Deformation flow analysis and symmetry of Goushti shear zone, Sanandaj-Sirjan metamorphic belt, Iran

Babak Samani
Faculty of Earth Sciences, Shahid Chamran University of Ahvaz, Ahvaz, Iran.
**Corresponding author, e-mail: samani.babak@gmail.com*

(*received: 01/11/2016 ; accepted: 08/04/2017*)

Abstract

Finite strain and vorticity analyses were carried out in the deformed conglomerate and micro-conglomerate of Goushti shear zone in the Sanandaj-Sirjan metamorphic belt (Zagros Mountains Iran). These deformed rocks are bounded between two northeastern and southwestern major thrust faults which are parts of the Zagros Thrust System. Finite strain measurements on three principal planes of strain ellipsoid show the higher amounts of strain ratios near the major thrust faults. Strain ellipsoid shape (K-value) and strain intensity (D-value) of the Flinn diagram reveal more prolate shape of strain ellipsoid and more amount of strain intensity around the major thrust faults. According to the amounts of natural logarithmic strain (ε), mylonite and protomylonite rocks have been formed in lower and higher distances from the main basal and dorsal faults respectively. Kinematic vorticity (W_m) analysis reveals an important strain partitioning in the Goushti shear zone. Results show simple shear dominant deformation near the major thrust faults and pure shear dominant deformation far from them. The amounts of shortening were determined between 0.37 and 0.62 which shows shortening between 38 to 63%. According to the angle between lineation and intersection of the foliation and shear zone boundary (χ), triclinic symmetry was determined for the shear zone.

Keywords: *Finite Strain, Kinematic Vorticity, Triclinic Symmetry, Zagros, Iran*

Introduction

Faults or shear zones that are not purely strike-slip and include additional perpendicular components of shortening or extension known as transpression and transtension areas (Tikoff & Fossen, 1995). The boundary-parallel simple shear component in transpression/transtension zones can be horizontal (transcurrent transpression/transtension) or oblique (oblique transpression/transtension) (Robin and Cruden, 1994; Lin *et al.,* 1998). Oblique-slip is commonly observed in present plate boundary regions and strain geometry in many ancient shear zones suggests possible oblique relative slip at the boundaries (Jones *et al.,* 2004). Most classical shear zone models (e.g. Ramsay & Graham, 1970; Ramberg, 1975) and some transpression models (e.g. Sanderson & Marchini, 1984) have monoclinic symmetry where the rotation axis of the finite deformation or the vorticity vector of the instantaneous flow is set to be parallel to one of the principal strain or principal strain rate axes. A common feature of most transpressional models is that they have used a basic symmetry of an upright zone deformed between vertical zone boundaries. Non-vertical zones were considered briefly by Dutton (1997), who termed such deformation 'inclined transpression'. Inclined transpressional models are applicable to most natural collisional plate margins and many other shear zones in the crust, which are often non-vertical (Jones *et al.,* 2004). In the last twenty years several field, theoretical and modelling studies have been carried out in order to unravel the complex 3D finite and incremental deformation that controls the development of geological structures during oblique convergence (Harland, 1971; Sanderson and Marchini, 1984; Richard and Cobbold, 1990; Fossen & Tikoff, 1993; Tikoff & Teyssier, 1994; Jones & Tanner, 1995; Dutton, 1997; Holdworth *et al.,* 2002; Jones *et al.,* 2004; Sullivan and Law, 2006). Oblique convergence can be described in terms of homogenous deformation and simultaneous superposition of pure and simple shear with a variable geometrical distribution of the fabric attractors and repulsors that have either a monoclinic symmetry or a more general triclinic one (Iacopini *et al.,* 2007). The finite configuration of the deformation can be described by an ensemble of geometric measurable quantities such as foliation and lineation attitudes, finite strain ellipticity and shear strain, their gradients across the shear zone. By means of these quantities, assuming that the finite configuration records of the whole deformation history (Horsman and Tikoff, 2007) it is possible to define the main characteristics of the deformation. This research mainly consists of strain analysis in the deformed conglomerate rocks in the Sanandaj-Sirjan metamorphic belt (SSMB) for determination of deformation flow, strain partitioning and symmetry of deformed area. The study area in this research (named as Goushti shear zone because of the nearest Goushti village) is a part of Dehbid shear zone or Koli-kosh shear zone complex that has been studied by several researchers (Haji Hosseinlou, 2003; Hosseni, 2009). Previous kinematic, structural and petrofabric studies reveal an important dextral sense of shear in this deformed area. There are a few studies in the realm of strain and vorticity analysis in different parts of this area. The most important principle in the strain and vorticity analyses is finding of proper markers that help geologists. The key to strain analysis lies in finding objects with known initial packing arrangement or features which enable final lengths or angles to be calculated. Therefore in this research pebbles of deformed conglomerate and micro-conglomerate were used as potential strain markers. However there is several strain and vortisity studies in the Sanandaj-Sirjan metamorphic belt (Sarkarinejad, 2007; Samani, 2013, 2015 & Keshavarz, 2015) but results show different amounts of strain partitioning and vorticity related to lithology inhomogeneities. In this paper strain and vorticity analyses were made with application of meso and micro-scale pebbles of deformed conglomerate rocks using R_f/Ø method (Ramsay, 1967; Dunnet, 1969) and Rigid Grain Net (RGN) method (Jessup *et al.,* 2007). Finally the symmetry of shear zone was determined with combination of strain and vorticity results and some structural aspect such as foliations, lineations and the attitude of shear zone boundaries.

Geological Setting

Iran is an assemblage of marginal Gondwana fragments that detached from the Gondwanan–Arabian plate during Permian or Early Triassic times (Stöcklin, 1968; Saki, 2010; Nance *et al.,* 2010). The Zagros orogenic belt is part of the Alpine–Himalayan Mountain Range and extends in a NW–SE direction from eastern Turkey to the Minab Fault System in southern Iran (Haynes and McQuillan, 1974; Stöcklin, 1968). The Zagros orogenic belt is considered to be a complex product of an Early Mesozoic separation of the Iranian continental block from the rest of the Gondwana landmass followed by a NE-dipping subduction of the newly generated Neo-Tethyan oceanic crust below the Central Iranian microcontinents and subsequent collision between the Afro-Arabian and Central Iranian microcontinents (Berberian & King, 1981; Alavi, 1994, 2004). The Late Cretaceous to Tertiary convergence between the Afro-Arabian continent and Central Iranian microcontinents accounts for thrusting and large-scale strike-slip faulting associated with crustal shortening in the Zagros Orogen (Alavi, 1994; Mohajjel and Fergusson, 2000; Sepehr & Cosgrove, 2005; Lacombe *et al.* 2006). Collision is still an ongoing orogenic process (Talebian & Jackson, 2002; Allen *et al.,* 2004; Regard *et al.,* 2004; Vernant *et al.*, 2004) with a convergence rate of approximately 20 ± 2 mmyr−1 (Vernant *et al.,* 2004). The Sanandaj-Sirjan metamorphic belt (SSMB) is one of the major tectonic units of the Zagros orogeny that extends 1500 km long from the Bitlis area in Turkey to the western end of Makran and 100 km wide, (Mohajjel *et al.,* 2003), The SSMB is shown as a coherent assemblage of tectonometamorphic units belonging to the upper Iranian plate (Mohajjel *et al.*, 2003). This belt is a zone of thrust faults that have transported numerous slices of variously metamorphosed Phanerozoic stratigraphic units (Samani, 2013). The study area within the SSMB is located near the Goushti village, 280 km northern of Shiraz city (Figs. 1a, 1b), southwestern Iran. The most abundant rocks are metamorphic rocks such as slate, recrystalized limestome, marble and deformed sandstone and conglomerate from upper Paleozoic to Cenozoic ages. The tectonics of this are characterized by several thrusts, all transporting older rock units on the younger rocks from NE to SW in piggyback style. Strain and vorticity analyses are the main goal of this paper, therefore was specially focused on the pebbles of J^{l}_{2-3} rock unit as a strain marker. The J^{l}_{2-3} rock unit is one of the most important rocks in the area. The most abundant rocks in this unit are moderate to strong deformed conglomerate and micro-conglomerate. These deformed rocks are bounded between two major northeastern and southwestern thrust sheets (Fig. 1c) which are part of the Zagros Thrust System. The deformed conglomerate and micro-conglomerate mainly consist of quartz pebbles and grains. The matrix around the quartz pebbles and grain is mainly composed of altered mica and feldspar minerals. The approximate similarity of composition between pebbles, grains and matrix indicates low rheological contrast between them. The metamorphic grade in this conglomerate is greenschist facies conditions (Sarkarinejad, 1999, 2007).

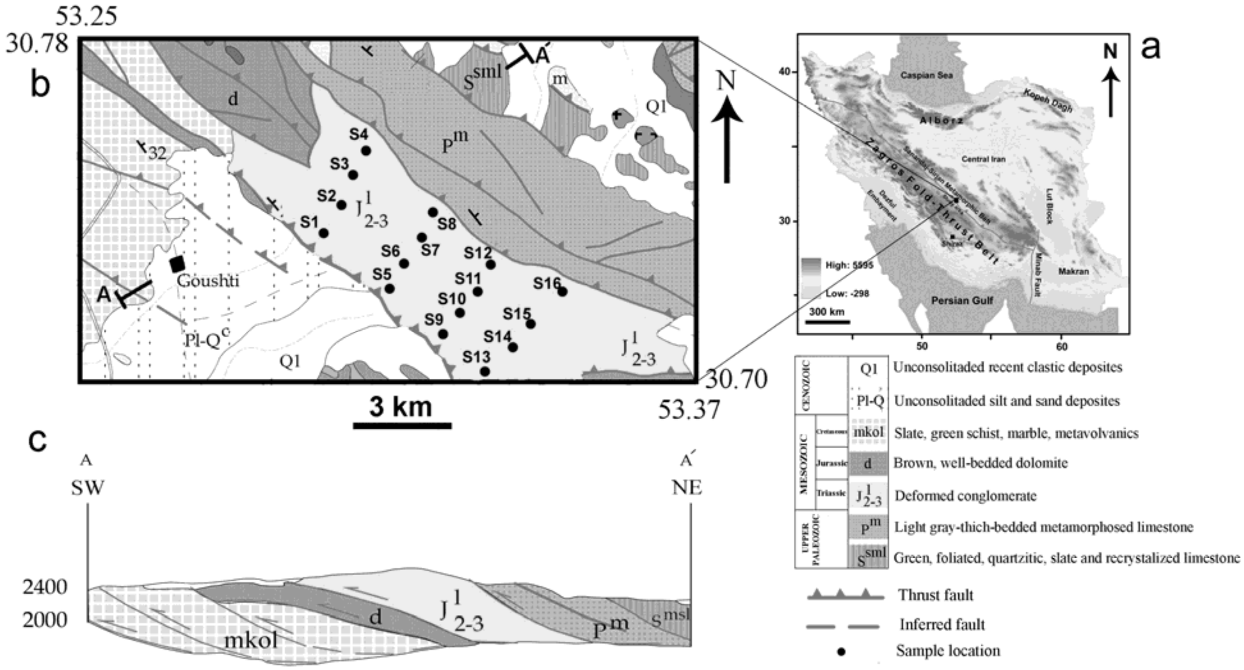

Figure 1. a: Regional setting of Zagros Fold - Thrust Belt and Sanandaj-Sirjan Metamorphic Belt. B: Geological map of the study area. C: Geological cross section along AA´.

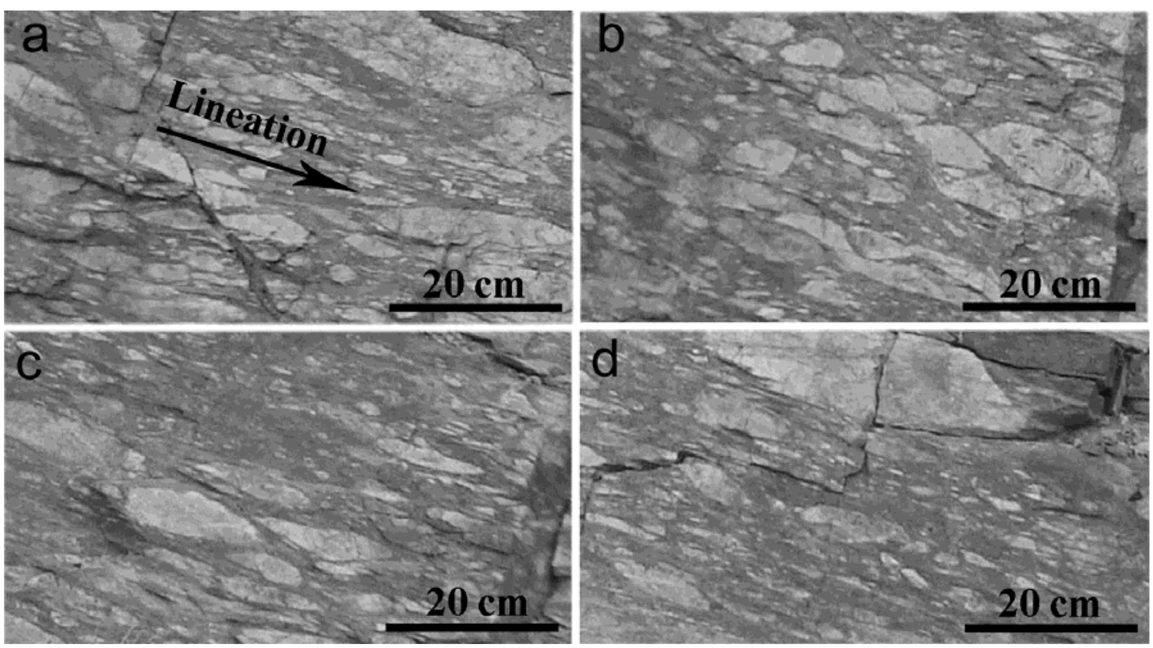

Figure 2. (a, b, c, d) Preferred orientation of pebble long axes in the Goushti deformed conglomerate and micro-conglomerate.

These rocks are strongly foliated and records a lineation marked by preferred orientation of pebble long axes (Fig. 2).

Foliation in conglomerate is defined by alignment of flattened pebbles in the XY plane of finite strain ellipsoid. The attitude of the preferred orientation of the conglomerate or foliation varies between N70°W/30°NE to N40°W/60°NE. The long axis preferred orientation of pebbles in the foliation plane shows a stretching lineation trending from 310°/5°NW to 340°/30°NW.

Material and method

As mentioned, the existence of strain markers is the most important factor for strain studies. Following the classic paper of Cloos (1947), a diversity of methodologies has been proposed in order to estimate finite strain in deformed rocks. The $R_f/Ø$ method (Ramsay, 1967; Dunnet, 1969) and Fry method (Fry, 1979) are the most common methods that have been used by structural geologists; they use the shape ($R_f/Ø$ method) and distribution (Fry method) of objects (e.g. deformed ooids, pebbles of deformed conglomerate and deformed fossils) or of points (e.g. quartz grain centers in quartzite). In this paper with application of pebble markers the amounts of strain ratio in three principal planes of strain ellipsoid was determined using the $R_f/Ø$ method (Ramsay, 1967; Dunnet, 1969). In some cases preparation of 3 perpendicular cut sections were difficult or impossible, therefore it was at list tried to provide 2 perpendicular cut sections parallel to the principal plane of strain ellipsoid. Finally the strain ratio on the one another principal plane was calculated by equation: $R_{XZ} = R_{XY}.R_{YZ}$ (Ramsay and Hubber, 1983). Flinn Digram was used to understand the 3-D nature of the strain ellipsoid. Strain ellipsoid shape parameter K=(R_{XY}-1/ R_{YZ}-1) and strain intensity $D=((R_{XY})^2+(R_{YZ})^2)^{1/2}$ were calculated for each sample. In order to determination of strain partitioning the amounts of kinematic vorticity number (W_k) was evaluated with application of strain markers. W_k was originally defined as an instantaneous rotation relative to the instantaneous stretching at a point (Truesdell, 1953; Means *et al.,* 1980). In most settings structural geologists observe the end product of flow and cannot measure instantaneous quantities. Tikoff and Fossen (1995) transformed W_k to represent three-dimensional finite deformation parameters, defined in terms of shear strains and stretches. Other studies have used W_n (neutral vorticity number, Passchier, 1988) and W_m (mean vorticity number, Passchier, 1988) to define vorticity. In natural systems the vorticity of flow may vary with both position and time (Fossen and Tikoff, 1993). This research uses the mean vorticity number, W_m (Passchier, 1988) which integrates the vorticity of the flow (W_k) over time and space. The amounts of W_m were calculated based on Rigid Grain Net (RGN) method (Jessup *et al.,* 2007). Jessup *et al.* (2007) have used the Rigid Grain Net (RGN) to unify the most commonly used W_m plots by comparing the distribution of theoretical and natural porphyroclasts within a flowing matrix. As shown by Gosh and Ramberg (1975), the sense and rate of rotation of a particle depend on its orientation, axial ratio (R) and the ratio between the elongation in the shear plane and shear strain. For W_m=1, i.e. simple shear, all particles which behave as active markers with R≥1 will rotate freely as the shear strain increases and the rate of rotation equals the rate of stretching. If W_m is lower than 1, i.e. a component of pure shear accompanies shearing (general or sub-simple shear of Simpson and De Paor 1993) and the rotation of particles with progressively smaller aspect ratios is subdued (Cowan, 1990). For any flow regime with W_m<1, not all rigid particles are free to rotate continuously. According to the strain and vorticity data the amounts and percentages of shortening were determined in the area based on graphical function of Passchier (1988). Finally the symmetry of the shear zone was determined based on the amount of (χ) angle that defined as the angle between the stretching lineation and intersection of foliation and shear zone boundary (Jones *et al.,* 2004).

Strain analysis

Strain analysis has been performed on 16 representative samples from the deformed conglomerate. Samples localities are shown on Fig. 1b. It tried to all samples collected from quartz rich horizons containing low competency contrast between pebbles and matrix and well-defined planar and linear fabric elements. The pebble shape of ductile deformed conglomerates was used as strain marker. Length-to-width ratios of pebbles were determined from measurements made on sections cut normal and parallel to the foliation and lineation (Fig. 3).

Finally, the strain ratios (R_{xz}, R_{xy} and R_{yz}) were estimated for each sample applying $R_f/Ø$ shape analysis (Ramsay, 1967; Lisle, 1985).

Ø is the angle between the X-axis of the elliptical strain marker and an arbitrary reference direction. In order to measure the angular relationship between long axes of pebbles in the XZ, YZ and XY planes, intersection foliation and stretching lineation trace lines were considered as the references line. Figure 4 shows the $R_f/Ø$ shape analysis for samples 7, 11 and 15. The strain ratios on the three principal strain planes for all samples are presented in table 1.

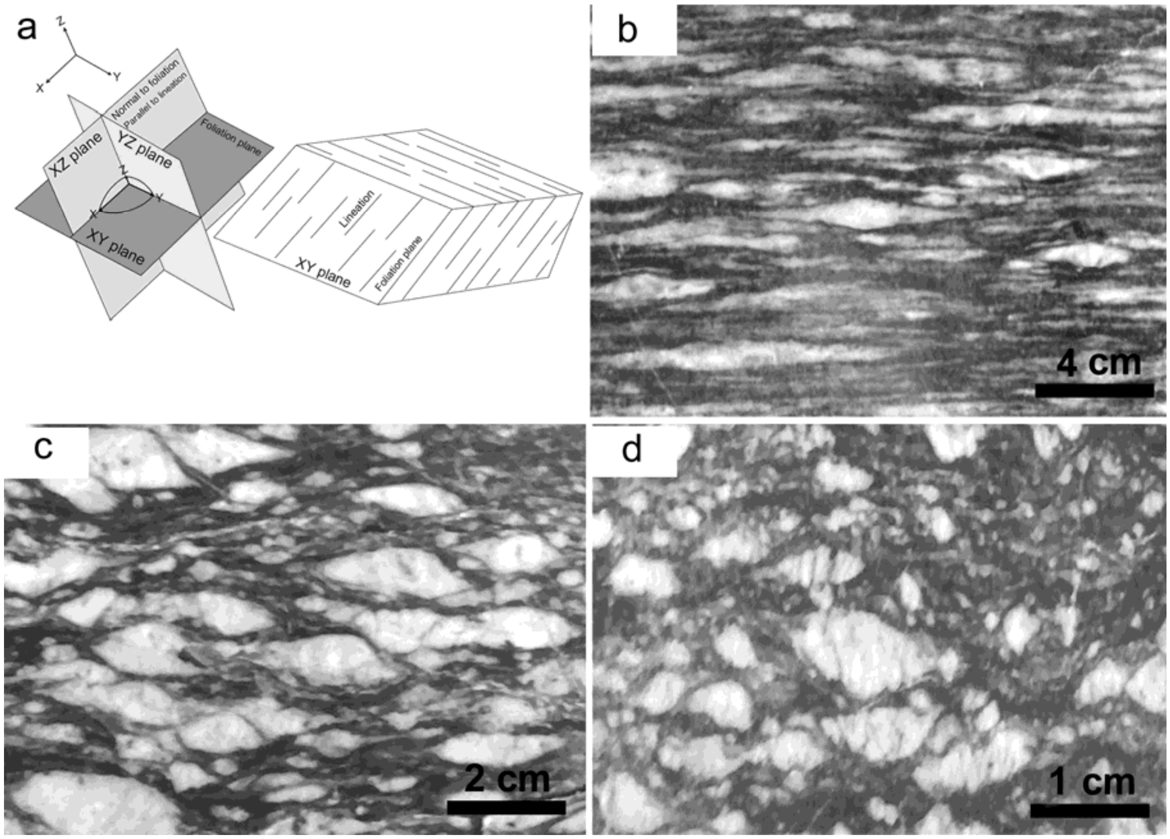

Figure 3. a: Position of the principal planes of strain ellipsoid related to the planar and linear structural elements. (b, c and d) Mesoscopic photographs of the deformed micro-conglomerate in the three principal planes (XZ, XY and YZ) of strain ellipsoid.

Harmonic mean (H) of axial ratio of XZ, XY and YZ surfaces was determined for each sample according to (Ramsay & Huber, 1983): $H = n/(R_{f1}^{-1} + R_{f2}^{-1} + ... + R_{fn}^{-1})$. These data were plotted in the Flinn diagram (Fig. 5) and K (strain ellipsoid shape) and D (strain intensity) values were determined for each sample. Table 1 represents the K and D resulted from all localities.

Kinematic vorticity analysis

Kinematic vorticity (W_K) is a dimensionless measure of rotation relative to strain and characterizes the amount of shortening proportional to displacement. For cases of simple shear and sub-simple shear, W_m is measured on a scale between 0 and 1, with 0 being pure shear and 1 being simple shear. The W_m scale is not linear, but can be converted to a linear scale by considering the percent of a deformation resultant from simple shear. Although some workers have categorized deformations as either pure- or simple shear dominated, in reality there is a continuum from pure shear to simple shear. Fort and Bailey, 2007 proposed three separate fields for pure, general, and simple shear dominated deformations. Pure-shear dominated deformations have W_m-values of 0–0.3, corresponding to less than 20% simple shear. In contrast, simple shear dominated deformations have W_m-values of greater than 0.95, corresponding to greater than 80% simple shear. General shear occupies the range between 0.3 and 0.95. There are several methods for estimating the W_m parameter (Xypolias, 2009 and references cited therein). The use of porphyroclasts rotating in a flowing matrix to estimate the mean kinematic vorticity number (W_m) is important for quantifying the relative contributions of pure and simple shear in deformed rocks.

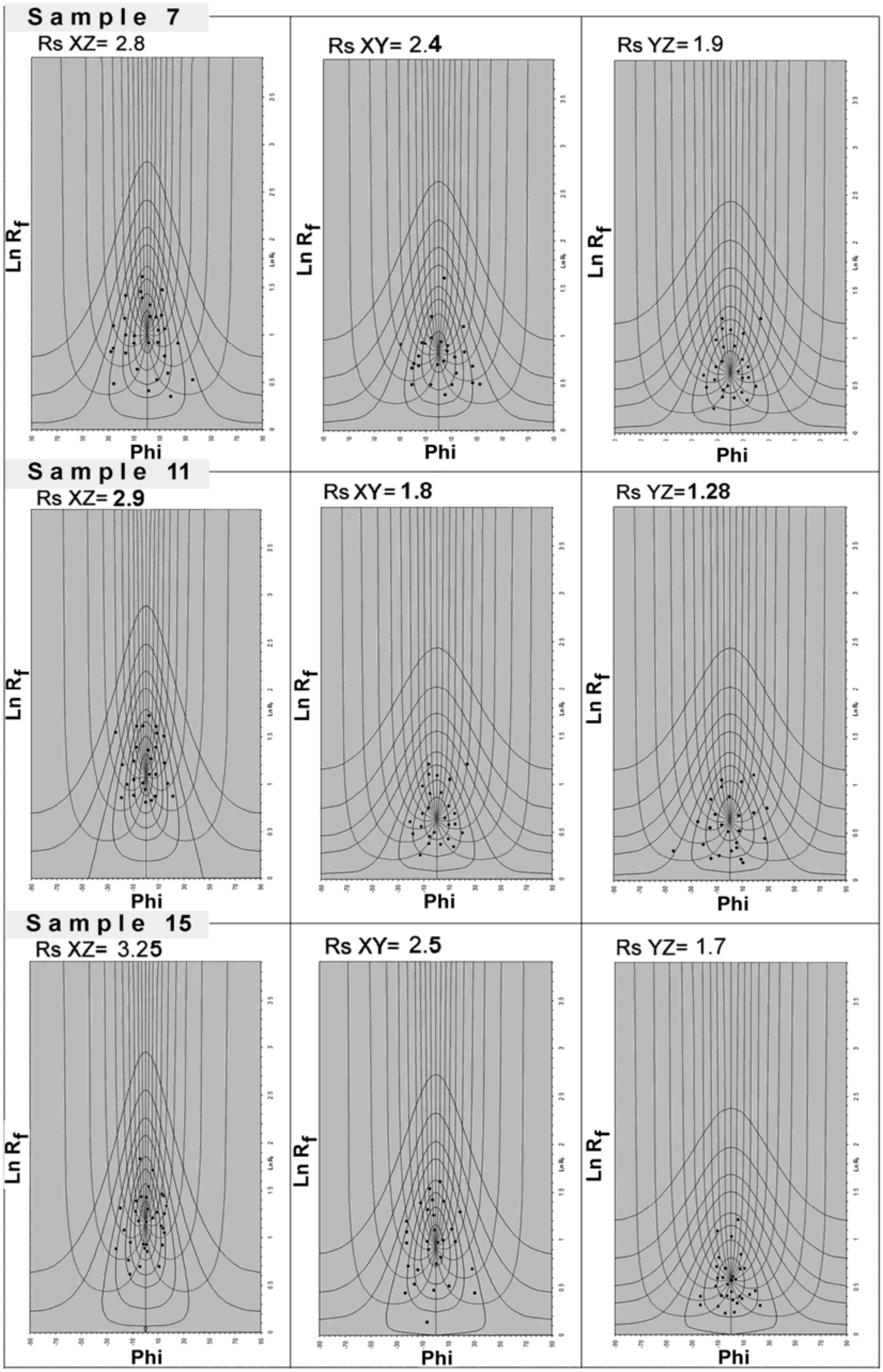

Figure 4. Strain analysis in the three principal planes of strain ellipsoid using Rf/Ø method (samples 7, 11 and 15).

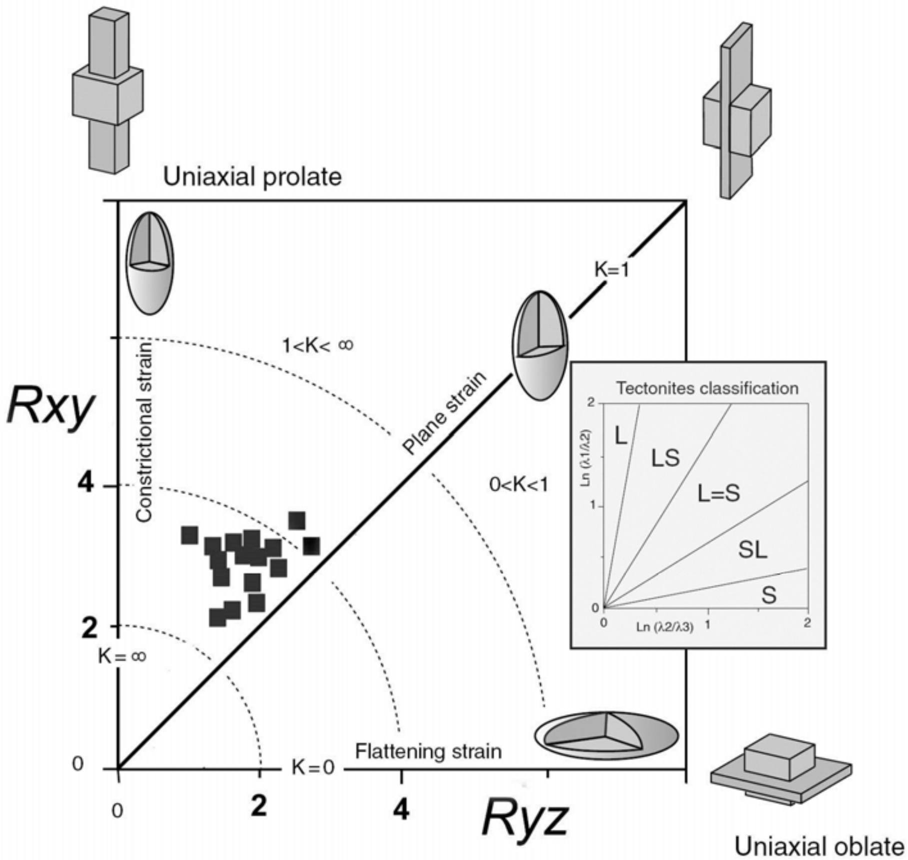

Figure 5. Ellipsoid shape and strain intensity analyzed by plotting the finite strains for XY and YZ principal sections on a Flinn diagram.

The most common methods, broadly grouped into those that use porphyroclasts, have been applied to many different tectonic settings. Particles with an aspect ratio above a certain critical value, R_c, will rotate until they reach a stable orientation. For aspect ratios less than the critical value, rotation is unrestricted. The value of R_c that divides freely rotating objects from those that have reached a stable orientation is a function of the degree of non-coaxiality (Passchier 1987):

$$W_m = (R_c^2 - 1)/(R_c^2 + 1)\ .$$

The RGN makes an important new contribution that advances the current methods for quantifying flow in shear zones (Jessup *et al.,* 2007). In this research we applied the Rigid Grain Net (RGN) method for sixteen samples in the study area which yields $0.5 \le W_m \le 0.85$ (Fig. 6 and table 1).

Discussion

Strain Partitioning

Strain and vorticity analyses show an important strain partitioning in the study area. It reveals that the amounts of finite strain ratios and mean kinematic vorticity numbers are dominantly controlled by distance from major basal and dorsal thrust faults in the area. Vorticity analysis shows that, strain partitioning causes observed different distributions of pure and simple shear components of strain in the study area. Deferent amounts of W_m represent simple shear dominant deformation near the major thrust faults and pure shear dominant deformation far from them (Fig. 7a). All the strain parameters such as strain ratio in the XZ principal plane of strain ellipsoid (R_{XZ}), strain ellipsoid shape (K) and strain intensity (D) show increasing mode towards the major basal and dorsal thrust faults (Fig. 7b, c, d).

The amounts of natural logarithmic strain (ε) calculated using the following equation (Ramsay and Huber, 1983):

$$\varepsilon = (\frac{1}{3})^{\frac{1}{2}}\left[(\ln(R_{XZ}))^2 + (\ln(R_{YZ}))^2 + (\ln(R_{XY}))^2\right]^{1/2}.$$

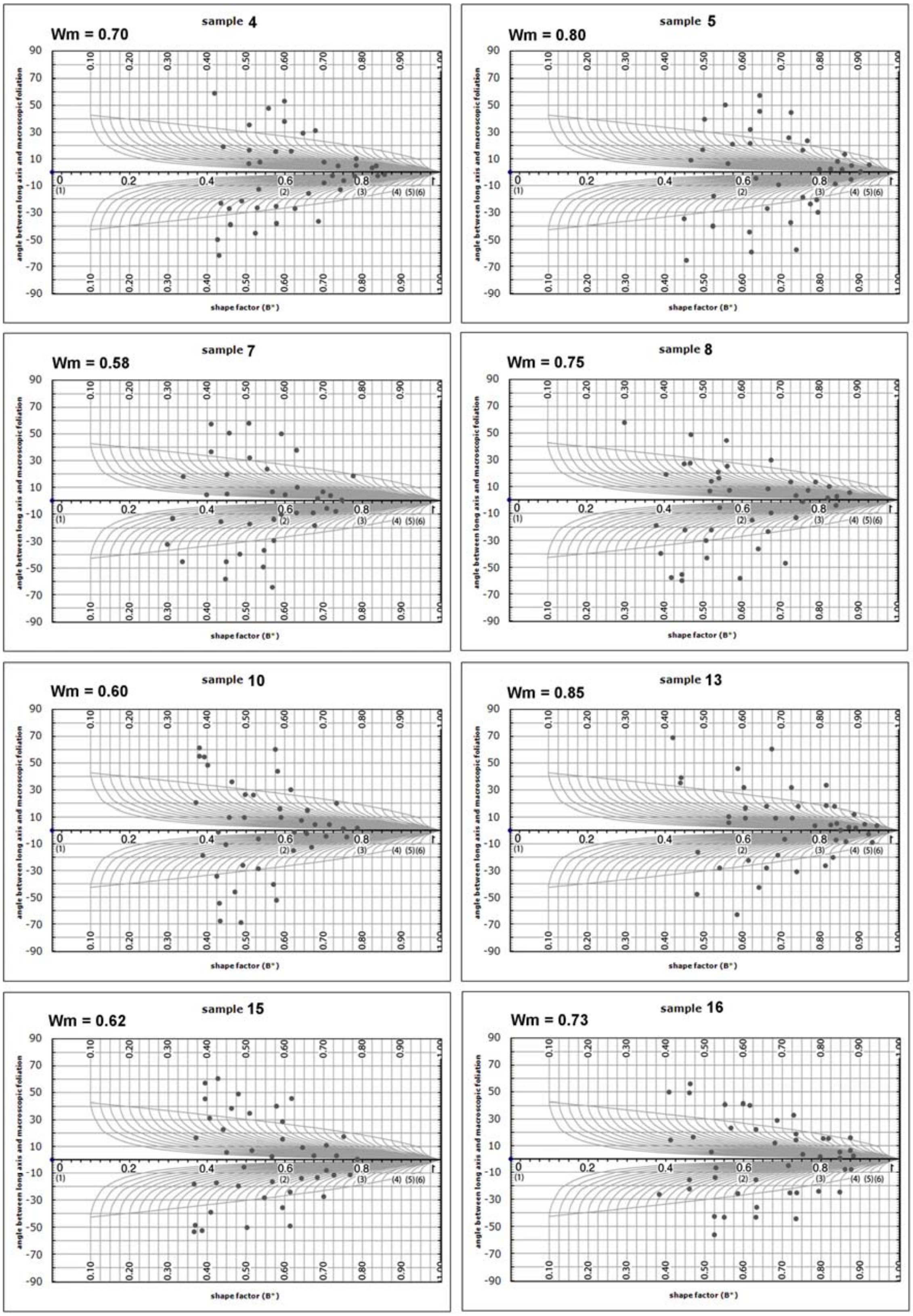

Figure 6. Kinematic vorticity analysis based on Rigid Grain Net (RGN) method.

Table 1. Strain measurements and kinematic vorticity number data for all the stations in the study area.

Sample	R_{XZ}	R_{XY}	R_{YZ}	K	D	W_k	ε
1	4.3	3	1.82	2.4	2.2	0.82	1.2
2	2.1	2.3	1.8	1.6	1.5	0.65	0.73
3	2.4	2.2	1.7	1.7	1.4	0.67	0.75
4	3.9	2.8	2	1.8	2.1	0.7	1.1
5	4.15	2.8	1.65	2.7	1.9	0.8	1.1
6	2.3	2.1	1.7	1.5	1.3	0.5	0.71
7	2.8	2.4	1.9	1.5	1.7	0.58	0.87
8	3.65	2.9	1.8	2.3	2.1	0.75	1.2
9	3.85	2.9	1.9	2.1	2.1	0.7	1.1
10	2.5	2.3	1.7	1.8	1.5	0.6	0.78
11	2.9	1.8	1.28	2.8	0.8	0.6	0.72
12	4.5	3.4	2.1	2.1	2.6	0.85	1.2
13	4	3.1	2.4	1.5	2.5	0.75	1.2
14	2.7	2.5	2.1	1.3	1.9	0.5	0.9
15	3.25	2.5	1.7	2.1	1.7	0.62	0.91
16	3.8	3	2	2	2.2	0.73	1.1

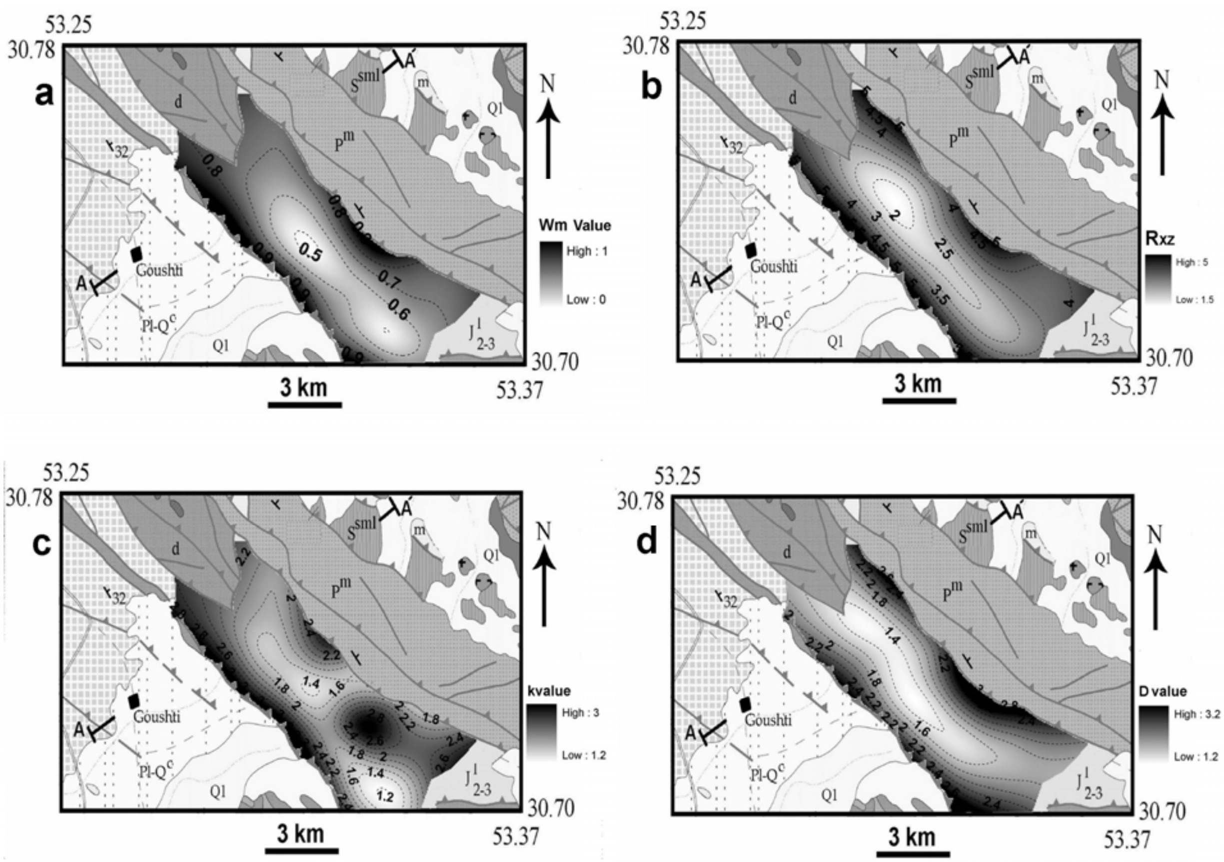

Figure 7. (a, b, c and d) Couture plot of variations of Wm, strain ratio (RXZ), K and D values in the study area.

Vitale and Mazzoli (2009) discriminated mylonite types using strain intervals of ε=0–1 (protomylonites), ε=1–2.5 (mylonite), and ε>2.5 (ultramylonite). According to this classification, the higher values of the natural logarithmic strain were observed around the southern basal and northern dorsal thrust faults (mylonite); however, the area between two thrust faults shows the lower ε values (protomylonites) (Fig. 8).

An analytical solution and graphical function to determine the shortening have been given by Wallis (1992) and Passchier (1988), respectively. Using the R_{xz} and W_m values and application of the graphical function of (Stretch-R_{xz}-W_m) (Passchier 1988), the amounts of shortening was determined between 0.37 and 0.62 (Fig. 9a). Figure 9b shows the percent amounts of the shortening in the area (38 to 63 %) that is related to regional convergence between Afro-Arabian continent and Iranian microcontinent. According to the Fig 9a in addition to shortening component of strain, some stretching component of strain (S≤1.3) have been occurred in the study area. This stretching is the main factor for lateral extrusion and development of the stretching lineation in the area.

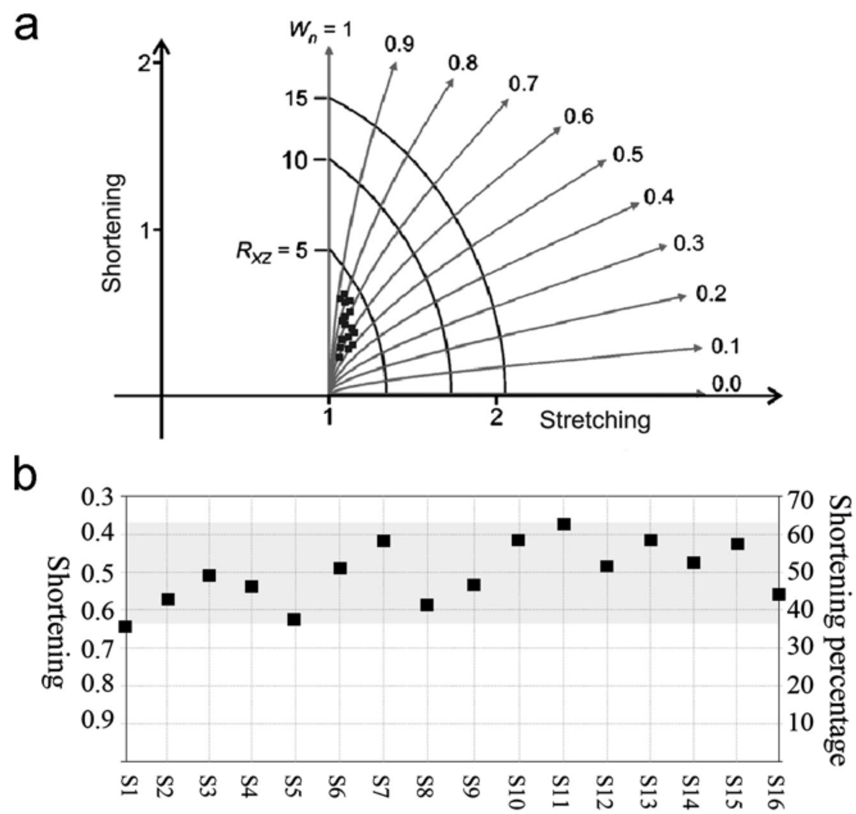

Figure 8. a: Amounts of shortening using graphical function of (Stretch-Rxz-Wm). b: The percent amounts of shortening in the area.

Symmetry of the Shear Zone

Theoretical modeling has shown the relation between structural pattern evolution and the symmetry of shear zones. A major effort has been made to predict the foliation and stretching lineation patterns in various kinematic types of shear zones, assuming that the foliation plane is parallel to the local $\lambda_1\lambda_2$ plane and the stretching lineation is parallel to the λ_1 direction (Lin *et al.,* 2007a,b and references cited therein). In many shear zones the variation of stretching lineation and development of foliation with non-parallel strike and dip respect to the shear zone boundaries are affected from the simple shear/ pure shear component ratio (Lin *et al.,* 1998). The strike obliquity of the foliation will reflect the strike-slip component of deformation. The dip of the foliation will generally be steeper than dip of the zone boundary due to the non-coaxial component of overthrusting (Lin *et al.,* 1998). Stretching lineations lying almost strike-parallel, or significantly oblique, or approximately down-dip are all feasible in inclined transpression zones, so that the angle of pitch of the lineation within the foliation plane can vary from almost 0° to nearly 90° (Jones *et al.,* 2004). When the elongation direction remains roughly down-dip or strike-parallel respect to the shear zone boundaries the shear zone symmetry can appear almost monoclinic (Lin *et al.,* 1998; Jiang *et al.,* 2001), whereas intermediate values of pitch are associated with strains that are more recognizably triclinic. Therefore the χ angle (angle between the stretching lineation and intersection of foliation and shear zone boundary that calculate in the foliation plane) can be defined when the deformation zone boundaries are well exposed and well defined. Shear zone symmetry can be recognized as triclinic when the angle (χ) neither 0 nor 90. The deformation zone boundaries in the study area are irregular in shape because of the complex thrust structures of the Sanandaj-Sirjan metamorphic belt. According to the field observations and geological maps the thrust faults confining the conglomerate unit considered as the shear zone boundaries. Therefore, the attitude of the zone boundary has been considered as N40°W, 30°NE approximately. The average dip of foliation is steeper than the dip of the zone boundaries. The angle between the lineation and intersection of the foliation and zone boundary measured in the plane of foliation (χ) is of 52° (Fig. 10).

This value of χ indicate triclinic strain symmetry (Jones *et al.,* 2004) for the transpression zone which is consistent with the obliquity of three axes of finite strain ellipsoid with respect to the boundaries of the transpression zone (e.g. Sarkarinejad, 2007; Sarkarinejad & Azizi, 2008). Using stereographic analysis and structural characters of the study area (i.e. deformation zone boundary, stretching lineation and foliation) the angle χ was determined 52° (Fig. 10a). The stretching lineation is neither strike-parallel nor parallel to the true dip of the zone boundary in the study area. The stretching lineation is oblique with respect to the strike and true dip of the deformation zone. The plunge of the stretching lineation varies with average amount of 330°/20°NW. Average strike and the dip of foliation in the deformed rocks of the study area is N50°W, 45°NE. Figure 10b shows the schematic inclined triclinic transpression model for the study area. This model describe the occurrence of non-coaxial deformation in all principal plane of strain ellipsoid, shortening and stretching or lateral extrusion along the Z and X axes respectively.

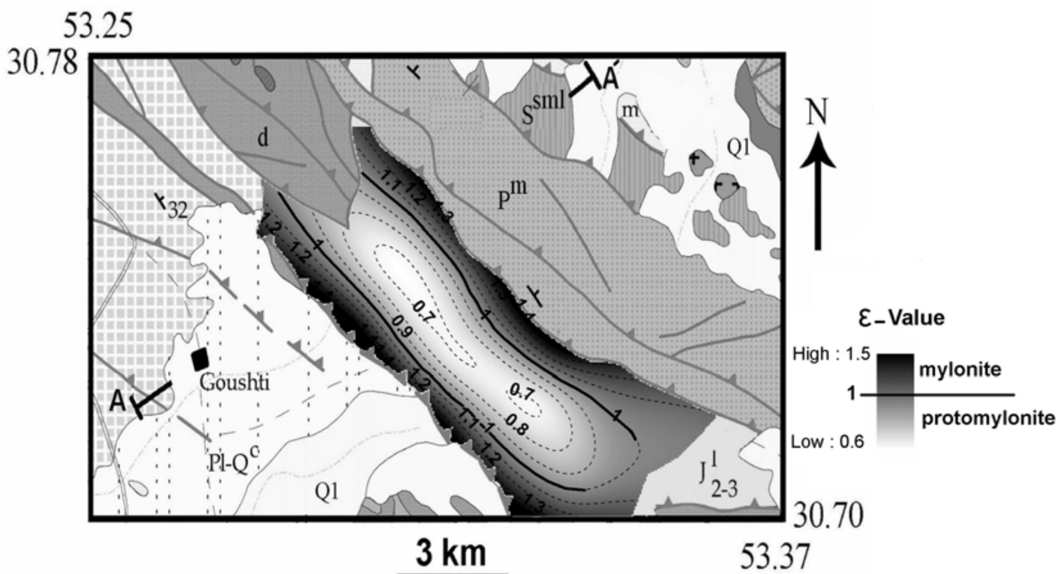

Figure 9. Natural logarithmic strain variations and rock type classification (mylonite and protomylonites) in the study area

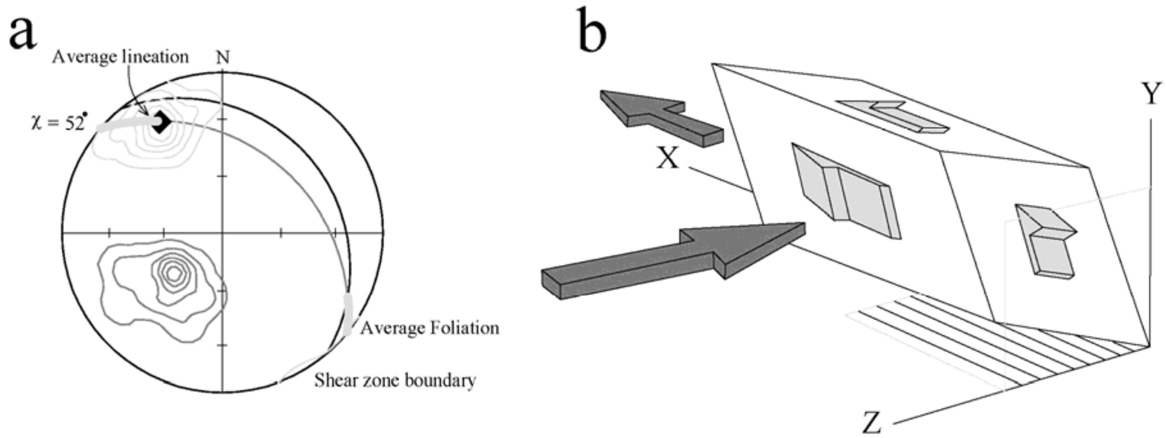

Figure 10. (a) Equal-area lower hemisphere projection showing quantitative interpretation of the data in the study area. Foliation generally dips more steeply than the zone boundaries. The pitch of stretching lineation within the plane of foliation can be steep, moderate, or shallow depending on the state of finite strain. χ is the angle between the stretching lineation and the intersection of the foliation and zone boundary, measured in the plane of the foliation, and for triclinic strains $0° < \chi < 90°$. (b) 3D model of the study area shows the non-coaxiality components of strain in the three principal planes of strain. The lateral extrusion of material depending on the shear zone narrowing has provided the penetrative stretching lineation in the area.

Conclusion

Strain and vorticity analyses show an important strain partitioning in the Goushti shear zone. Results show that the changes in the amounts of finite strain ratios and mean kinematic vorticity are dominantly controlled by distance from the major basal and dorsal thrust faults. Vorticity analysis shows strain partitioning causes different distributions of pure and simple shear components of strain in the study area. Deferent amounts of W_m shows simple shear dominant deformation near the major thrust faults and pure shear dominant deformation far from them. Strain parameters such as strain ratio in the XZ principal plane of strain ellipsoid (R_{XZ}), strain ellipsoid shape (K) and strain intensity (D) show increasing mode towards the major basal and dorsal thrust faults. The higher values of the natural logarithmic strain near the southern basal and northern dorsal thrust faults reveal development of mylonite rocks in these places; however, the area between two thrust faults shows protomylonite rocks. Results show (38 to 63 %) shortening in the area related to the convergence between Afro-Arabian continent and Iranian microcontinent. The relation between foliation and stretching lineation and the condition boundary of the deformed area offer an important parameter which help to understanding the symmetry of shear zones. The angle between lineation and intersection of the foliation and zone boundary measured in the plane of foliation (χ) was determined 52° which indicate triclinic strain symmetry for the study area.

Acknowledgements

The author would like to thank the editor of Geopersia (Dr. Ebrahim Ghasemi-Nejad) for the editorial authority. I am grateful to anonymous reviewers for helpful and valuable reviews of the manuscript. Also I would like to thank Professor Adrian Foster for final grammar review of the paper. The research was supported by the Research Council grant of the Shahid Chamran University of Ahvaz which is gratefully acknowledged.

References

Alavi, M., 1994. Tectonics of the Zagros orogenic belt of Iran: new data and interpretations. Tectonophysics, 229: 211–238.

Alavi, M., 2004. Regional stratigraphy of the Zagros Fold-Thrust Belt of Iran and its proforeland evolution. American Journal of Science, 304: 1–20.

Allen, M.B., Jackson, J., and Walker, R., 2004, Late Cenozoic reorganization of the Arabia-Eurasia collision and comparison of the short-term and long-term deformation rates. Tectonics, 23, 639–654.

Berberian, M., King, G.C.P., 1981. Towards a paleogeography and tectonic evolution of Iran. Canadian Journal of Earth Sciences, 18: 210–265.

Cloos, E., 1947. Oolite deformation in the South Mountain fold, Maryland. Bulletin of Geological Society of America, 68: 843–918.

Cowan, D.S., 1990. Kinematic analysis of shears zones in sandstone and mudstone of the Shimanto belt, Shikokn, SW Japan. Journal of Structural Geology, 12: 431–441.

Dunnet, D., 1969. A technique of finite strain analysis using elliptical particles. Tectonophysics, 7, 117–136.

Dutton, B.J., 1997, Finite strains in transpression zones with no boundary slip. Journal of Structural Geology, 19: 1189–1200.

Fort, A.M., Bailey, C.M., 2007. Testing the utility of the porphyroclast hyperbolic distribution method of kinematic vorticity analysis. Journal of Structural Geology, 29: 983–1001.

Fossen, H., Tikoff, B., 1993. The deformation matrix for simultaneous simple shearing, pure shearing and volume change, and its application to transpression–transtension tectonics. Journal of Structural Geology, 15: 413–422.

Fry, N., 1979. Random point distribution and strain measurements in rocks. Tectonophysics, 60: 89–105.

Gosh, S.K. and Ramberg, H., 1976. Reorientation of inclusions by combination of pure shear and simple shear. Tectonophysics, 34: 1–70.

Haji Hosseinlou, H., 2003.Structural Analysis of Eqhlid Fault Shear Zone, southwest of Iran, M.Sc. Thesis, University of Shiraz, Shiraz Iran,

Harland, W.G., 1971. Tectonic transpression in Caledonian Spitzbergen. Geological Magazine, 108, 27–42.

Haynes, S.J., McQuillan, H., 1974, Evolution of the Zagros Suture zone, southern Iran. Geological Society of America Bulletin, 85: 739–744.

Holdworth, R.E., Tavarnelli, E., Clegg, P., Pinheiro, R.V.L., Jones, R.R., McCaffrey, K.J.W., 2002. Domainal deformations pattern and strain partitioning during transpression: an example from the southern Uplanda terrane, Scotland. Journal of the Geological Society of London, 159: 401–415.

Horsman, E. and Tikoff, B., 2007. Constraints on deformation path from finite strain gradients. Journal of Structural Geology of London, 29: 256–272.

Hosseni, SH., Structural, microstructural and kinematics of Zagros thrust system in the Dehbid area, Iran, M.Sc. Thesis, University of Shiraz, Shiraz Iran, 2009.

Iacopini, D., Passchier, C.W., Koehn, D., Carosi, R., 2007, Fabric attractors in general triclinic flow systems and their application to high strain shear zones: a dynamical system approach. Journal of Structural Geology, 29: 298–317.

Jessup, M.J., Law, R.D., Frassi, C., 2007: The Rigid Grain Net (RGN): an alternative method for estimating mean kinematic vorticity number (Wm). Journal of Structural Geology, 29: 411–421.

Jiang, D., Lin, S., and Williams, P.F., 2001, Deformation path in high-strain zones, with reference to slip partitioning in transpressional plate boundary regions. Journal of Structural Geology, 23: 991–1005.

Jones, R.R., Holdsworth, R.E., Clegg, P., McCaffrey, K., Tavarnelli, E., 2004. Inclined transpression. Journal of Structural Geology, 26, 1531–1548.

Jones, R.R., Tanner, G.P.W., 1995. Strain partitioning in transpression zones. Journal of Structural Geology, 17, 793–802.

Keshavarz, S., 2015.Structural and microstructural analyses of the HP-LT Sanandaj-Sirjan metamorphic belt, southwestern Sirjan, Iran, Ph.D. Thesis, University of Shiraz, Shiraz Iran.

Lacombe, O., Mouthereau, F., Kargar, S., Meyer, B., 2006, Late Cenozoic and modern stress fields in the western Fars (Iran): implications for the tectonic and kinematic evolution of central Zagros. Tectonics, 25.

Lin, S., Jiang, D., Williams, P.F., 1998. Transpression (or transtension) zones of triclinic symmetry: natural example and theoretical modelling. In: Holdsworth, R.E., Strachan, R.A., Dewey, J.F. (Eds.), Continental Transpressional and Transtensional Tectonics, Special Publication of the Geological Society of London, 135: 41–57.

Lin, S., Jiang, D., and Williams, P.F., 2007a, Importance of differentiating ductile slickenside striations from stretching lineations and variation of shear direction across a high-strain zone. Journal of Structural Geology, 29: 850–862.

Lin, S., Jiang, D., Williams, P.F., 2007b. Kinematics of shear zones: a review with emphasis on a triclinic model and the importance of differentiating ductile slickenside striations from stretching lineations. Geological Bulletin of China, 26: 19–31.

Lisle, R.J., 1985, Geological Strain Analysis: A Manual for the Rf/Ø Technique. Pergamon Press, Oxford, 99 p.

Means, W.D., Hobbs, B.E., Lister, G.S., Williams, P.F., 1980, Vorticity and noncoaxiality in progressive deformations. Journal of Structural Geology, 2: 371–378.

Mohajjel, M. and Fergusson, C.L., 2000, Dextral transpression in Late Cretaceous continental collision, Sanandaj–Sirjan Zone, western Iran. Journal of Structural Geology, 22: 1125–1139.

Mohajjel, M., Fergusson, C.L., and Sahandi, M. R., 2003. Cretaceous–Tertiary convergence and continental collision, Sanandaj–Sirjan Zone, western Iran. Journal of Asian Earth Sciences, 21: 397–412.

Nance, R.D., Gutierrez-Alonso, G., Keppie, J.D., Linnemann, U., Murphy, J.B., Quesada, C., Strachan, R.A., and Woodcock, N.H., 2010. Evolution of the Rheic Ocean. Gondwana Research, 17: 194–222.

Passchier, C.W., 1987. Stable positions of rigid objects in non-coaxial flow: a study in vorticity analysis. Journal of Structural Geology, 9: 679–690.

Passchier, C.W., 1988. Analysis of deformation paths in shear zones. Geologische Rundschau, 77: 309–318.

Ramberg, H., 1975. Particle paths, displacement and progressive strain applicable to rocks. Tectonophysics, 28: 1–37.

Ramsay, J.G., 1967, Folding and Fracturing of Rocks. McGraw-Hill, New York, 568p.

Ramsay, J.G., Graham, R.H., 1970, Strain variation in shear belts. Canadian Journal of Earth Sciences, 7, 786–813.

Ramsay, J.G., Huber, M.I., 1983. The Techniques of Modern Structural Geology. Strain Analysis, Academic Press, London, 1: 307p.

Richard, P., Cobbold, P., 1990, Experimental insight into partitioning of fault motion in continental convergent wrench zones. Annu Tectonnics, IV (2): 35–44.

Robin, P.F., Cruden, A., 1994. Strain and vorticity patterns in ideally ductile transpression zones. Journal of Structural Geology, 16: 447–466.

Saki, A., 2010. Proto-Tethyan remnants in northwest Iran: geochemistry of the gneisses and metapelitic rocks. Gondwana Research, 17: 704–14.

Samani, B., 2013. Quartz c-axis evidence for deformation characteristics in the Sanandaj-Sirjan metamorphic belt, Iran. J. Afr. Earth Sci. 81: 28–34.

Samani, B., 2015. Strain analysis using structural elements and geometry of surfaces of no finite longitudinal strain. Arabian Journal of Geosciences, 8: 8461–8468.

Sanderson, D.J. and Marchini, W.R.D., 1984, Transpression. Journal of Structural Geology, 6: 449–458.

Sarkarinejad, K., 1999. Tectonic finite strain analysis using Ghuri deformed conglomerate, Neyriz area Southwestern Iran. Iranian Journal of Sciences and Technology, 23: 351–363.

Sarkarinejad, K., 2007. Quantitative finite strain and kinematic flow analyses along the Zagros transpression zone, Iran. Tectonophysics, 442: 49–65.

Sarkarinejad, K., Azizi, A., 2008. Slip partitioning and inclined dextral transpression along the Zagros Thrust System, Iran. Journal of Structural Geology, 30: 116–136.

Sepehr, M., Cosgrove, J.W., 2005. Role of the Kazerun fault zone in the formation and deformation of the Zagros Fold-Thrust Belt, Iran. Tectonics, 24.

Simpson, C., De Poar, D.G., 1993. Strain and kinematic analysis in general shear zones. Journal of Structural Geology, 15: 1–20.

Stöcklin, J., 1968. Structural history and tectonics of Iran. A review. American Association of Petroleum Geologists Bulletin, 52: 1229–1258.

Sullivan, W.A., Law, R.D., 2006: Deformation path partitioning within the transpressional White Mountain shear zone, California and Nevada. Journal of Structural Geology, 29:583–599.

Talebian, M., Jackson, J., 2002. Offset on the Main Recent Fault of NW Iran and implications for late Cenozoic tectonics of the Arabia–Eurasia collision zone. Geophysical Journal International, 150: 422–39.

Tikoff, B., Fossen, H., 1995, The limitations of three-dimensional kinematic vorticity analysis. Journal of Structural Geology, 12: 1771–1784.

Tikoff, B., Teyssier, C., 1994. Strain and fabric analysis based on porphyroclast interaction. Journal of Structural Geology, 16: 477–492.

Truesdell, C., 1953. Two measures of vorticity. Journal of Rotational Mechanical Analysis, 2: 173–217.

Vernant, P., Nilforoushan, F., Hatzfeld, D., Abbasi, M.R., Vigny, C., Masson, F., Nankali, H., Martinod, J., Ashtiani, A., Bayer, R., Tavakoli, F., and Chery, J., 2004. Present-day crustal deformation and plate kinematics in the Middle East constrained by GPS measurement in Iran and northern Oman. International Journal of Geophysics, 157: 381–398.

Vitale, S., Mazzoli, S., 2009. Finite strain analysis of a natural ductile shear zone in limestones: Insights into 3-D coaxial vs. non-coaxial deformation partitioning. Journal of Structural Geology, 31: 101–113.

Wallis, S.R., 1992. Vorticity analysis in a metachert from the Sanbagawa belt, SW Japan. Journal of Structural Geology, 14: 271–280.

Xypolias, P., 2009. Some new aspects of kinematic vorticity analysis in naturally deformed quartzites. Journal of Structural Geology, 31: 3–10.