a sharp increase in depth of the shallow-water Sylhet Limestone (Alam 1989; Alam et al. 2003). Deep seismic sounding lines crossing the hinge line confirm that the crust thins sharply here but it is concluded that, close to the hinge line, the evidence is consistent with thinned continental crust rather than with true oceanic material (Kaila et al. 1992). Receiver functions sampling the crust immediately to the south of the Shillong Plateau also support this conclusion (Mitra et al. 2005).

Unfortunately both these studies sample crust only close to the inferred palaeoshelf edge. Consequently, there is little solid evidence for the location of the continental–oceanic transition beneath the Bengal Basin. Mitra et al. (2008) present a receiver function in the Eastern Bengal Basin at Aragatala, which has a 6 km thick high-velocity layer in the lowermost crust, and suggested that this may represent oceanic crust. This would require a 30 km thick sedimentary layer, with very high velocities in the lowermost 10 km of the sediment pile. Recent work, involving a joint inversion of P-receiver functions and Rayleigh wave group velocity dispersion, suggests that the sediment–basement transition may occur at the much larger velocity discontinuity observed at 18–20 km, and that the crust here is likely to be thinned continental material rather than true oceanic crust (Mitra et al. 2014). Similarly Singh et al. (2016) found that receiver functions from Dhaka revealed evidence for a relatively thin (c. 16 km) crystalline basement, and suggested that this may represent oceanic crust that was thickened due to the activity of the Kerguelen plume. These results suggest that much of the Bengal Basin is underlain by thinned continental material, or thickened oceanic material, which may extend further east beneath the Indo-Burman Ranges. In the northern Bay of Bengal, Mitra et al. (2011) determined that surface-wave dispersion patterns are consistent with oceanic crust, overlain by an extremely thick and actively metamorphosing pile of sediments. They suggested that the earlier conclusion that this area was underlain by thinned continental crust (Brune & Singh 1986) was a misinterpretation of unusually high velocities within the highly metamorphosed base of this exceptionally thick sediment pile. However, recently multi-channel seismic data from the offshore northern Bay of Bengal have been interpreted as indicating 15 km thickness of thinned continental crust injected by Mesozoic volcanism (Sibuet et al. 2016; Rangin & Sibuet 2017). Talwani et al.
The character of the Indian crust currently beneath the Indo-Burman Ranges is of great importance, because the density difference between continental and oceanic crust may play an important role in controlling the tectonics of this region (Clark & Bilham 2008). The presence of a seismically active slab, extending to depths of c. 160 km beneath the central Myanmar lowlands, suggests that the downgoing plate is oceanic at this point. The ability of continental lithosphere to be subducted to these depths is controversial as, unlike oceanic lithosphere, it would be expected to be positively buoyant. Note that some workers have suggested that continental material is subducted to great depths beneath the Hindu Kush (e.g. Searle et al. 2001; Schneider et al. 2013). The presence of Pliocene–Recent calc-alkaline volcanism near the Sagaing Fault (Mounts Popa, Taungthonlon and Mount Loinyee; Stephenson & Marshall 1984) confirms that the Indian slab is oceanic at depth. It is tempting to place the transition between continental and oceanic lithosphere beneath the high Indo-Burman Ranges where earthquake locations indicate that the slab abruptly steepens (Fig. 2.13); however, this remains a matter of conjecture.

The plate boundary

The main sources of seismic hazard in NE India and Bangladesh remain uncertain. Vast amounts of sediment (>1 Gt a−1) are transported from the Himalaya and Indo-Burman Highlands by the Brahmaputra and Ganges rivers and deposited in the Bengal Basin (Milliman & Syvitski 1992). The accumulated post-Eocene sediments are over 20 km thick in places, and shroud the surface expression of the active tectonics (Alam et al. 2003; Steckler et al. 2008). The continuation of the plate boundary between the Indian and Eurasian continents must pass beneath this cover, probably projecting to the surface near eastern Dhaka, but the position and nature of this boundary remains an issue of active debate.

On Boxing Day 2004 the Mw 9.2 Sumatra–Andaman earthquake, ruptured 1300 km of the Indian Ocean plate boundary, immediately to the south of Bangladesh and Myanmar. This devastating earthquake and the associated tsunami killed more than 280 000 people and caused widespread destruction (Lay et al. 2005). The stress changes associated with the 2004 earthquake will have loaded the northwards extension of the plate boundary fault, potentially bringing it closer to seismogenic failure. If the extension of the megathrust beneath Bangladesh (the Dhaka segment) is capable of rupturing in large events, then it could easily produce the most devastating earthquake in history. Extreme population densities, the vulnerability of the existing building stock, the potential for liquefaction and seismic-wave amplification within the thick sediment pile and the low seismic attenuation of the Indian Shield (which would result in greater shaking intensities over wider areas) combine to produce extremely high earthquake vulnerability. It is, therefore, particularly important to determine whether this plate-boundary fault ruptures in large earthquakes or slips aseismically, either continuously or in periodic slow-slip events.

Within the northernmost SW–NE-trending part of the Indo-Burma Ranges, the Naga Hills GPS data indicate slow (1–4 mm a−1) ongoing convergence (Vernant et al. 2014). Here, the underthrust Indian Plate is continental. Ophiolite obduction onto the eastern margin of India is sealed by the Late Eocene–Oligocene Jopi Formation, suggesting that oceanic subduction on the northernmost side of the Indo-Burma Ranges may have ceased by this time (Ghose & Chatterjee 2014). Further south, it is clear that subduction of an oceanic plate has occurred in the recent geological past because there is a continuous dip-slip plane of earthquakes observed in earthquake relocation studies, and because there is Pliocene–Recent calc-alkaline volcanism above the Burma Seismic Zone (Stephenson & Marshall 1984; Stork et al. 2008; Hurukawa et al. 2012). However, there is an active debate about whether continental material is now being thrust beneath the Tripura Fold Belt, and even the existence of an active megathrust in this region is disputed.

Some workers view the Tripura Fold Belt as the expression of an active accretionary prism associated with ongoing subduction beneath the Indo-Burman Ranges (e.g. Steckler et al. 2008; Wang et al. 2014). Others suggest that subduction beneath the Indo-Burman Ranges has now completely ceased and that the shortening associated with the TFB is related to the gravitational collapse of Tibetan material in the east, transmitted through the relatively rigid Myanmar Central Basin (Rangin et al. 2013). Alternatively, the gravitational driving force could simply be provided by the pre-existing topography of the adjacent Indo-Burman Ranges (Copley & McKenzie 2007). It is, of course, possible, even expected, for gravitational collapse to be partially driving deformation in an area with an active subduction zone. For example, this is currently occurring in the Aegean (e.g. Floyd et al. 2010).

There is no clear evidence in the instrumentally recorded seismicity of rupture on the subduction zone interface in this area (Kundu & Gahalaut 2012; Rangin et al. 2013). As discussed above, the only low-angle thrust with a well-determined centroid depth is too shallow to be located on the plate interface. This absence of identified interface seismicity cannot rule out an active interface conclusively. Convergence could be accommodated aseismically, or the interface could fail only in very large and infrequent earthquakes and be currently locked and accumulating strain.

The historical record

We now consider evidence for large pre-instrumental earthquakes in the area (Fig. 2.12). The rapid sedimentation rate in this area might be expected to obscure the geomorphic signatures of even large earthquakes quickly, although there is a reasonably long historical account of major events. The 1762 Arakan earthquake is thought to have ruptured the megathrust beneath the Bay of Bengal. This earthquake uplifted a number of islands along the Arakan coast and caused extensive damage as far north as Chittagong (Cummins 2007). Wang et al. (2013) collected extensive field data for coastal uplift associated with this earthquake on Cheduba and Ramree islands. They inferred that 9–16 m of slip occurred beneath the islands corresponding to an $M_w$ 8.5 earthquake, and suggested that the repeat time of such an earthquake would be 500–700 years. The double-humped pattern of uplift beneath the two islands led them to conclude that slip on the megathrust was transferred, in part, to two steeper imbricate faults associated with the long-term anticlinal growth of the two islands. A repeat of this earthquake would have regionally devastating societal effects, especially if it were associated with a tsunami.

North of the Arakan segment, as the plate boundary continues onto land, there is no unambiguous historical record of megathrust rupture and even the precise location of the fault becomes controversial. Steckler et al. (2008) reviewed the location and nature of this plate boundary. They suggested that in the south the westernmost expression of the plate boundary may be the Comilla Terrace, which has been uplifted by 3–4 m relative to the rest of the delta. North of Dhaka the
location of the fault is less clear. Most workers suggest that it most likely follows the trend of the Tripura fold belt and curves east, passing to the east of the Shillong Plateau (e.g. Wang et al. 2014). Steckler et al. (2008) suggest that it may continue to the NW, meeting the western edge of the Shillong Plateau.

Steckler et al. (2008) suggest that a major earthquake, reported to have caused damage across a wide area from Syllah to Chittagong in the south in 1548, may have occurred on the section of the megathrust east of Dhaka. This earthquake is more commonly associated with faults further to the north. Morino et al. (2011) found palaeoseismic evidence for a significant rupture on the Dauki Fault with an age consistent with the 1548 rupture. OSL dating of sand pipes in the Brahmaputra Valley in the NE also provides dates consistent with the suggestion that this earthquake may have occurred on the northern margin of the Naga Hills (Thomas et al. 2007). The attribution of this earthquake to a specific fault would be an important step in the analysis of seismic hazard in this area. If it did occur on the megathrust, it would reveal the nature of this important plate boundary. However, unlike the 1762 earthquake, the evidence remains circumstantial.

**Geodesy**

India–Sunda plate motion is highly oblique to the plate margin in this region (e.g. Socquet et al. 2006; Banerjee et al. 2008). The Indian Plate is moving NNE relative to Sundaland. This means that there is a significant component of convergence (although still oblique) across the Arakan segment of the subduction zone, which strikes NW. On the other hand, as the subduction zone curves round in the north the overall plate motion becomes increasing hyper-oblique, and it becomes less clear whether active subduction may be expected to occur. Some workers have suggested that the motion between the Indian Plate and Sundaland is taken up by strike-slip faults, such as the Sagaing Fault in the East and the Churachandpur-Mao Fault in the high Indo-Burma Ranges, rather than through slip on a subduction zone interface (e.g. Socquet et al. 2006; Gahalaut et al. 2013).

Recent GPS measurements from the Tripura Fold Belt show c. 10–13 mm a$^{-1}$ of east–west shortening within the TFB (Fig. 2.13) (Gahalaut et al. 2013; Steckler et al. 2016). It is therefore clear that, despite the hyper-oblique plate convergence, significant east–west shortening is occurring in this area. Steckler et al. (2016) have also demonstrated that the distribution of this shortening is consistent with the accumulation of elastic stresses on a locked plate interface. The geodetically measured rate of shortening is consistent with a $M_w$ 8.5 earthquake every c. 500 years if the plate boundary is locked (although this estimate of the repeat time is dependent on poorly constrained parameters such as the geometry and locking depth of the plate boundary). This means that if the 1548 earthquake did occur on the plate boundary then shortening could be taken up predominately elastically on the fault and, furthermore, this section of the fault could be due for another major earthquake in the near future. Further work is needed to examine if this east–west shortening could instead be explained by gravitationally-driven largely aseismic folding and pressure solution creep within shallow water-saturated sediments.

A number of authors have suggested that subduction zones characterized by the huge quantities of sediment and associated high pore-fluid pressures may limit the strain-weakening behaviour required for great earthquakes (e.g. Pacheco et al. 1993). However, as Steckler et al. (2008) point out, major events do occur in high-sediment-input systems such as the 1964 Alaskan earthquake (Ruff 1989, 1992), and the high geothermal gradients associated with thick sediment piles may actually promote stick-slip behaviour at relatively shallow depths. Wang et al. (2013) confirmed that slip did occur on the megathrust interface on the Arakan segment beneath the offshore extension of the Ganges–Brahmaputra delta in the Bay of Bengal, suggesting that sediment-clogged system is capable of failure. The fact that slip in the Arakan earthquake appears to have been transferred from the megathrust onto imbricate faults within the accretionary prism (Wang et al. 2013) supports the suggestion that severe damage was not reported in western cities such as Dhaka after the 1548 earthquake, due to a similar process (Steckler et al. 2008). Steckler et al. (2008) suggest that this behaviour may occur because the burial of the tip of the accretionary prism beneath the sediments of the Ganges–Brahmaputra delta promoted the transfer of megathrust slip onto imbricate thrusts to thicken the deforming wedge.

Nowhere else are such large populations exposed to so potentially high, yet so uncertain, seismic hazard. There is an urgent need to further investigate how the c. 13 mm a$^{-1}$ of shortening observed in GPS studies is accommodated, and to better determine the geometry of the system. Some workers have suggested that the plate boundary is unlikely to be an active, seismogenic megathrust in its onshore section, and consequently conclude that the seismic hazard in this area may be low. Even if this section of the megathrust is aseismic, the Arakan segment, the Shillong Plateau and faults within the Indian Plate buried beneath the thick sedimentary pile are all proven and potentially devastating sources of hazard.

**Andaman Sea**

The northern and northeastern part of the Andaman Sea lies within Myanmar (Fig. 2.14). However, to understand the tectonic development of the area it is necessary to view the region as a whole and discuss areas that lie in Indian, Indonesian and Thai waters. The region is located along a highly oblique subduction margin (north–south-trending trench, NNE–SSW motion of the India Plate relative to Sundaland) where a back-arc spreading centre (Central Andaman Basin) has developed in a pull-apart setting between the Sagaing Fault to the NE, and the West Andaman–Sumatra Fault Zone to the south and SW (Fig. 2.14). Instantaneous motions from GPS data along the Sagaing (1.8 cm a$^{-1}$) and Sumatra fault zones (2.3 cm a$^{-1}$; Genrich et al. 2000; Maurin et al. 2010) are sufficiently similar to support the regional linkage of the fault systems.

The Andaman Sea can be divided into six distinct tectonic regions. (1) First is the trench-accretionary prism complex (Andaman, Nicobar, Coco Islands) in the west (location A in Fig. 2.14). (2) Immediately to the east of the accretionary prism complex lies a region dominated by active transtensional strike-slip faults (particularly the Sumatra Fault Zone and the West Andaman Fault) and some east-vergent folds and thrusts in sedimentary basins. The Andaman Sea exhibits considerable bathymetric relief, with long, linear submarine ridges (Curray 2005; Cochran 2010) (location B in Fig. 2.14). (3) East of the strike-slip faults are the rugged, relatively high areas of the Alcock and Sewell rises. The rises are separated by the deep, NNE–SSW-trending Central Andaman Basin, which is interpreted as a spreading centre (Curray et al. 1979; Raju et al. 2004) (location C in Fig. 2.14). (4) East of the rises is a north–south- to NNE–SSW-trending sediment-filled deep-water trough called the East Andaman Basin. This basin has been the focus of Pliocene–Recent sedimentation in the eastern part of the Andaman Sea (location D in Fig. 2.14). (5) Passing up the continental slope to a drowned shelf in the southern and central part of the eastern Andaman Sea, a basement high called the Mergui Ridge is encountered. East of the ridge lie Cenozoic rift basins including the Mergui Basin (Thailand), the North
Sumatra Basin (Indonesia) and a number of less well-developed rift basins are present on the shelf in Myanmar, including the North and South Mali basins and the Taninthari basin (location E in Fig. 2.14). Unlike other parts of the Andaman Sea, this shelf has not been the site of significant Pliocene–Recent tectonic activity although a few faults, such as the Ranong Fault, do appear to have undergone minor reactivation during this time. (6) North of the Alcock Rise lies the slope and shelf region of the Gulf of Martaban (Fig. 2.14). This region is the southerly extension of the Myanmar Central Basin, and also contains the offshore extension of the Sagaing Fault. The Gulf has been affected by Eocene and Early Oligocene rifting, localized Miocene inversion, and the Pliocene–Recent or Late Miocene–Recent strike-slip fault activity.

Trench and accretionary prism

The Indo-Burma Trench has been surveyed by the Andaman Cruise (Nielsen et al. 2004), Scripps cruises from 1963 to 1979 (Curry 2005), Lamont-Doherty Earth Observatory, the US Navy and NOAA (Cochran 2010), as well as by some petroleum industry seismic surveys including a 2500 km survey by the Directorate General of Hydrocarbons (India; Goli & Pandey 2014). According to Nielsen et al. (2004), the upper plate at the trench is marked by the West Burma Scarp which links the 3000 m deep Bengal Basin to the west to the Coco Ridge to the east (Fig. 2.15). Typical accretionary prisms have a relatively low taper slope to the top of the wedge (1°–7°) which contains extensive large folds, thrusts and ponded basins (e.g. Dahlen 1990; Morley et al. 2011). The width of the deforming part of the accretionary prism narrows from 80–100 km along the southern part of the Andaman Trench to c. 40 km between 10° N and 11°30′ N (Cochran 2010). The unusually steep (12°–15°) and narrow West Burma Scarp is interpreted to be a strike-slip-dominated margin (Nielsen et al. 2004). This interpretation fits the NNE motion of the Indian Plate relative to Eurasia. North of the Andaman Islands the change from a north–south to N40° E trend is associated with a change from transpressional convergence to pure dextral strike-slip (Nielsen et al. 2004). The NE-trending part of the margin is associated with multi-scale en echelon folds, with axes trending N10° E to N35° E, and rhomboidal fault patterns (Fig. 2.16). There is a gradual change northwards towards more north–south-trending segments that show mixed NE–SW-trending strike-slip faults and north–south–to NNW–SSE-trending thrusts (Nielsen et al. 2004).

West Andaman Fault Zone

The duration of activity of the major strike-slip faults is difficult to establish. The West Andaman Fault and other faults (e.g. Diligent Fault, Eastern Margin Fault) on the western margin of the Andaman Sea (Fig. 2.14 in location B) are major geomorphological features and were clearly very important fault zones during the Pliocene (Curry 2005; Cochran 2010). The considerable topographic relief and c. 70° dip of the West Andaman Fault zone (Goli & Pandey 2014) suggests there is a considerable extensional component to the displacement. Between the fault-bounded highs are basins several kilometres deep, thought to be filled with Miocene–Recent sediments. Seismic reflection data along the southern portion of the West Andaman Fault Zone suggests that it was only active since the Middle Miocene or younger (depending upon how unconformities on the seismic data are correlated; Berglar et al. 2010). Further north, however, long-distance extrapolation of horizons on seismic data from shallow-water wells east of the Andaman Islands suggests that fault activity may extend...
back to the Early Miocene or even the Oligocene (Goli & Pandey 2014).

Barren Island and Narcondam Island are subaerial volcanic edifices west of the main Sewell Rise, forming seamounts within the region of strike-slip-transtensional faulting. The lavas are calc-alkaline and subduction-related, with the Narcondam lavas exhibiting a contribution from thinned continental crust and/or sediments from the downgoing...
slab (Streck et al. 2011). Earthquake swarms to depths of about 35 km, associated with the volcanoes, are interpreted by Špičák & Vanečk (2013) to be related to the intrusion of magma.

Plagioclase xenoliths from the Barren Island lavas have been $^{40}\text{Ar}/^{39}\text{Ar}$ dated at 106 ± 3 Ma by Ray et al. (2015). They infer, based on dating, isotope compositions and mineralogy, that the xenoliths are derived from a lower oceanic crust gabbro that is genetically linked with the 95 Ma Andaman ophiolite. However, there is a problem with inferring an oceanic crust basement to the Andaman Islands. The shear wave velocity structure of the Andaman islands from joint inversion of teleseismic receiver functions and Rayleigh wave group velocity measures indicates the following layers are present: 2–6 km $v_c$. 1.3–2.5 km/s (Andaman flysch sediments); 12–14 km $v_c$. 3.5 km/s (silicic material), 8–12 km mafic layer ($v_c$. 4.0 km/s). Hence the crustal thickness is around 24–32 km, with a structure akin to continental crust (Gupta et al. 2016). Consequently, the xenoliths from Barren Island indicate oceanic crust that is too old to be part of the Central Basin, and there is continental as well as ophiolitic basement underlying the Andaman Islands. Instead the Barren Islands appear to be underlain by Upper Cretaceous oceanic crust trapped between regions of continental crust to the west (Andaman Islands) and continental crust or Neogene oceanic crust to the east (Alcock, Sewell Rises, Central Basin, see following section).

Alcock and Sewell rises

Understanding the crustal structure and geological evolution of the Alcock and Sewell rises is the key to understanding the development of the Andaman Sea. The rises are separated by the ENE–SWS-trending Central Andaman Basin, discussed in the following section. Seismic reflection data across the rises indicate they have a thin covering of sediment in places to virtually no sediment. They have a very irregular topography, attributed to a mixture of volcanic cones, normal faults and strike-slip faults (Srisuriyon & Morley 2014). Curray (2005) noted the occurrence of dredged tholeiitic basalts dated at c. 20 Ma from the Alcock Rise, which he associated with the formation of back-arc crust in the Andaman Sea.

Interpretation of gravity data suggests the crust beneath the Sewell and Alcock rises varies considerably in thickness, but may be in the order of 15–20 km in many places (Radhakrishna et al. 2008). It can be argued that the relatively thick oceanic crust is due to the effects of magmatic underplating and volcanism. However, gravity modelling of Seasat data suggests the rises are composed of thinned continental crust (Morley & Alvey 2015).

East Andaman Basin

The East Andaman Basin is a large sedimentary basin on the eastern side of the Sewell Rise, downfaulted against the rise, and lies between the rise to the west and the Mergui Ridge to the east. In the southern part of the basin two main sedimentary packages are present: a lower package that is intensely faulted by predominantly ENE–WSW-trending normal faults and occasional north–south-trending strike-slip faults (Morley et al. 2011); and an upper package that is much less affected by faulting. The upper package overlies, and seals, most of the faults in the lower package. The age of this event is early Middle Miocene (by regional correlation on seismic lines to the Mergui Basin). The underlying sequence can be thick in places (>4 km) and has not been drilled, but is probably of Oligocene–Lower Miocene age.

Further north the Middle Miocene–Recent section, overlying the Middle Miocene event, is found within a west-thickening basin that is extensively affected by normal faults (Fig. 2.17). The predominantly Late Miocene–Recent sediments infilling the basin are present both north and south of where the ENE–WSW-trending Central Andaman Basin intersects the East Andaman Basin (Figs 2.14 and 2.18). Considerable extension of the area north and south of the Central Andaman Basin during the Late Miocene–Recent is implied.

**Fig. 2.17.** Schematic cross-section from the eastern Alcock Rise across the East Andaman Basin, SE offshore Myanmar. The section illustrates a model for hyper-extended continental crust flooring much of the East Andaman Basin. See Figure 2.14 for location.
Central Andaman Basin

The Central Andaman Basin (CAB) is generally accepted to be underlain by back-arc oceanic crust (e.g. Curray et al. 1979; Raju et al. 2004; Curray 2005; Diehl et al. 2013). In his regional plate reconstructions, Hall (2002) showed the prevailing view at the time that the Andaman Spreading Centre opened during the Middle Miocene (following Curray et al. 1979). However, the geophysical study of the Central Basin by Raju et al. (2004), including reappraisal of the magnetic data, indicated the basin has formed by continual spreading from c. 4.0 Ma to the present. This interpretation is supported by Curray (2005). The eastern half of the CAB is covered by a blanket of sediment, while only in the western part, in a region 60 km wide, have poorly constrained oceanic-crust-type magnetic anomalies been interpreted (Raju et al. 2004).

Curray et al. (1979) and Curray (2005) reported data acquired over the eastern Central Basin that showed the presence of sediment up to the axial trough, and recognized the sediments could indicate that spreading was episodic rather than continual. However, they ultimately favoured a continuous spreading model. Raju et al. (2004) interpreted the magnetic anomalies as showing initial slow spreading rates of 1.6 cm a⁻¹ beginning at c. 4 Ma, that increased to 3.8 cm a⁻¹ from anomaly 2 to the present (i.e. the past 0.7 Ma).

The CAB is characterized by extensional earthquakes (e.g. Raju et al. 2004; Diehl et al. 2013). Earthquake activity indicates that only 10% of the long-term spreading rate of 3.0–3.8 cm a⁻¹ is accounted for by extensional faulting, and modelling of the earthquake swarms suggests the presence of intrusive dyke activity (Diehl et al. 2013). Consequently, these authors concluded that igneous intrusions account for 90% of current extension.

Srisuriyon & Morley (2014) and Morley & Alvey (2015) argued that the geometry of sediments (Fig. 2.19) and their thickness (up to 1 km within the central trough) around the eastern half of the CAB indicate that continuous spreading was not possible. Instead, they suggest episodic spreading with a probable Late Miocene–Early Pliocene phase of spreading followed...
by a hiatus, then very recent renewed spreading or extension in the trough, perhaps only beginning in the last several tens of thousands of years.

Jourdain et al. (2016) provide details about the spreading centre geometry from 2D seismic reflection data. They suggest that the spreading centre is composed of oceanic crust, and they show some younging of deeper sediments towards the spreading centre. Yet the central problem on the seismic reflection data remains explaining the thick sedimentary section that occupies the central trough in terms of active spreading (Morley & Alvey 2015). One atypical feature of the spreading centre is the apparent absence of extrusive basalt, which instead is replaced by a sill-segment sequence of c. 50% sediment and 50% igneous material (Jourdain et al. 2016). However, sills do not accommodate crustal extension in the same way as dykes (sills enlarge vertically, dykes horizontally). Hence, perhaps the sills also indicate a quiescent phase in the spreading centre development.

Gulf of Martaban

The offshore area in the Gulf of Martaban shows two main stages of structural development: Oligocene–Early Miocene rift development; and Late Miocene–Recent rapid subsidence, accompanied by extensional and strike-slip faulting in the eastern half of the area (Figs 2.15 & 2.20). Evidence for Middle–Early Late Miocene deposition is sparse for a variety of reasons (well locations on highs or in areas of very thick Late Miocene–Recent sediments), but is also due to the lack of sediments reaching the offshore area during this time, particularly in the western gulf.

In southern onshore Myanmar, and offshore in the Gulf of Martaban, this basinal area becomes much wider and deeper (Fig. 2.20). Offshore, the Sagaing Fault comprises three and, further south, two north–south-trending main fault traces or principal displacement zones that lie at the deepest part of a synformal depocentre (Figs 2.15 and 2.20). This strike-slip basin, located predominantly offshore, is remarkable in that its axis lies parallel to the Sagaing Fault Zone, and there is no obvious releasing bend geometry to explain the location of the basin. The basin depocentre is not confined to one side of any of the principal displacement zones, but overall the basin is thickest west of the strike-slip faults (Figs 2.15 and 2.20). This basin represents the northern extension of the East Andaman Basin.

There are several factors contributing to the location and the high rate of subsidence in the basin. First, the north–south-trending Peninsular Thailand margin underwent oblique extension/transension from the Late Eocene–Early (or in places Middle) Miocene (Morley & Racey 2011; Morley et al. 2011). This extension evolved into a thermally subsiding north–south-trending passive margin segment. The second factor is the later offshore east–west-trending Miocene–Recent rifting of the Alcock and Sewell rises and the Central Basin (Fig. 2.3). The east–west segment is also now in the early stages of post-rift subsidence, and intersects the older north–south rift trend in the vicinity of the Sagaing Fault Zone. Third, the Sagaing Fault Zone is an important fault zone, accommodating about half of the northwards motion of India relative to Indo-China (Vigny et al. 2003). Since it is effectively a plate boundary, the fault zone is likely to be of lithospheric extent (e.g. Searle & Morley 2011; Searle et al. 2011). Such a fault zone in the context of isostatic modelling would have greatly weakened and considerably reduced the effective elastic thickness of the crust in comparison with adjacent areas. The fourth factor is the very high sediment supply, focused along the synformal trough by the Sittoung and Salween rivers. A rough estimate of the modern sediment supply to the eastern Gulf of Martaban by these rivers is around 240 million tons/year and for the Irrawaddy River about 364 ± 60 million tons/year (Robinson et al. 2007). This represents 20% of the total flux of material from the Himalayan–Tibetan orogeny (Robinson et al. 2007). This combination of deep-water passive margin segments, high sediment supply, low effective elastic thickness and a north–south-trending strike-slip fault zone suited to focus sediment along a north–south extending synformal basin, where offshore the Pliocene–Recent section can exceed the 6 s two-way travel time on the 2D seismic record (i.e. thicknesses exceeding c. 8 km).

Offshore, the Miocene–Recent fault pattern in the Gulf of Martaban can be related to a combination of three factors: crustal extension; delta-type gravity tectonics; and localized strike-slip faulting (particularly along the Sagaing Fault) (Fig. 2.15). The two principal displacement zones that comprise the Sagaing Fault Zone offshore are narrow, vertical features that pass up to the sea floor. Close to the principal displacement zones (i.e. about 1–2 km on either side) there is intense deformation, uplift of beds and formation of pressure ridges, well exposed in places onshore. Locally, accompanying the fault zone, there are overpressured fluids (onshore these are manifest as sand and shale injection features). More than a few kilometres away from the principal displacement zones there is a general absence of any faults that could be described as R (Riedel) or P shear trends (i.e. faults oriented at an acute angle of about ±16° to the master strike-slip fault) (Fig. 2.20). Instead, the offshore fault pattern is dominated by ENE–WSW-trending extensional faults and north–south-trending strike-slip faults (Figs 2.3 and 2.20). This pattern is broken passing towards the tips of the strike-slip fault trends where a fanning, horse-tail pattern of faults has developed.

These east–west– to ENE–WSW-trending faults are not associated with significant seismicity and many appear to have a listric fault shape, suggesting that they detach within the sedimentary section. The density of growth faults for a deltaic depocentre is, however, unusual. Typically deltas become organized.

---

Fig. 2.19. Line drawing interpretation of a seismic line across the Central Andaman Basin, and putative spreading centre (original seismic line in Raju et al. 2004). See Figures 2.14 and 2.18 for location.
Fig. 2.20. Map of the main basins and tectonic features in the northern part of the Andaman Sea.
into systems of large growth faults with widely spaced (15–20 km) depocentres, bounded by large-displacement growth faults (commonly 2 km to >6 km offset) (e.g. Morley 2003). This organization is seen on the west side of the Late Miocene–Recent basin, but is lost passing eastwards towards the depocentre. The width of the fault pattern has increased (averaging 1.5 km apart), and large displacement faults are lacking. Throws on most faults are in the range of tens to hundreds of metres. The structural style comprises conjugate convergent sets of faults associated with broad synformal and antiformal geometries. The fault pattern probably reflects the effects of delta loading, overpressure, weak sediments and growth faulting, interacting with the large-displacement, seismically active Sagaing Fault. The western principal displacement zone of the Sagaing Fault dies out north of the eastern principal displacement zone, and it is the eastern one that transfers its displacement onto structures in the central Andaman Sea (Figs 2.3 and 2.20).

Models for the tectonic development of the Andaman Sea

The widely accepted model for the Andaman Sea is that a spreading centre east of the Alcock Rise depicted in Curray (2005) is not supported by unpublished petroleum industry seismic data. However, inversion of gravity data suggests that there is a small patch of oceanic crust east of the Alcock Rise (Fig. 2.18). A speculative, schematic cross-section from the Mergui Shelf to the Alcock Rise is shown in Figure 2.17. The section is based partially on seismic reflection data, and shows that the north–south-trending East Andaman Trough is possibly underlain by hyper-extended continental crust. A thick section of Late Miocene–Recent sediments overlies the area of thinnest crust, and is offset westwards from earlier rift basins. It is far from clear whether the 3.8 cm a\(^{-1}\) modern spreading rate proposed by Raju et al. (2004) is reasonable for the tectonic model for the region. A rate of 3.8 cm a\(^{-1}\) implies that the West Andaman and Sagaing faults are transform faults driven by seafloor spreading; 1.8 cm a\(^{-1}\) occurs when the West Andaman and Sagaing faults are transient faults driven by strain partitioning and/or drag by the northwards motion of India, and in their zone of overlap have caused a pull-apart basin to form. Indeed, published tectonic models interpret the West Andaman and Sagaing faults as the result of strain partitioning and drag associated with the northwards motion of India (e.g. McCaffrey et al. 2000; Maurin & Rangin 2009; Rangin et al. 2013). Conversely there is a paradox in accepting these tectonic models along with 3.8 cm a\(^{-1}\) spreading in the CAB. We suggest that the tectonic development of the CAB is not yet fully understood, and that further investigations are required to determine the nature of the crust in the Alcock and Sewell rises and even in the area presently thought to represent a spreading centre. The spreading rate of 3.8 cm a\(^{-1}\) determined for the last 2 Ma of activity by Raju et al. (2004) is faster than that required by strain partitioning models for the Sumatra–West Andaman–Sagaing faults. Sedimentation patterns within and adjacent to the central trough indicate that episodic extensional activity, not continuous spreading, has occurred. The episodic spreading of the Red Sea, where around 24 Ma extension shifted from the ridge axis to the continent–ocean transition zone before spreading renewed at c. 5 Ma (Almalki et al. 2014), is possibly an indication of how extension has changed location in the Andaman Sea with time.

Discussion

Understanding the dynamics of continental-scale deformation is a challenging task. Experimental studies can provide important insights, but natural processes necessarily operate on very different time and length scales. The Myanmar region is particularly interesting because a dramatic reorganization of deformation has occurred over the last 15 Ma, providing an example of a system evolving due to changing stresses and material properties. If these changes can be untangled, then it may be possible to gain a new quantitative insight into the parameters that control continental deformation. In the sections above, we have discussed the recent tectonic history of this region. This reorganization is summarized in Figure 2.21. The timing of many of these changes remains poorly constrained, and it is still not clear what processes have driven this reorganization.

One interpretation could be that these changes represent an ongoing westwards shift of the left-lateral shear associated with the eastern flank of the Indian indenter. Over the course of the collision, eastern Myanmar and the Indo-Burma Ranges have rotated clockwise due to the northwards indentation of the Indian Plate. Originally, shear may have been taken up by oblique slip on the more SE–NW orientated Indo-Burma Ranges boundary and right-lateral slip on the Wanding, Nanting and Mengliang faults (which would have trended closer to north–south), before being transferred to the Gaoligong Shear Zone in the north. This would have resulted in shortening in the area currently NE of the Wanding, Nanting and Mengliang faults. This clockwise rotation would have resulted in increasingly oblique subduction. This may have resulted in increased partitioning of shear strain away from the plate boundary. At the same time, the southeastwards extrusion of Tibetan material between the relatively rigid Sichuan Basin and central Myanmar lowlands would have resulted in the suppression of right-lateral slip on SW–NE strike-slip faults, due to compression from the NE. This may have triggered the reversal in the sense of slip of these strike-slip faults and the accommodation of right-lateral shear through clockwise rotations of left-lateral faults. Surface-wave tomography suggests that the thick lithosphere beneath Tibet extends to the vicinity of the northern boundary of these faults, but not further. The southeastwards flow of Tibetan material may also have controlled the reversal or cessation of activity on the Red River–Ailao Shan Shear Zone, which lies at a high angle to the direction of inferred crustal flow.

During the Pliocene some of the shear seems to have been focused onto the Sagaing Fault in central Myanmar. This requires significant shortening to be taken up to the north of the western sliver of the Myanmar Central Basin. It is not clear how this shortening is being accommodated, especially as the crust of both the central Myanmar lowlands and the Brahmaputra Valley are likely to be relatively strong. If the total displacement estimates of 300–400 km are taken as accurate, and these are far from certain, the present-day slip rate on this fault is likely to be slower than it was in the past. The difficulty of absorbing this deformation in the north may be driving the rapid right-lateral shear currently observed within the Indo-Burman Ranges in the GPS velocity field.

The GPS velocity field in the east shows that Tibetan material spreads outwards to the east and west after passing between the Eastern Syntaxis and the Sichaun Basin. This westwards component of motion could have contributed to the c. east–west component of shortening within the Myanmar Central Basin during the Late Miocene. The development of
the Shillong Plateau, and the associated clockwise rotation of the Brahmaputra Valley, may also represent a westwards shift in this shear, and may in part be driven by compression associated with the northern termination of the Sagaing Fault. The development of the Shillong Plateau may also have triggered an acceleration of deformation within the recently developed Tripura Fold Belt. There is no reason to believe that the Indian Plate has begun to move increasingly eastwards relative to the Sundaland during the last 2 Ma, and so the development of the fold belt may instead be the result of rapid deposition of large amounts of sediment in the flexural accommodation space, associated with the growth of the Shillong Plateau. This rapid sedimentation may have allowed the development of a very weak, shallowly dipping detachment, allowing gravity-driven deformation (associated with the topography of the Indo-Burma Ranges in the east) to trigger shallow folding. It is not clear if the rotation of the plate boundary may have led to the recent cessation of subduction beneath the northern Indo-Burman Ranges, and this remains controversial.

Conclusions and outstanding questions

(1) Much of the right-lateral shear associated along the eastern margin of India is localized on the Sagaing Fault. The rest of this shear is distributed across a wide area, including poorly known faults in the interior of the Myanmar Central Basin, major structures in the high Indo-Burman Ranges such as the Churachandpur-Mao Fault (which are currently only known in discontinuous sections) and beneath the Tripura Fold Belt. Many of the structures taking up this shear are not well described, and the possible role of oblique-slip on a subduction interface remains controversial.

(2) Estimates of the total geological offset on the Sagaing Fault are not currently based on secure pinning points. It has been suggested that the total offset may be on the order of 400 km; however, this would require a significant slowing of the fault slip rate over its history, or a much older age of initiation than is suggested by the timing of deformation in the Myanmar Central Basin. One possibility is that deformation on the Sagaing Fault began in the Late Miocene (as suggested by marine data on the southern extension of the fault), and that the structure took up a larger proportion of the overall right-lateral shear in the past.

(3) The location of the transition between continental and oceanic crust beneath NE India remains unknown. Extended continental crust appears to underlie the greater part of the Bengal Basin and may extend well beneath the Indo-Burman Ranges. Oceanic material must be present in the steeply dipping BSZ east of the Indo-Burman Ranges, due to the existence of recent calc-alkaline volcanism above the downgoing slab. While most workers assume that the ocean–continent transition occurs west of the Indo-Burma Ranges at the hinge-line shown in Figure 2.12, here we suggest (in line with Rangin et al. 2013) that thinned continental material may extend beneath most of the Indo-Burman Ranges where the Indian Plate remains shallowly dipping.

(4) There are large uncertainties concerning the seismic hazard in Bangladesh and easternmost India. Recent work suggests that there may be a megathrust interface elastically accumulating strain beneath the Tripura Fold Belt. Significant subduction, although oblique, most likely does occur on the southern Arakan section. Further north, where overall Indian Plate motion suggests little convergence, there is nevertheless significant east–west shortening as observed in local GPS studies. There is an urgent need to better understand how deformation is taken up in this region.

(5) Regardless of the status of the plate boundary, the unusually high seismogenic thickness of the underlying Indian Plate means that it is capable of failing in extremely large events. Such events have most likely occurred in the past, but the high sedimentation rates of the area mean that it is very difficult to constrain this hazard using surface observations.
(6) The tectonics of this region have changed dramatically over the last c. 10–20 Ma, providing an opportunity to test theories about how actively deforming areas evolve. More work is needed to determine the precise timing of these changes. The relative importance of the changing distribution of gravitational potential energy and the stresses associated with the Indian Plate boundary remain controversial. The ongoing extrusion of material out from the SE margin of Tibet may be a principal control on these changes.

(7) Complex deformation processes are operating offshore. In the Gulf of Martaban we need to understand how the fault and fold pattern in the Miocene–Recent sediments has evolved as an interaction of basement-involved rifting, strike-slip movement and gravity-driven deformation as a consequence of rapid sediment loading. In the Andaman Sea the timing of seafloor spreading and extent and age of oceanic crust remains uncertain.

(8) A simple strike-slip faulting model which relates pull-apart in CAB with the Andaman–Sagada Fault Zone predicts that the Sagaing Fault should accommodate most of the 3.8 cm a−1 differential motion between India and SE Asia. The reality (from GPS observations) is that the Sagaing Fault accommodates only half the motion, and the rest of the motion is distributed north of the spreading centre. This indicates that the simple model does not work. The proposal that the main timing of spreading was Middle–Late Miocene (Morley & Alvey 2015) fits better with the distributed strike-slip deformation observed onshore than a Pliocene–Recent spreading model.

References


