Microseismicity and seismotectonics of the South Caspian Lowlands, NE Iran

Majid Nemati, James Hollingsworth, Zhongwen Zhan, Mohammad Javad Bolourchi and Morteza Talebian

SUMMARY
This paper is concerned with the microseismicity and seismotectonics of the eastern South Caspian Sea region, where the East Alborz mountains descend to meet the South Caspian Lowlands of NE Iran. To better understand the present-day tectonics and seismicity of this region, which includes the cities of Gorgan and Gonbad-e-Qabus (combined population 500 000), we installed a temporary local seismic network across the area for 6 months between 2009 and 2010. We analysed the seismicity and focal mechanisms together with data from the permanent networks of the Institute of Geophysics, University of Tehran (IGUT) and the International Institute of Earthquake Engineering and Seismology (IIEES), based in Tehran. Microseismicity is focused primarily on the Shahrdud fault system, which bounds the east Alborz range to the south. Relatively few earthquakes are associated with the Khazar thrust fault, which bounds the north side of the range. A cluster of shallow microseismicity (<15 km depth) occurs 40 km north of the Khazar fault (within the South Caspian Lowlands; SCL), an area typically thought to be non-deforming. This area coincides with the location of three relatively deep thrust earthquakes (Mw 5.3–5.5) which occurred in 1999, 2004 and 2005. Inversion of teleseismic body waveforms allows us to constrain the depth of these earthquakes at 26–29 km. Although significant sedimentation throughout the SCL obscures any expression of recent fault activity at the surface, focal mechanisms of well-located events from the shallow cluster of micro-seismicity show a significant component of left-lateral strike-slip motion (assuming slip occurs on NE–SW fault planes, typical of active faults in the region), as well as a small normal component. Inversion of timeslides for well-located events in our network yields a velocity structure for the region, and a Moho depth of 41 km. The pattern of deep thrust and shallow normal seismicity could be explained by bending of the rigid South Caspian crust as it underthrusts the East Alborz mountains and Central Iran. Late Quaternary reorganization of drainage systems in the SCL may be the result of shallow normal fault activity within the SCL.

Key words: Earthquake source observations; Seismicity and tectonics; Continental neotectonics; Tectonics and landscape evolution; Crustal structure; Asia.

1 INTRODUCTION

Iran is currently deforming as a result of the northward collision of Arabia with Eurasia. Despite a relatively good understanding of how the broad-scale deformation is currently accommodated throughout this wide (>1000 km) region (e.g. Berberian 1981; Jackson & McKenzie 1984; Jackson et al. 1995; Axen et al. 2001; Jackson et al. 2002; Vernant et al. 2004b; Walker & Jackson 2004; Allen et al. 2006; Copley & Jackson 2006; Talebian et al. 2006; Walpersdorf et al. 2006; Djamour et al. 2010; Hollingsworth et al. 2010a; Allen et al. 2011; Nissen et al. 2011), the detailed tectonics of many earthquake-prone regions throughout the country still remain poorly understood. This study focuses on the South Caspian Lowlands (SCL), a unique area in NE Iran which forms the southern margin of the relatively non-deforming, aseismic South Caspian block, yet also lies between the deforming Kopah Dagh and Alborz mountains to the east and south, respectively. The SCL and neighbouring areas have experienced many destructive earthquakes over the past few centuries (Fig. 1), which highlight their importance in accommodating regional shortening (Tchalenko 1975; Berberian...
Figure 1. Summary topographic map of NE Iran corresponding to the yellow star in the inset figure. Historical and instrumental seismicity for the Central and East Alborz, West Kopeh Dagh and South Caspian Lowland (SCL) are shown: historical earthquakes (white circles) are from Ambraseys & Melville (1982), black focal mechanisms are from the Global CMT catalogue, red circles are epicentres from the catalogue of Engdahl et al. (2006), blue circles are epicentres recorded by the Institute of Geophysics, University of Tehran (IGUT) seismic network \( (M_n > 4.0) \). The 1970 Karnaveh and 2004 Baladeh earthquakes are labelled in yellow. Yellow numbers show the dates for various earthquakes in the SCL region. Blue squares show major cities across the region; GeQ, Gonbad-e-Qabus. Black lines show various faults. Blue ellipse highlights the SCL region, which is the focus of this study.

& King 1981; Ambraseys & Melville 1982; Jackson et al. 2002; Hollingsworth et al. 2006; Shabanian et al. 2009b). The primary aim of this paper is to better understand the role of the SCL within the wider Arabia–Eurasia collision zone. Furthermore, with a combined population of >500 000 people in the cities of Gorgan and Gonbad-e-Qabas (Fig. 1), plus many more living in neighbouring towns and villages, and a regionally important power plant at Neka, any attempt to improve our understanding of the seismic hazard posed to this area by active faults is of fundamental importance.

Due to the relative lack of seismicity in the SCL, and very few geological outcrops across the region, a seismological study was undertaken to help clarify the subsurface tectonics. We investigated the microseismicity of the SCL region during a 6-month period (spanning 2009–2010) using a network of 10 medium band CMG-3ESP Guralp instruments, owned by the Geological Survey of Iran. This data was combined with 4 yr of seismicity data recorded by seismic networks of the Institute of Geophysics, University of Tehran (IGUT) and the International Institute of Earthquake Engineering and Seismology (IIIES) of Iran. Because the seismic moment release from micro-earthquakes is relatively small compared with larger earthquakes \( (M_w > 5.2) \), and they rupture small fault areas away from the major fault structures, they may not always provide a representative picture of the active tectonics for a region. The approach used in this study is to combine data from local and teleseismic seismicity, the crustal structure and tectonic geomorphology so we can provide a coherent picture of the active tectonics of the SCL region of NE Iran; where information from any one of those data sources alone might be ambiguous and inconclusive.

2 TECTONIC SETTING AND SEISMICITY

The 3000 m high Alborz mountains of Northern Iran have formed in successive tectonic events related to the collision of Central Iran with Eurasia in the Late Triassic (Cimmerian Orogeny) and the present-day collision of Arabia with Eurasia (Berberian & King 1981; Alavi 1996; Zanchi et al. 2006). Immediately north of the Alborz lie the SCL, which form a low lying area around the edge of the South Caspian Basin (SCB). This region is notable for its thick sedimentary deposits, which yield significant volumes of oil and gas (Devlin et al. 1999; Brunet et al. 2003). The SCB is bordered to the south and west by major thrust belts of the Alborz, Talysh and Kopeh Dagh, and to the north by a young subduction zone, the Apsheron-Balkhan Sill (Jackson et al. 2002). The Alborz mountains bound the SCB to the south, forming a boundary separating rocks with a Central Iran affinity from rocks with Eurasian origins (Stöcklin 1974). The SCB contains up to 20 km of sedimentary deposits, making it one of the deepest sedimentary basins in
Figure 2. Map showing the location of both permanent and temporary seismic networks used in this study. Black triangles are IGUT stations, purple squares are from the IIEES network, and red triangles are from our temporary network. Orange stars show mud volcanoes (Ansari & Bolourchi 2003) in the SCL region, which highlight the ductile nature of thick and shallow sedimentary sequences characteristic of this foreland basin setting. Thin blue line shows the Gorgan river. Grey arrows show the direction of fold axes, adapted from Hinds et al. (2007). The Khazar thrust fault is highlighted by a thick blue line; the Shahrud fault system is highlighted by a thick red line.

Despite the clear expression of active faults within the east Alborz region (Ritz et al. 2006; Hollingsworth et al. 2008, 2010b; Javidfakhr et al. 2011), there has been relatively little seismicity associated with these structures over the last 50 yr. The Alborz mountains are currently being thrust over the SCB by the Khazar fault, which runs along the northern edge of the range (Tatar et al. 2007). Although the Khazar fault is not clearly expressed in the Late Quaternary geomorphology, compared with other thrust faults in the desert regions of Iran (e.g. Fattahi et al. 2006; Hollingsworth et al. 2010a), recent slip on the central part of this fault is thought to be responsible for the 2004 Baladeh thrust earthquake (M, 6.4, event 5 in Fig. 1, Tatar et al. 2007). The focal mechanism for this event indicated N–S shortening across the range, while the after-shock distribution revealed a S-dipping fault plane which projected to the surface along the Khazar fault trace, strongly suggesting a dominantly thrust-style of deformation on the Khazar fault. Although no large earthquakes have ruptured the southern range-front of the eastern Alborz in recent times, various studies have revealed active left-lateral slip along major range-parallel faults within the southern half of the range (Ritz et al. 2006; Hollingsworth et al. 2008, 2010b; Nemati et al. 2011). Deformation of the east Alborz mountains appears to be partitioned onto NNW-directed shortening on the Khazar thrust fault and ENE–WSW left lateral shear on the Shahrud fault system (Wellman 1966; Jackson et al. 2002; Hollingsworth et al. 2008). Using campaign GPS velocities from either side of the East Alborz range (Vernant et al. 2004a,b; Masson et al. 2007; Djamour et al. 2010), Hollingsworth et al. (2008, 2010b) estimated a left-lateral shear of ~3 mm yr⁻¹ across the range (consistent with Quaternary rates of shear, see Javidfakhr et al. 2011 and Rizza et al. 2011). Reconciling seismicity data with GPS velocities
and active fault mapping, Jackson et al. (2002) and Hollingsworth et al. (2006, 2008) concluded the SCL has a westward component of motion, relative to both Central Iran and Eurasia.

The South Caspian interior and adjacent lowland areas, especially to the east (the region under investigation in this study), have experienced relatively few earthquakes both in recent times (Jackson et al. 2002), and historically (Ambraseys & Melville 1982)—for example, the 72-m-high pure-brick tower of Gonbad-e-Qabus, built in 1006 AD as a tomb for the sultan Qabus ibn Wushmgir remains in pristine condition. The most significant earthquake from the East Alborz region occurred within the southern East Alborz mountains 50 km south of Gorgan, where slip on the Astaneh fault resulted in the destruction of Damghan city in 856 AD (Ambraseys & Melville 1982; Hollingsworth et al. 2010b, and Fig. 1). Nevertheless, a large earthquake occurred ~25 km SE of Gorgan in 1985, and three mid-sized earthquakes occurred 20–40 km north of Gorgan in 1999, 2004 and 2005 (Fig. 1).

The 1985 October 29, Gorgan earthquake (Mw 6.2) is the largest instrumentally recorded event to have occurred in the region of our local network (Fig. 1, Priestley et al. 1994). In the 2 months following the 1985 main shock, five aftershocks occurred (Mw 4.3–5.0, NEIC catalogue). Assuming a typical fault slip-to-length ratio of $5 \times 10^{-3}$, the approximate fault dimensions given a seismic moment of $2.18 \times 10^{38}$ N m (Priestley et al. 1994) are ~11 km, with 0.6 m of slip. Based on the 13 ± 5 km depth for the Gorgan earthquake (Priestley et al. 1994), and the 30–35° dip of the south-dipping nodal plane (Priestley et al. 1994), the fault plane projects to the surface along the Khazar fault, which runs along the northern edge of the East Alborz mountains (see later discussion, and Fig. 9). Assuming this earthquake ruptured the south-dipping Khazar fault, the focal mechanism of Priestley et al. (1994) indicates NWW-directed shortening, perpendicular to the Alborz range-front, suggesting a dominantly pure-thrust style of deformation for this structure (earthquake 2 in Fig. 1).

The 1999, 2004 and 2005 events are notable for being relatively deep (~30 km; e.g. Global CMT and hypocentral relocations of Engdahl et al. 2006) and for their unusual location 30–50 km north of the range front within the generally aseismic SCL (Fig. 1)—most earthquakes in Iran are shallower than 20 km (e.g. Maggi et al. 2000; Jackson et al. 2002; Talebian & Jackson 2004; Nissen et al. 2011). Hollingsworth et al. (2008) suggested these deeper earthquakes may be related to flexural compression at the base of the SCL as it is underthrust beneath the Alborz, although they could also result from deep and blind thrust splays propagating north of the Alborz range front. Various kinematic models have been proposed for the region (Jackson et al. 2002; Hollingsworth et al. 2006, 2008, 2010a; Ritz et al. 2006; Shabanian et al. 2009a; Djamour et al. 2010; Shabanian et al. 2010; Javidfakhr et al. 2011) which show that Arabia–Eurasia shortening is accommodated by both thrust and strike-slip fault systems in the Koppeh Dagh and East Alborz ranges, with some component of westward extrusion of the Western Koppeh Dagh region between the right lateral Ashkabad fault and left-lateral Shahrud fault systems (Fig. 1). However, these models do not adequately address active deformation occurring within the SCL or West Koppeh Dagh regions, where large earthquakes such as the 1970 Karnaveh earthquake (Ambraseys et al. 1971) have occurred, and where recent fault movements are visible in the late Quaternary geomorphology. Northeast of Gorgan, the pattern of active faults becomes distributed, and obscured by extensive loess deposits and vegetation cover. Little is known about how faults slip in this area, and few faults have surface traces continuous over more than a few kilometres.

3 METHODS

In the following section, we first explain the various approaches we use to examine the seismicity of the SCL region using teleseismic, regional network and local network data. We then determine the velocity structure for the SCL by inverting traveltimes for well-located events in our network. The results from each analysis are outlined separately in Section 4.

3.1 Methods: teleseismic seismicity

While the 1985 Gorgan earthquake appears to have ruptured the Khazar fault (the most obvious active fault in this region), three additional earthquakes occurred in the SCL north of Gorgan between 1999 and 2005, which are less easy to associate with an active structure: 1999 November 19 (Mw 5.3, depth 30 km), 2004 October 7 (Mw 5.5, depth 32 km) and 2005 January 10 (Mw 5.4, depth 32 km)—see Figs 1 and 9. The ~30 km depths reported by the Global CMT and Engdahl catalogues (e.g. Engdahl et al. 1998, 2006) for these earthquakes are unusually deep for Iran, where depths in the plateau region are typically <20 km (Maggi et al. 2000). Deeper earthquakes have been reported from the Trans-Caspian zone, which strikes NW–SE across the central Caspian Sea, and are thought to result from subduction of the South Caspian beneath the North Caspian Sea (part of stable Eurasia, see Jackson et al. 2002). Since depth estimates from the Global CMT catalogue are often poorly constrained in the shallow crust (e.g. Maggi et al. 2002), we use both body-waveform inversion (Zwick et al. 1994) and Cut-And-Paste (CAP) methods (Zhao & Helmberger 1994; Zhu & Helmberger 1996) to determine the best-fitting source parameters of these earthquakes.

We use the MT5 version (Zwick et al. 1994) of the McCaffrey & Abers (1988) and McCaffrey (1991) algorithm, which inverts the P and SH waveform data to obtain the strike, dip, rake, centroid depth, seismic moment and source time function of the 1999, 2004 and 2005 earthquakes. The method and approach we use are described in detail elsewhere (Nábělek 1984; McCaffrey & Nábělek 1987; Molnar & Lyon-Caen 1989; Taymaz et al. 1991). The CAP method first cuts teleseismic records into P-wave segments in the vertical components and SH-wave segments in the transverse components, and then fits them to synthetic waveforms independently allowing different time shifts and weights. We perform a grid-search of magnitude, depth, source duration, strike, dip and rake to find the optimal source parameters for the 2004 event with the smallest waveform misfit. Uncertainties are calculated using the bootstrapping method.

3.2 Methods: regional network seismicity

We examined 4 yr of seismicity data recorded by the regional seismic networks of the IGUT and the IIIES of Iran. The IGUT permanent network instruments are SSI short period Nanometrics instruments each consisting of three components. The IIIES seismometers are mainly broadband Guralp CMG-3T sensors. We selected 1389 well-located events from 3941 earthquakes recorded by the IGUT network. About 4800 events were also recorded by the IIIES network between 2004 and 2010. Fig. 2 shows the locations of the IGUT and IIIES stations throughout the East Alborz region. We only consider events, which were recorded by at least four stations. To get a picture of the overall seismicity in the East Alborz and SCL region, we did not select the epicentres based on their maximum azimuthal gap. We filtered the data to include only the best-located events, thereby improving the location of the diffuse seismicity (Bondár et al. 2004;
3.3 Methods: local network microseismicity

To better constrain the recent microseismicity of the SCL region, a temporary network of 10 seismometers was deployed in a 100 km × 100 km region north of the city of Gorgan (Fig. 2). The instruments installed were medium-band sensors with a cut-off period of 60 s, each connected to a DM-24 digitizer. The local network was operational between 2009 October and 2010 January, while the timing of IGUT and IIIES data used in this study ranged from 2006 to 2010 and 2004 to 2010, respectively. Fig. 2 shows the location of the temporary network stations. The instruments recorded the waveform data in continuous mode, and at a sampling rate of 100 Hz. The temporary stations were visited each week for maintenance (e.g. checking of power supply and consistency of internal versus external GPS time). The appropriate horizontal amplitudes were picked to calculate a \( M_f \) equivalent Wood-Anderson magnitude. The waveforms were edited manually using the Seisam software (Guralp systems LTD), and the processing was performed using the Seisan software (Havskov & Ottemøller 2005) and Hypo71 program (Lee & Lahr 1972).

3.4 Methods: velocity structure of the SCL

To better constrain the crustal structure of the SCL region, we estimated the \( V_p \) to \( V_s \) ratio from the average \( t_s/t_P \) versus \( t_P-t_0 \) of a selected set of 32 earthquakes recorded with our local network, obtaining a \( V_p/V_s = 1.70 \). Each earthquake was recorded by a minimum of four stations, with a maximum azimuthal gap of 180°, a maximum epicentral distance of 150 km, RMS less than 0.3 s, standard deviation of 0.03 and minimum correlation coefficient of 0.98 (Nemati et al. 2010).

Due to the lack of a \( P \)-wave velocity structure in this area, we used the shallow velocity structure of Mangino & Priestley (1998) as an initial model. We determined the crustal velocity model by inverting local traveltimes of reliable events for a 1-D velocity structure (Kissling 1988). Only events recorded with a minimum of eight phases, rms less than 0.5 s and both horizontal and vertical errors less than 5 km were used in the inversion (totaling 47 events). Because there is a non-uniqueness and correlations between the input and output models, we investigated 50 randomly distributed initial models to avoid bias in the estimated structure. The inversion process was divided into two stages. First, all the initial models consisted of 20 layers with a mean velocity of 6.0 km s\(^{-1}\) and a thickness of 2.0 km from the surface to 40 km depth, and we allowed for a maximum change of 0.5 km s\(^{-1}\) for each layer (e.g. Nemati et al. 2011). This allowed us to estimate the approximate depths and velocities of the real discontinuities in the velocity structure, which results in a three layer model with velocity contrasts located at 6 km, 12 km (upper crystalline layer) and 20 km depth, overlying a half-space.

4 RESULTS

4.1 Results: teleseismic seismicity

The best-fitting source parameters determined for the 1999, 2004 and 2005 earthquakes using the MT5 body-waveform modelling software are shown in Fig. 3. The best-fitting depths for each event are: 1999, 26 km; 2004, 28 km and 2005, 29 km.

The best-fitting source parameters for the 2004 earthquake, determined using the Cut-and-Paste method, are shown in Fig. 4. Figs 4(a) and (b) show the fit between the raw data and synthetic waveforms for both \( P \) and \( SH \) waves for our best-fitting source model (\( M_s = 5.5 \), depth 32 km, strike 26°, dip 44°, rake 53°). Figs 4(c)-(i) show the uncertainties in our best-fitting solution. Although the rupture duration is not well constrained, we find it does not trade-off with depth (Fig. 4c), forming a well-defined minimum on the misfit curve. Although the azimuthal station distribution is not optimal, our bootstrapping error estimations give consistent results, suggesting both the focal mechanism and depth are well resolved. The best-fitting depth of 31–33 km is similar to the 28 km depth estimate determined using MT5 (Fig. 3).

4.2 Results: regional seismicity

The distribution of seismicity in the SCL for the two different regional networks used in this study is shown in Fig. 5a and b. The earthquake epicentres of the IGUT network (\( M_b \) of 0.5–4.5) were plotted from 2006 to 2010, while the epicentres of the IIIES network are from 2004 and 2010.

The majority of seismicity is focused on the left-lateral Shahrud fault system of the southern Central and East Alborz, and in the SCL north of Gorgan, in the same area as the 1999, 2004 and 2005 earthquake (Figs 1 and 5). Relatively little seismicity is associated with the Khazar fault, which strikes along the northern margin of the Alborz range, and ruptured during the 2004 Baladeh earthquake (Fig. 1, Tatar et al. 2007) and probably during the 1985 Gorgan earthquake (Priestley et al. 1994).

4.3 Results: local network seismicity

Of the 900 earthquakes (\( M_l \) range of 0.5–4.0) recorded by the temporary network, 280 events have been well-located, of which 70 are high quality, and 210 are lower quality (Fig. 6a). The depth distribution of both high and lower quality earthquakes recorded by the local network between October 2009 and January 2012 shows the majority of earthquakes occurring in the top 14 km, although earthquakes occur all the way down to the Moho at 41 km depth (Fig. 6b). Table 2 gives the nearest station distances, the number of \( P \) and \( S \) readings and azimuthal gaps used to determine the depths for each micro-earthquake.

The relatively good station covering of our local network, and also the IGUT and IIIES surrounding stations (Fig. 2), especially south of Gorgan, allows us to compute 20 first motion micro-earthquake focal mechanisms (Fig. 7 and Table 3). Based on the number of polarities, maximum rotation of the planes and polarity scattering over four quadrants, we split the focal mechanisms into two groups: black (higher quality) and grey (lower quality). In nearly all cases, the steeper NE–SW-striking nodal planes are relatively well constrained by the station coverage, compared with the NW–SE planes. The distribution of first motions and the nodal planes for each earthquake can be found in Appendix A. Depth errors are typically <5 km (Nemati et al. 2011). Focal mechanisms: 2, 3, 4, 5, 6, 11, 14, 16, 18, 19 and 20 have a left-lateral, and in some cases a small normal component of slip, assuming the NE–SW nodal plane represents the fault which ruptured. The mechanisms of 4 and 7 may be related to left lateral shear on fault systems in the region of Karnevah and Kara Kala (Figs 10 and 11). Focal mechanisms 12 and 13 show a deep strike-slip (44.3 km) and thrust (40.9 km) mechanism, respectively, although both are not well constrained—nevertheless,
Figure 3. $P$ and $SH$ waveforms for the 1999, 2004 and 2005 earthquakes in the SCL of NE Iran. The event header shows the strike, dip, rake, centroid depth and scalar seismic moment (in Nm) of the best-fitting solution. The upper focal sphere shows a lower hemisphere stereographic projection of the $P$ waveform nodal planes, along with positions of the seismic stations used in the inversion. The lower focal sphere shows the $SH$ nodal planes. Capital letters next to the station codes indicate their position on the focal sphere, and are ordered clockwise by azimuth, starting at north. Solid lines show observed waveforms, and dashed lines show synthetics. The inversion window is marked by vertical lines on each waveform. The source time function (STF) is shown, along with time scale for the waveforms. The amplitude scales for the waveforms are shown below each focal sphere.
Figure 4. (a) Best-fitting fault plane solution for the 2004 October 7 earthquake determined using the Cut-and-Paste method. Red crosses show station distributions used in the inversion. (b) Teleseismic waveform fitting and the corresponding source parameters for the best-fitting solution. Black is the data and red is the synthetic. Both $P$ and $SH$ waveforms are used in the inversion. (c) Misfit function for a range of depths and durations for the 2004 event. Different durations cause very little difference in waveform misfit, and therefore have little trade-off with depth. All misfit curves for different durations show well-defined minima at 31–33 km. The solution outlined in red shows the optimal focal mechanism. Solutions outlined in black show 1000 bootstrapping results. Red and blue dots show $P$- and $T$-axes for all 1000 focal mechanisms. (d, e, f) Optimal focal mechanism (red dots) and the 1000 bootstrapping focal mechanisms (small green dots) plotted over the misfit function. (g, h, i) Histograms of strike, dip and rake from the 1000 bootstrapping solutions. The titles show corresponding means and standard deviations for strike, dip and rake.
both events are within error (±5 km) of the Moho (41 km), and therefore probably occur in the base of the crust. Mechanisms 5 and 8 may reflect left-lateral oblique shortening on the Khazar fault, with similar depths as aftershocks from the 2004 Baladeh earthquake in the Central Alborz (Tatar et al. 2007). The red solution in Fig. 7 shows the focal mechanism for the 1985 Gorgan earthquake (Priestley et al. 1994), which probably ruptured the south-dipping Khazar fault.

4.4 Results: velocity structure of the SCL

Table 4 shows our initial and final velocity models, calculated by inversion of travel times. At depths shallower than the Moho, the range of velocities is between 6 and 6.7 km s\(^{-1}\) (Nemati et al. 2010).

Because the majority of the events are located shallower than 20 km, we could not estimate the velocity for the deeper layers. We
Table 1. Source parameters of significant instrumentally recorded earthquakes occurring in NE Iran over the last 30 yr (Fig. 1). Except where constrained by body-waveform modeling, locations and depths for all events are taken from Engdahl et al. (2006, prior to 2005), and the updated catalogue of Engdahl et al. (1998, after 2005). Data sources are as follows: 1, Priestley et al. (1994); 2, this study; 3, Global CMT catalogue; 4, Tatar et al. (2007).

Table 2. Nearest station distances, number of P and S readings and azimuthal gaps used to determine the depths for each micro-earthquake.

Therefore used traveltimes of refracted waves from distant events to estimate the depth and velocity structure deeper than 20 km. Using 946 traveltimes recorded in the IGUT network, we computed a P-wave velocity of 8.2 km s\(^{-1}\) for the upper mantle with a Moho depth of 41 km. This velocity is consistent with the 8.2–8.5 km s\(^{-1}\) velocity estimated from Pn tomography for this region (personal communication, Abdooreza Ghods, 2012). The computed average velocity (≈6.5 km s\(^{-1}\)) of the upper crust, shown by the slope of the
Figure 7. Topographic map showing fault plane solutions for 20 micro-earthquakes recorded by our local network. Each event was recorded by at least five stations, with an azimuthal gap less than 180° and horizontal and depth uncertainties less than 5 km (see Table 3 for details). Black solutions are higher quality, grey solutions are lower quality. Red solution is for the 1985 October 29 Gorgan earthquake (Priestley et al. 1994).

Table 3. Source parameters of 20 micro-earthquakes for which we were able to generate reliable first-motion focal mechanisms (Fig. 7)—see also Fig. A1. “∗” Denotes lower quality focal mechanisms.

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5 DISCUSSION

5.1 Summary of results

In this study we use both teleseismic and local network seismicity data to investigate the subsurface tectonics of the SCL region of NE Iran. Data from our local seismic network is used to compute the ratio of \( V_p/V_s \) (about 1.70), while inversion of local traveltimes from reliable events yields a velocity range in the crust of lower left fitted line in Fig. 8, is consistent with the result of the 1-D inversion. This is not a unique method of Moho determination; the refraction range in Fig. 8 indicates \( \sim 3 \) km uncertainty in the Moho depth. Furthermore, the refracted traveltimes of distant events is sensitive only to horizontal discontinuities. Nevertheless, despite location uncertainties of \( \sim 5 \) km, and possible artifacts due to lateral heterogeneities for the regionally recorded events this method is efficient for resolving shallow discontinuities (\(<45 \) km).
Table 4. Starting velocity structure (Mangino & Priestley 1998), and final velocity structure computed using locally recorded events.

<table>
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<th>Initial velocity model</th>
<th>Final velocity model</th>
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<td>Velocity (km s⁻¹)</td>
<td>Depth (km)</td>
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<tr>
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6.0–6.7 km s⁻¹, consistent with other studies in the area (Mangino & Priestley 1998). Using traveltimes of refracted waves from distant events, we obtain a velocity of 8.2 km s⁻¹ for the upper mantle Pn wave velocity and a Moho depth of 41 km.

For the Tehran region of the Central Alborz, Ashtari et al. (2005) computed a Moho depth of 34 km and 8.0 km s⁻¹ for the P-wave velocity in the upper mantle using locally recorded micro-earthquakes. Sodoudi et al. (2009) found deeper Moho depths beneath the Central Alborz (51–54 km), using P- and S-receiver function methods. However, a more recent study by Radjaee et al. (2010), which uses simultaneous joint inversion of teleseismic receiver functions with fundamental mode Rayleigh wave group velocity measurements across the entire Central Alborz, gives even deeper Moho depths of 55–58 km beneath the range, 48 ± 2.5 km immediately south of the Alborz (i.e. Central Iran), and 46 ± 2.5 km immediately north of the Alborz (i.e. SCL). Using a partitioned waveform inversion method, Shad Manaman et al. (2011) image the S-velocity structure of the Iranian Plateau, and find similar Moho values as Radjaee et al. (2010) for the Central Alborz of 55–60 km, although they obtain significantly lower values of 35–37 km for the West and East Alborz regions. One explanation for the different estimates of Sodoudi et al. (2009) and Radjaee et al. (2010) is that Sodoudi et al. (2009) use the IASP91 earth model (Kennett & Engdhal 1991) to convert the Moho Ps delay times to an interface conversion depth. The IASP91 model is faster (average Vp 3.81 km s⁻¹) than the average velocity Radjaee et al. (2010) find for the central Alborz region (Vp 3.59 km s⁻¹) resulting in a 3–4 km deeper depth estimate. Based on the results of Radjaee et al. (2010) and Shad Manaman et al. (2011), the Alborz appear to have a moderate crustal root but of insufficient thickness to compensate the elevation of the range, where numerous peaks rise above 3000 m. The analysis of free-air gravity shows that the elevation of the Alborz Mountains is largely supported by the elastic strength of the Iranian Plate, the South Caspian Plate or both (Radjaee et al. 2010). Their Moho depth estimate of 41 km for the SCL region north of Gorgan is therefore broadly consistent with the 46 km estimate of Radjaee et al. (2010) for the SCL north of the Central Alborz range. However, it raises questions regarding the depth of the Moho beneath the East Alborz mountains south of Gorgan, where Nemati et al. (2011) computed a P-wave velocity for the upper mantle of 8.0 km s⁻¹ and a Moho depth of 34 km using 458 traveltimes recorded in the IGUT network, and Shad Manaman et al. (2011) obtain a Moho depth between 35 and 37 km.

The crustal velocity model and depth distribution of micro-earthquakes suggests a sedimentary cover of ~6 km, similar to the 7 km depth in the NE Alborz region, calculated using receiver functions analysis (Mangino & Priestley 1998). Our estimates are less consistent with the 10 km estimate of Tatar et al. (2007) for the north-central Alborz, which ruptured in the 2004 Baladeh earthquake. Although the majority of micro-earthquakes occur in the upper 15 km of the crust, events occur all the way down to the Moho at 41 km depth, suggesting the South Caspian crust is strong throughout.

Earthquake epicentres recorded between 2004 and 2010 by the IGUT and IIEES regional networks are focused on the left-lateral Shahrud fault system, within the southern half of the East Alborz mountains south of our study area. Activity on the Khazar fault is relatively low, although a large earthquake probably ruptured this section of the fault in 1985. Three mid-sized earthquakes, with relatively deep depths for Iran (~30 km), occurred north of the Khazar fault within the stable SCL in 1999, 2004 and 2005. A small cluster of micro-earthquakes recorded between 2009 and 2010 also coincide with these deep earthquakes. These events occur mostly at shallow depths, and focal mechanisms for the largest events indicate predominantly strike-slip motion (probably left-lateral on NE–SW-oriented planes), with a small normal component in the shallower events.

5.2 Variation of deformation-style with depth in the SCL

A cross-section through the SCL crust is shown in Fig. 9, which helps illustrate the subsurface deformation occurring at the present day. Earthquakes larger than Mw 5.2 are shown in blue (Global CMT) and red (Priestley et al. 1994). Grey and black solutions are micro-earthquakes (Ml1.9–4.8) recorded by our local network. Earthquakes south of the Alborz range front project from depths of ~25 km to the Khazar fault, which strikes along the edge of the range front, and is poorly expressed in the geomorphology in this region. Seismicity within the SCL north of the range front is broadly grouped in deep thrust earthquakes (30–41 km), and shallow left-lateral micro-earthquakes with a small normal component (~<20 km). These shallow events lie along-strike from known left-lateral strike-slip faults to the NE (e.g. at Karnaveh, Fig. 10, and Kara Kala, Fig. 11). Therefore, they may represent re-activation of similar structures which have been buried by sediments within the SCB. Alternatively, they could simply be shallow earthquakes responding to stress changes after the earlier 1999, 2004 and 2005 events. Moho depth estimates in this region (Mangino & Priestley 1998, and this study) increase from 35 to 41 km between distances of ~100 km and ~50 km north of the Alborz range front (Fig. 9c). Consequently, the 1999, 2004 and 2005 events all occur near the base of the crust, implying the South Caspian crust is strong throughout (see also Jackson 2002). Therefore, the pattern of deep thrust and
shallow normal earthquakes may result from bending of the South Caspian crust beneath the over-riding Alborz mountains and Central Iran. Thus, the Alborz mountains are supported, at least in part, by the strong South Caspian crust (e.g. Radjaae et al. 2010). The separation by depth of shallow normal and deep thrust events associated with bending is common in the oceans, and seen on the outer rises of many subduction zones (e.g. Chapple & Forsyth 1979). However, it much less common in the continents, with the clearest examples coming from Central Asia, where northern India (Jackson 2002) and the Tarim basin (Sloan et al. 2011) underthrust Tibet. Nevertheless, the transtensional rather than normal mechanisms of the shallow micro-earthquakes recorded by our local network could result simply from the westward motion of the South Caspian block relative to Central Iran and Eurasia, rather than a response to bending of the rigid South Caspian crust beneath the Alborz (JavidiFakhir et al. 2011). Furthermore, despite deep thrust events having occurred in this region, there have been no recorded large normal earthquakes, as might be expected with such a bending mechanism. Therefore, it is also possible the 1999, 2004 and 2005 events are related to thrust faults stepping northward of the Alborz range front. Similar basinward migration of thrust faults is seen elsewhere in the East Alborz and Sabzevar mountains (Hollingsworth et al. 2010a), elsewhere in Iran (Berberian et al. 2001; Walker 2003), and is typical of older continental thrust systems such as the Alps and Himalayas (Boyer & Elliott 1982).

5.3 Tectonic geomorphology: observations of the West Kopeh Dagh region

Although micro-earthquakes may not always provide a clear picture of the regional tectonics, due to their relatively small contribution to the regional seismic moment release, compared with larger earthquakes (e.g. $M_w > 5.2$), these events lie along strike from NE–SW left-lateral fault systems to the east, near Kavnaveh and Kara Kala, which are known to be seismogenic. On the 1970 July 30, the Kavnaveh earthquake ($M_w 6.4$) occurred ∼100 km NE of the area covered by our local network, where the Western Kopeh Dagh mountains drop down to the meet the SCL (Fig. 1). The earthquake caused widespread destruction to villages in the region, killing ∼200 people (Ambraseys et al. 1971). No clear surface rupture was reported, although the rupture may not have propagated through the extensive loess deposits covering the area, or may have degraded rapidly after
the earthquake (Fig. 10, see also Fig. 1 for location). The worst affected villages of Baba Shamalak (25 per cent of population killed), Karnaveh (<10 per cent) and Goldagh Shahrak (<10 per cent) span a distance of 30 km. Four subparallel NE–SW faults lie within this region of maximum destruction (Figs 10a and b), although only the eastern fault is clearly visible in satellite imagery; the other two are obscured by vegetation, and are inferred from lineations visible in the topography (Fig. 10c). One of these faults runs SW of Baba Shamalak, which lies 10 km east of the teleseismic epicentral location of Engdahl et al. (2006), and is visible as a small ridge uplifting Quaternary river deposits to the east (Fig. 10c). The orientation and sense of uplift is consistent with the left-lateral nodal plane in the solution of Priestley et al. (1994, see Fig. 10a), which also has a small normal component (down to the west) and Jackson & Fitch (1979), who recorded aftershocks aligned in a NE–SW direction. The two eastern faults, shown in Fig. 10(d), have similar lengths to the fault west of Baba Shamalak. Drag of the bedrock geology into the fault zone is consistent with left-lateral motion (3 km minimum offset estimated from deflection of a fold axis, white dotted line in Fig. 10d).

Another system of active left-lateral faults lies 50 km north of Karnaveh, near the town of Kara Kala in Turkmenistan (Fig. 11). Although the rocks offset Mesozoic geology, small river systems have been deflected across the fault against their drainage direction, suggesting they are active at the present-day. Active left-lateral shear on NE-SW striking faults both at Karnaveh and Kara Kala may partly...
accommodate the westward motion of the South Caspian/Western Kopeh Dagh regions (relative to Eurasia and Central Iran), in a similar manner to the Shahrud fault system to the south (Hollingsworth et al. 2008).

5.4 Tectonic geomorphology: expression of flexural bending in the SCL

A small topographic high occurs in the SCL, at the eastern end of the zone of micro-earthquakes recorded by our local network (Fig. 9). At this location (N37.25°E57.75°, see Fig. 12), a number of streams (dark green lines) appear to drain south into the west-draining Gorgan river (heavy blue line). A topographic profile across this area indicates the south draining streams (green lines) actually lie on a north-facing slope. Some of these streams have even reversed their flow direction (red lines), joining up with another west draining river to the north. Thus, a subtle topographic high appears to have formed in the late Quaternary, disrupting the drainage in this relatively flat area. One possibility is that, over time, buried left-lateral faults slipping with a small normal component have produced a small amount of footwall uplift, consistent with the focal mechanisms of micro-earthquakes in this region. If the faults could not propagate all the way to the surface, through the weak sedimentary cover, footwall uplift may have warped the overlying sediments to form a topographic high. Any diffuse extension occurring in the crest of the uplifted area may have been excavated by the Gorgan river. Repeated slip at depth would result in uplift of the area, and incision of the Gorgan river to maintain its current course. With continuing uplift, streams draining southward into the Gorgan river will have been back-tilted and abandoned. Figs 12(c) and (d) show two cartoons illustrating this concept. Although this hypothesis could potentially explain the formation of this bulge in the geomorphology, its lies only 25 km north of the range front, which may be too close to be caused by bending. In the neighbour- ing Kopeh Dagh mountain range, several hundred kilometres to the northeast, bending of the Turkmen foreland extends up 100 km north of the range (Maggi et al. 2000). The Kopeh Dagh foreland is unlikely to be significantly different from the East Alborz. An alternative explanation is that the normal component in the various micro-earthquakes could result from the overall westward motion of the SCB relative to NE Iran (Jackson et al. 2002; Javidfakhri et al. 2011), although this would not necessarily explain the depth separation of normal and thrust mechanisms. Uplift could also result from slip on a north-dipping thrust fault, although the decreasing elevation to the north of this area, and the proximity to the major south-sipping Khazar fault make this explanation unlikely.

6 CONCLUSIONS

Using a combination of data on the local and regional seismicity, crustal structure and tectonic geomorphology of the SCL region of NE Iran, we provide a relatively coherent picture of the active tectonics of this little-studied region. Inversion of earthquake travel times yields a velocity structure for the SCL crust, and a Moho depth of 41 km. Micro-earthquakes occur down to 41 km, while three thrust earthquakes in 1999, 2004 and 2005 occurred at depths of ~30 km, implying shortening near the base of the South Caspian crust. Oblique-slip (left-lateral and normal) micro-earthquakes occur in the shallow crust (<20 km) above the deeper thrust events. If micro-earthquakes in this area are representative of the

Figure 11. (a) Landsat7 satellite image of the western Kopeh Dagh mountains near the town of Kara Kala (Turkmenistan, see Fig. 5d for location). Pale grey lines show various left-lateral faults (see inset map for clearer image of active faults in this area). (b) SPOT satellite image (GoogleEarth) showing left-lateral displacement of Mesozoic geology across one of the left-lateral faults located 20 km SW of Kara Kala. (c) Quickbird satellite image showing left-lateral deflection of a river, and its associated terrace deposits, 5 km SW of Kara Kala.
deformation produced by larger earthquakes, left-lateral shear, related to the westward extrusion of the SCB, may be accommodated within the SCL, north of the Alborz and Khazar fault. The pattern of shallow transtention and deep shortening may also be produced as the rigid/elastic South Caspian crust is thrust beneath the Alborz mountains, which it partially supports. Transtentional events may result from re-activation of pre-existing buried left-lateral faults, common in the region. Significant sedimentation and dense vegetation within the SCL obscures much of the fault-related geomorphology. Nevertheless, a small component of footwall uplift possibly associated with shallow events (assuming it is persistent over thousands of years), may be at least partially preserved in drainage systems throughout the area.

ACKNOWLEDGEMENTS

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APPENDIX

The distribution of first motions and the nodal planes for each earthquake is shown in Fig. A1.
Figure A1. Focal mechanisms showing the distribution of first motions and the nodal planes for micro-earthquakes 1–20. Each plot is shown in a stereographic projection.