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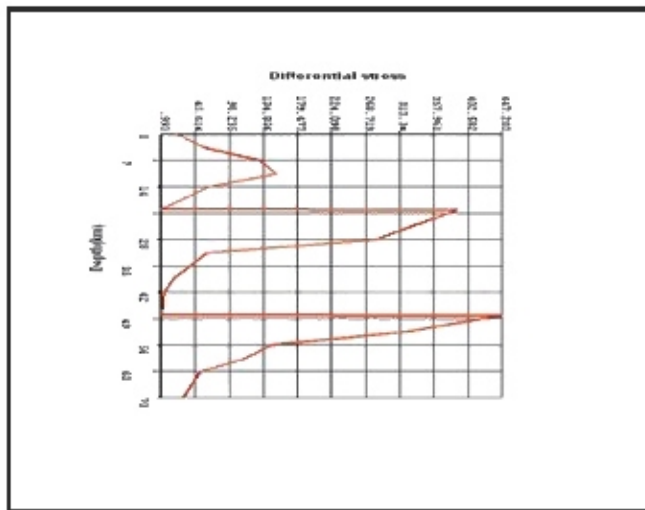Morgan, P., Sass, J. A., 1984- Thermal regime of the continental lithosphere, J. Geodyn. 1 (1984) 143–166.

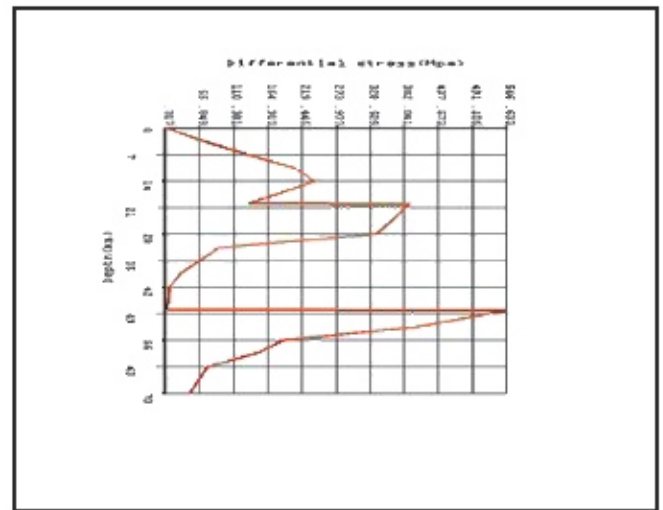Figure 6- Differential stress versus depth for cold geotherm and wet rheology



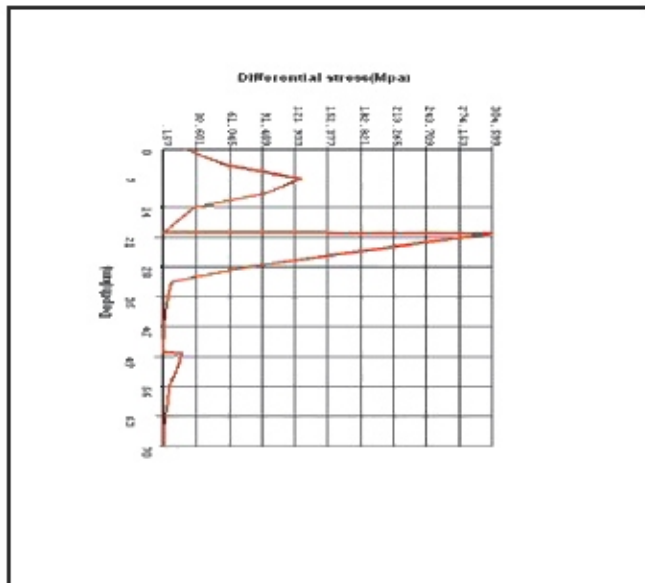Figure7- Differential stress versus depth for cold geotherm and dry rheology



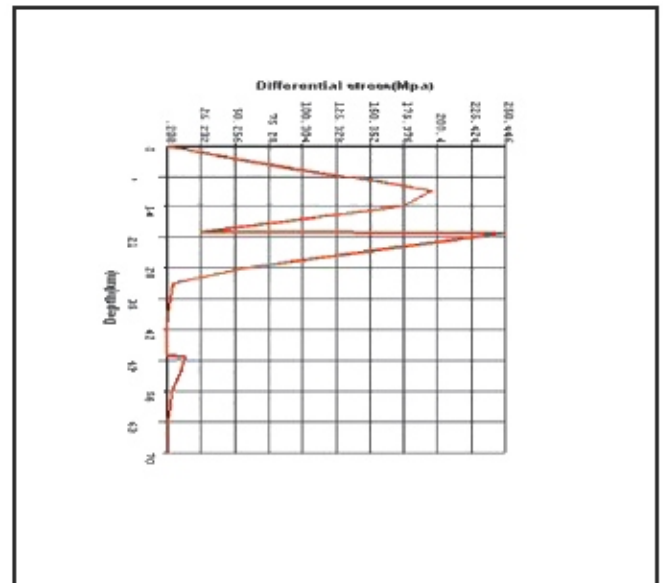Figure8- Differential stress versus depth for hot geotherm and wet rheology



Figure9- Differential stress versus depth for hot geotherm and dry rheology

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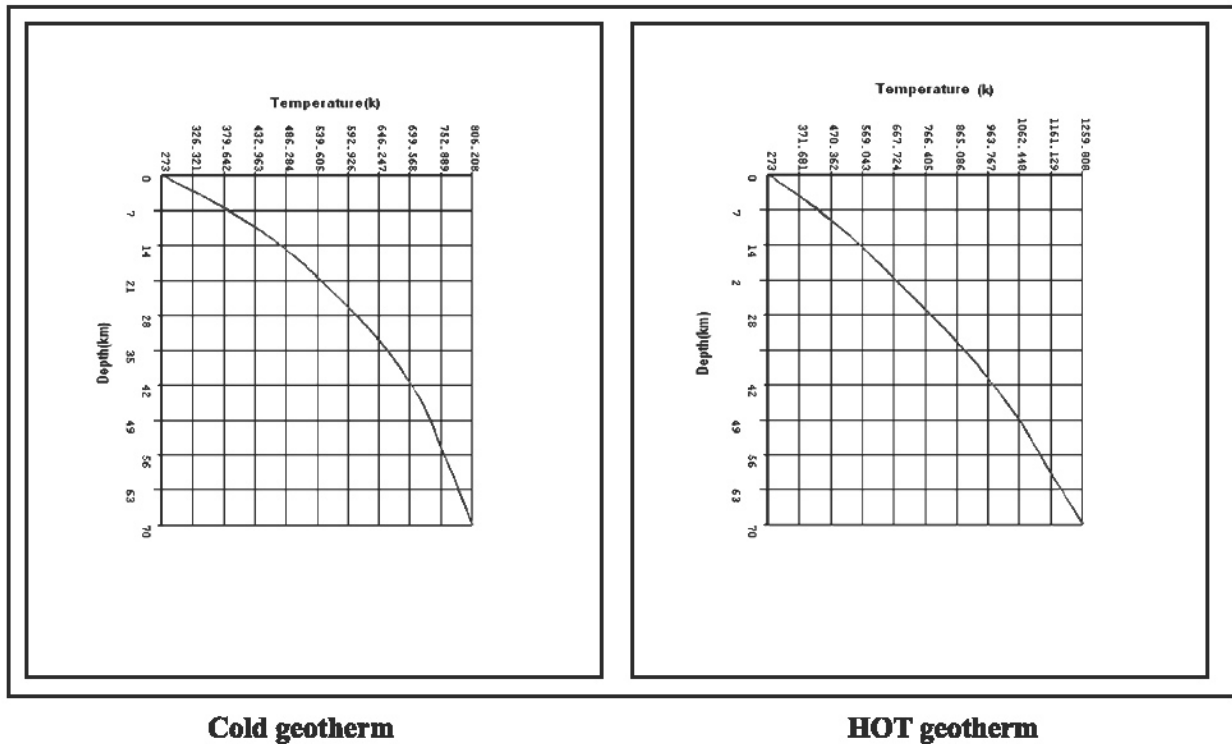Cold geotherm                                                    HOT geotherm

Figure4- The initial geothermal gradient (left, cold, right, hot) estimate from thermal analysis used in the model input to the creep equation that controls material behavior. Model elements behave according to Hook's law where temperatures are low and become transitionally more viscoelastic at higher temperatures. Horizontal axis represents the temperature in kelvin and the vertical axis represents the depth in km.
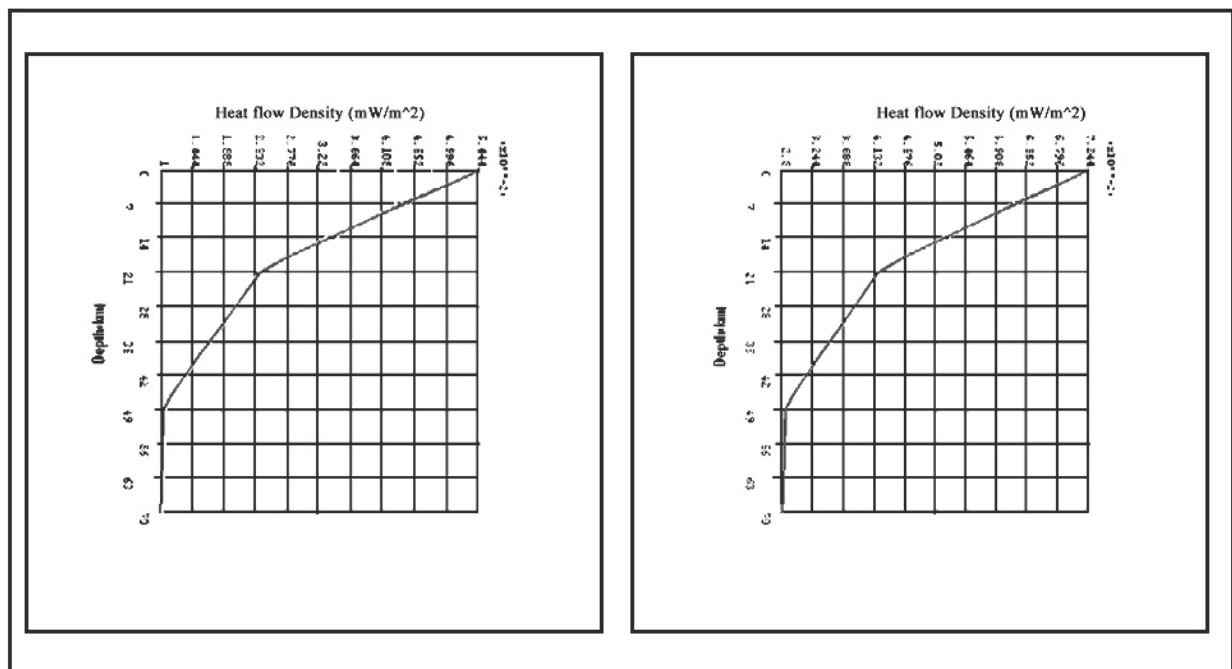


Figure5- Estimated initail heat flow in the Zagros (cold&hot)case. The surface heat flow is about 0.054 mWm-² for cold case (left) and 0.072mWm-² for hot case (right)
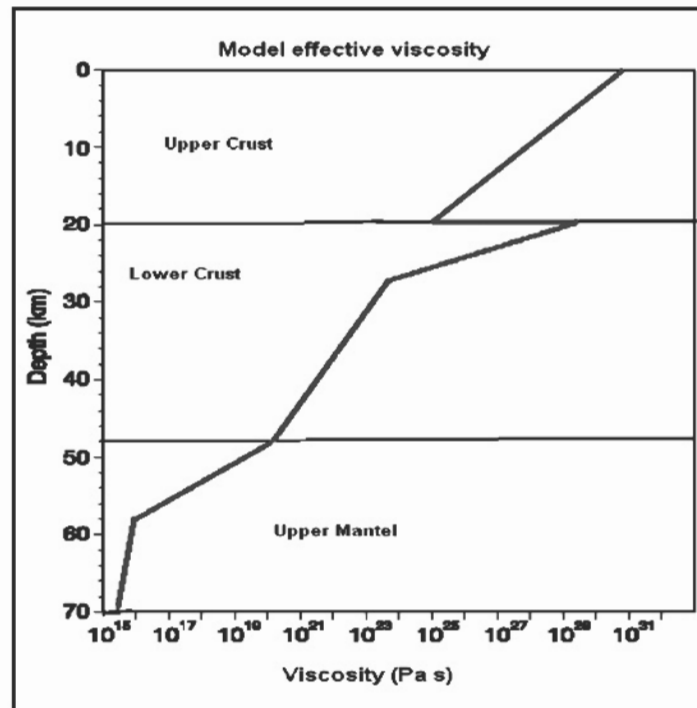
Figure3A- Effective viscosity versus depth Effective viscosity was calculated according to $\eta_c = \sigma^{(-n)} \exp (Qc/RT)/2A$, where s is differential stress (calculated with the finite element model), R is the gas constant, and Qc (activation energy), A, and n are experimentally determined constants (Table 1).



Finite Element Model(Zagros)
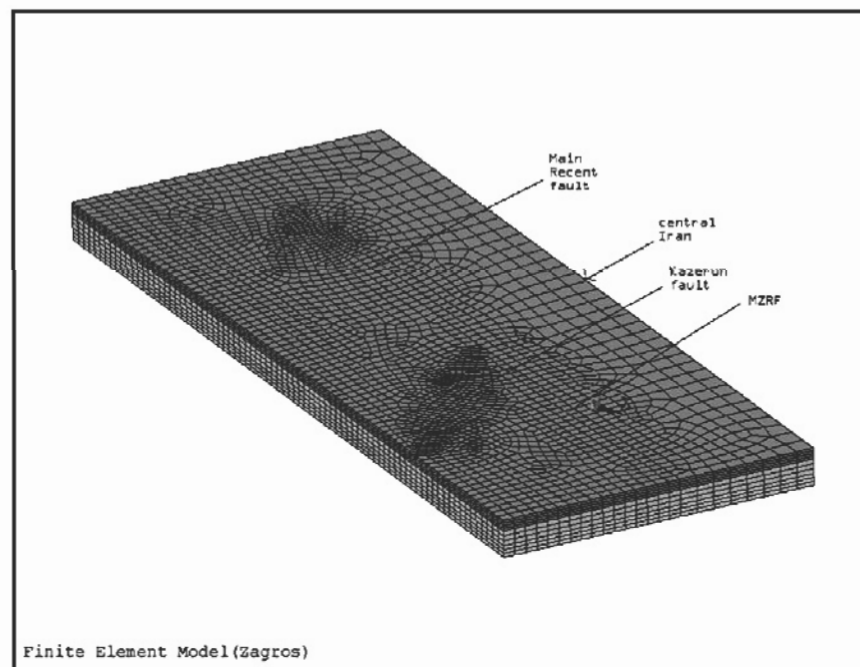
Figure3.B- Oblique view of the finite element mesh, model boundaries, and fault structures, elements have been chosen as the viscoelastic material, with viscous behavior controlled by the temperature gradient and have eight nodes located at the corners. The faults are discontinuous in the model and have deformable contact elements on their surfaces that obey a Coulomb failure criterion.

Table2- Material Constants used in the Three Layers of the Finite Element (Artemieva, 2001)

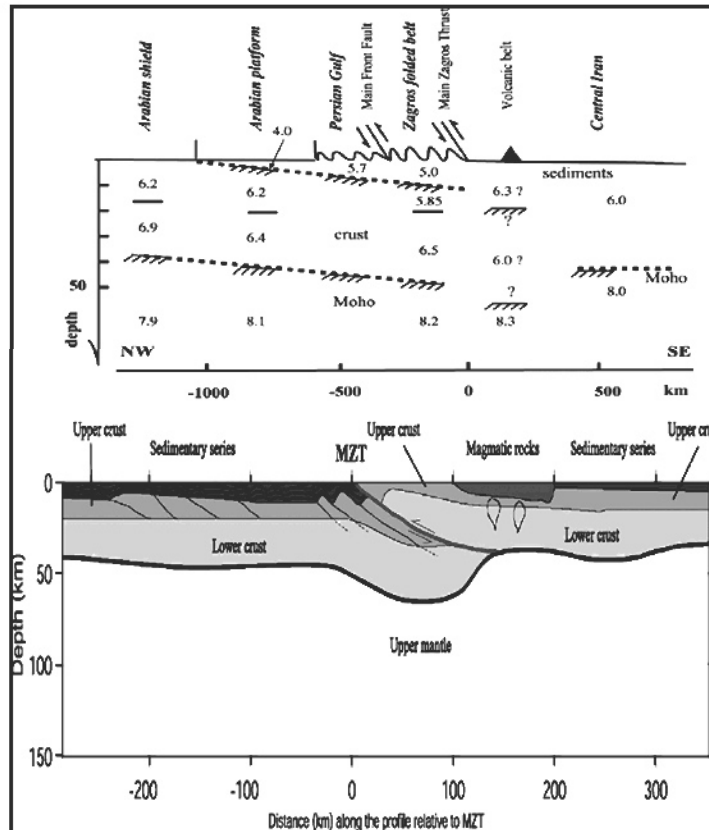| Surface temperature (K) | Ts | 273 |
|---|---|---|
| Lithospheric basal flow(Wm-²) | q* | 0.019 |
| | | |
| **Upper Crust** | | |
| Thermal conductivity (Wm⁻¹ ok ⁻¹) | K | 3 |
| Radiogenic heat Production (wm⁻³) | H | $1.5 \times 10^{-6}$ |
| **Lower crust** | | |
| Thermal conductivity (Wm⁻¹ °k ⁻¹) | K | 2.5 |
| Radiogenic heat Production (wm⁻³) | H | $0.5 \times 10^{-6}$ |
| | | |
| **Upper mantle** | | |
| Thermal conductivity (Wm⁻¹ °k ⁻¹) | K | 3 |
| | | |
| Radiogenic heat production (Wm⁻³) | H | $0.02 \times 10^{-6}$ |



Figure2- Seismic profile and the Moho depth and crustal section in Zagros
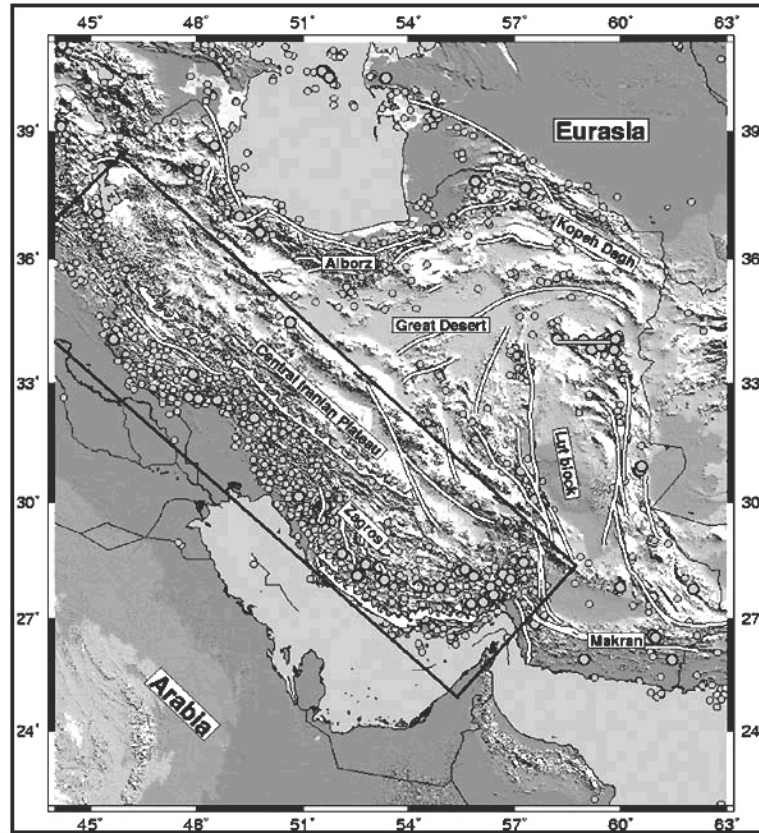(Up- Hatzfeld et al., 2003, down –Paul et al., 2006)

Figure1B- Tectonic setting of Iran (Masson, 2005), solid line shows the boundaries of this study

Table1- Material Constants used in the Three Layers of the Finite Element Model.

| Parameter | Layer 1 0-20km | Source | Layer 2 20-48km | Source | Layer3 >48km | Source |
|---|---|---|---|---|---|---|
| E Young's modulus, MPa | $6 \times 10^4$ | 1 | $9.4 \times 10^4$ | 1 | $1.6 \times 10^5$ | 1 |
| A physical constant, MPa$^{-n}$S$^{-1}$ | $2 \times 10^4$ | 2 | $6.3 \times 10^4$ | 5 | $5 \times 10^1$ | 6 |
| n Power law exponent | 1.9 | 2 | 3.1 | 5 | 3.8 | 6 |
| Q activation energy, kJ/mol | 140.6 | 2 | 276 | 5 | 492 | 6 |
| $V$ Poisson's ratio | 0.25 | 3 | 0.26 | 3 | 0.28 | 3 |
| $P$ density*10$^3$kg/m$^3$ | 2.775 | 4 | 2.905 | 4 | 3.200 | 4 |

Elements in the model are all viscoelastic, with viscous behavior controlled by the temperature gradient and the listed constants. References: 1, Hatezfeld [2003]; 2, Hansen and Carter [1983]; 3, Christensen [1996]; 4, Snyder and Barzangi [1986]; 5, Caristan [1982]; 6, Carter and Tsenn [1987].

The upper part of the lithosphere fails as a brittle/elastic medium, the lower part by creep, which fails aseismically. Thus the boundary between the two is often termed the brittle-ductile transition and is often considered the base of the seismogenic part of the crust. Note that this transition moves upward in response to higher temperatures. This variation in the brittle-ductile transition has been observed, at least crudely, around the globe in continental areas.

In figure 6 at the upper crust level brittle failure in occur around 10.5 km depth where the differential stress reaches 150 Mpa. From the surface to 10.5 km the differential stress gradually increase up to 150 MPa. This effect is the direct consequence of the mean stress on the yield criterion. At greater depth in the upper crust, it decreases to 3 MPa due to strong temperature dependence. In the lower crust stress range between 380 Mpa to 4Mpa and failure occur from 21km to 28 km. At greater depth in the uppermost mantle, stress ranges between 447 and 30 Mpa and there is no failure, in the mantel and therefore system is mechanically coupled. For case 2, when hot geotherm input in to the model, brittle failure occur in 8 km. In figure 8 the brittle failure in Zagros is around 8 km where the stress reaches 122 Mpa. From the surface to 8 km the differential stress gradually increase from up to 122 MPa. This effect is the direct consequence of the mean stress on the yield criterion. At greater depth in the upper crust, it decreases to 1 MPa due to strong temperature dependence. In the lower crust stress ranges between 304 Mpa to 1 Mpa. At greater depth in the uppermost mantle, stress ranges between 20 and 1 Mpa and failure occur from 48 km to 56 km. The system is mechanically decoupled.

## 10- Discussion & Conclusions

A histogram of the relocated earthquakes shows that the majority of earthquakes are located between 5–20 km depth with a maximum number of events at 12 km (figure1.A) No earthquakes are reliably located at depth greater than 20 km. The seismicity, therefore, is likely to be located in the upper part of the crystalline basement below the 8-10km thick sedimentary layer (Tatar et al., 2004; Hatzfeld et al,. 2003).Our results are consistent with the depth computed by body-wave inversion for all strong events in Zagros (Maggi et al., 2000 ,Tatar et al., 2004; Engdhal et al., 2006).

The calculated strengths with wet upper crust for Zagros are shown in figures 6 and 8. The maxima define the brittle–ductile transitions from which the upper crustal maximum can be considered to be more important. This transitional layer is located between the depths of 8 and 10.5km. The lower crust maximum is located between 21and 28 km. Typically the brittle–ductile transition is considered to be the lower limit of the earthquake focal depths. However, the real lower boundary of the seismicity is controlled by modification of the frictional behaviour. This transition between velocity-weakening and velocity-strengthening for granitic composition is approximately at the temperature of 250°-350° (Scholz, 1990). In our

thermal model depth of brittle ductile transition appears between 10 to 17 km and 18 to 25 km( Figure4). This transitional depth is nearly consistence with the brittle–ductile transition depth derived from our model. Analysis of the focal depth distribution in the Zagros has shown that 75% of the events occurred between the depth of 10 to 15 km and 65% between the depths 15 to 25 km (Engdahl et al., 2006 Tatar et al., 2004). Dry rheological conditions have a major impact on the rheology, both on the brittle and ductile behaviour. In the brittle regime an assumption of the dry conditions increases the strength (Kohlstedt, 1995), and this assumption in our model change the brittle ductile transition to14km for cold and 10.5-14km for hot geotherm as the pore fluid factor is neglected (Figures 7&9) Kohlstedt, 1995. In the ductile regime the consequences are similar, i.e. the ductile strength increases as the material becomes stronger. It seems that temperature has significant effect on the rheology of the lithosphere with the same material properties. It changes the BDT zone in the lithosphere entirely. The results of the model much better compatible with the cold geotherm in the Zagros. Based on the correlation between surface heat flow and seismicity depth in continental areas (Meissner & Strehlau1982; Sibson 1982), only initail thermal field appear compatible with a temperature of 250-350 °c occurring between 10 and 25 km in Zagros. This depth range appears consistent with the minimum and maximum seismicity depth deduced from local seismic surveys in the Zagros (Baker et al., 1993; Tatar et al., 2004; Yaminifar et al., 2006).
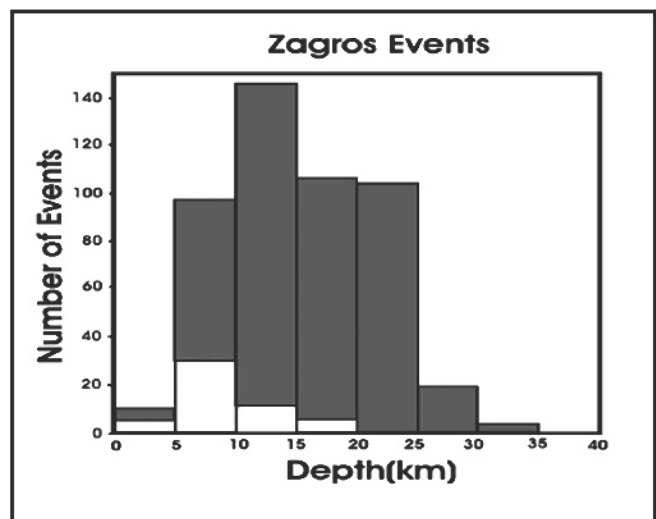
Figure1A- Earthquake depth distribution in Zagros (After Engdahl et al., 2006). Depths determined by waveform modelling are not shaded. Shaded are EHB focal depths

subdivided into a series of increments applied over several steps. Before each solution, the Newton-Raphson method evaluated the out-of-balance load vector, which is the difference between the restoring forces (the loads corresponding to the element stresses) and the applied loads. A linear solution was performed using the out-of-balance loads. If convergence criteria were not satisfied, then the out-of balance load vector have been reevaluated, the stiffness matrix updated, and a new solution was obtained.

## 8- Initial Results

The results of initial thermal model are shown in figures 4&5. Solid black line in horizontal direction shows temperature in Kelvin and in vertical direction shows the depth. Temperature in the level of Moho is 732 K for cold case and, 1062 K for hot case, these are agreements with previous thermal model done in this area (Bird, 1975; Vernant & Chery, 2006). The resulting surface heat flow density is 54mW/m² for cold case and 72mW/m² for hot case. At the base of the upper crust heat flow density is 19mW/m² for cold case and 42 m/Wm² in hot case. The 300 ± 50⁴⁻ isotherm (Fernandez and Ranalli, 2003) consider to set a limit for frictional behavior reached a depth between 18 km to31 km cold geotherm and 10 to17 km for hot geotherm.

## 9- Lithosphere strength

The mechanical behavior of the lithosphere can be characterized by various rheological laws which depend mainly on rock composition, temperature, pressure, strain and strain rate (e.g., Ranalli, 1995). The most important deformation mechanisms are elasticity, brittle failure and viscous flow. Elasticity is used to describe the recoverable strains of model materials at low differential stresses, but is restricted to a few percent of deformation only. Brittle deformation by fracture and frictional sliding on existing faults occurs if the applied stress exceeds the brittle strength. Brittle strength can be estimated according to the Coulomb-Navier criterion (e.g. Ranalli, 1995) which gives the critical shear stress required to overcome frictional resistance by Eq. 1. An examination of rock properties reveals a large variation in rock strength with lithology. This would suggest that lithology would be the most important factor in determining the strength of rocks. However, experiments have revealed that this variation is almost entirely restricted to the effort to break a rock; once broken, the frictional behavior of a fault remains pretty uniform from lithology to lithology. What is more, this behavior depends almost entirely on the effective pressure on the fault, with little dependence on temperature or strain rate. This observation is termed Byerlee's Law and is expressed as a relationship of the shear stress and effective normal stress on a fault. If one of the principal stresses coincides with the vertical stress as is the case close to the earth's surface, Eq. (1) can be recast in terms of the differential stress (Ranalli, 1995), (Sibson, 1977; Ranalli, 1995),

$$\sigma_1 - \sigma_3 = A \rho g z(1 - \lambda) \quad , \quad (5)$$

where g is gravitational acceleration, λ is ratio of pore pressure to lithostatic pressure (Observations down to a couple of kilometers depth in continents suggest that a pore pressure between (dry) and hydrostatic are probably about correct for much continental crust), ρ is the density of the rocks overlying a depth z and A is a parameter that depends on the coefficient of friction and the predominant type of faulting. This approach assumes that the orientation of the preexisting faults is that given by frictional theory and that cohesion on the faults is negligible when compared with effective normal stress. In the finite element model, brittle behavior is approximated by continuum deformation using an elastic – perfectly plastic flow law with a depth – (vertical stress-) dependent yield stress. Thus, material behaves elastically at stresses below the yield stress as defined by Eqs (1) and (5), but flows instantaneously according to an ideal plastic flow law if the yield stress is exceeded. This rheology is designed to account for permanent strain at stress levels above the yield stress, but does not explicitly describe frictional sliding and fault movement. As temperature increases with depth, ductile flow becomes important and lithospheric deformation is assumed to be governed by thermally activated power law dislocation creep. The non-linear relationship between strain rate, and differential stress and temperature T is given by (Kirby, 1983)

$$, \quad d\varepsilon / dt = A \, exp(-Qc / RT)\sigma^n \quad (6)$$

where R is the universal gas constant and A (strain rate coefficient), n (power law stress exponent) and Q (activation energy) are material properties derived from laboratory experiments (e.g., Ranalli, 1995). Power law creep holds for high temperatures and low to moderate differential stresses. At low temperatures or high stresses, this relation vanishes and strength becomes rather independent of strain rate and temperature (Carter and Tsen, 1987). The rheological laws outlined above are combined to calculate lithospheric strength in the mechanical model. The specific strength at any given point in the lithosphereis determined either by the modified Coulomb-Navier criterion (Eq. (5)) or by power law dislocation creep (Eq. (6)) depending on which deformation mechanism provides the lowest yield strength. There appear to be two fundamental modes of deformation of rocks: brittle failure and ductile flow. The former is related to earthquakes, which represent such failure; the latter should not generate earthquakes. The boundary between the two regimes is termed the brittle-ductile transition. Although it is often represented as a major discontinuity within the Earth, it is in fact most probably a broad zone within which a number of different mechanisms of rock deformation occur. Somewhere within this zone should be the base of that part of the lithosphere that generates earthquakes (the seismogenic layer).

For purposes of strength of the lithosphere, we may use (Eq. (6)) in terms of a stress difference as a function of temperature (Ranalli, 1995).

$$\sigma_1 - \sigma_3 = (d\varepsilon / dt \, exp(Qc / RT)/A)^{1/n} \quad (7)$$

approximately based on P-wave velocity isolines(Hatzfeld et al., 2003; Paul et al., 2006, Tatar, personal communication). The average crustal thickness in the Zagros is 48 km i.e., 20 km for the upper crust (include sediment) and 28 km for the lower crust. The upper mantle depth is 22 km from base of the lower crust. Our mesh covers a rectangular area in the south east to northwest Iran with horizontal dimensions of 1500 km×600 km (Figures. 3&3.1) and a depth extent of 70 km depth. The model is composed entirely of eight-node elements. It consists of 26520 elements with 109379 active nodes. The finite element model has cuts in it that represent the two major strike-slip faults of the Zagros i.e Main Recent fault and Kazerun(Figure.3). The faults are deformable and are constructed from contact elements that obey the Coulomb failure relation (Parsons, 2002, 2006)

$$\Delta CF = \Delta \tau + \mu(\Delta \sigma_i + \Delta P), \quad (1)$$

where $\tau$ is shear stress acting on a fault surface, $\mu$ is the friction coefficient, and $\sigma_i$ is the component normal to a fault surface and P is the pore pressure. Contact elements have zero thickness and are welded to the sides of elements. All elements of the Zagros finite element model strained through a combination of linear elasticity and rate dependent creep behavior. Time-independent elastic strain occurred in the model according to

$$\varepsilon = \sigma/E \quad , \quad (2)$$

Where E is the Young's modulus and $\sigma$ is differential stress. Modeled time-dependent inelastic strain rate was controlled by the creep equation (e.g., Kirby and Kronenberg, 1987)

$$d\varepsilon/dt = A\exp(-Qc/RT)\sigma^n \quad , \quad (3)$$

where A(elastic constant), Qc (activation energy), and n are experimentally derived elastic constants, R is the universal gas constant, T is temperature, and $\sigma$ is differential stress (Table 1). Elements in the model are all viscoelastic, with viscous behavior controlled by the temperature gradient and the listed constants. References: 1, Hatezfeld [2003]; 2, Hansen and Carter [1983]; 3, Christensen [1996]; 4, Synder and Barzangi [1986]; 5, Caristan [1982]; 6, Carter and Tsenn [1987].

## 5- Geotherms

Temperature has a major importance in rheological and the tectonic modeling as, for instance, ductile flow is dependent on temperature in a non-linear way. Conductive heat transfer, understood to be the main mechanism of the heat transport in the lithosphere, is described with Fourier's law of heat conduction where the heat flow is directly proportional to the temperature gradient

$$\nabla.(\nabla T) = -A/k \qquad Q = -k\frac{\partial T}{\partial z} \quad (4)$$

where Q is the near surface heat flow [Wm$^{-2}$], T is the absolute temperature [K], k=k(z) is the coefficient of thermal conductivity [Wm$^{-1}$K$^{-1}$]as a function of depth and temperature (Zoth and Hanell, 1988), and z is the depth [m] and A=A(z) is the heat production as a function of depth (z).

This equation can be used to determine the heat flow density (HFD) with temperature measurements in boreholes as a function of a depth, together with the laboratory estimates of thermal conductivity. If we assume steady state conditions i.e., no temperature change over time, we end up with an equation that can be used to calculate the stable geotherms of the lithosphere. We calculate two different geotherm input to the creep equation that control material behavior.

## 6- Geometry and Boundary Conditions

The geometry of our model is shown in figure3B, and the geographical boundaries are superimposed on the topographic map of figure1B. Our model boundaries are parallel and perpendicular to the Zagros belt trend, i.e extends from North Western Iran (west) near Urumieh to Southeast (East) end of the Zagros( strait of Hormoz ), It is approximately 1600 km×600 km. In the previous sections all the necessary model definitions were described for simulating deformation in Zagros. We assume the horizontal normal stress has two components (lithostatic stress and deviatoric stress). The model is subject to body force (gravity), which established an initial stress state. The bottom of the model was constrained to zero displacement in the vertical direction, and the model edges were not permitted to move laterally. All other nodes were given 3 degrees of freedom (ux, uy, and uz). Elements at the model base mimic low-viscosity asthenosphere because of high temperatures (Figure3A). Therefore while compressed because of the fixed basal boundary condition, elements at or near the model base could not support any stresses (contrast between the base of the lithosphere and the underlying asthenosphere is of several orders of magnitude) into the upper part of the model. After this stage tectonic loading is simulated by remotely moving Arabia past central Iran.

## 7- Numerical Experiments

In principle, the numerical approach would allow to study systematically the impact of all input parameters as well of the initial temperature and boundary conditions once the model geometry has been set up. However, such analysis would go far beyond the scope of this paper. Instead, work concentrates on the most important parameter influencing the modeling results which is temperature. Due to the strong temperature-dependence of the creep laws and the depth of the brittle–ductile transition, temperature has a profound control on the style of lithosphere. Two numerical experiments differing in the initial temperature distribution and, hence, strength of the continental lithosphere are studied. The scenarios intend to represent a mechanically weak, and strong lithosphere. These differences in strength are achieved by varying the basal heat flow and consequently the initial lithospheric temperatures. All modeling presented here was conducted using the ANSYS finite element program. ANSYS employs the Newton-Raphson approach to solve nonlinear problems. In this method, a load was

Reverse Fault (MZRF) separates the Zagros mountain belt from Central Iran, which is a major structural discontinuity (Stocklin, 1968; Berberian, 1995).

The Zagros mountains are affected by the active NS trending Kazerun fault that offsets the folds and the lower Miocene terranes. Maximum and minimum displacement rates on the fault have been inferred from these offsets by Berberian (1981, 1995) and Authemayou (et al., 2005) to 15 and 4 mm yr⁻, respectively. Present-day activity of the Kazerun fault is evidenced by recent earthquakes with right lateral mechanisms located on the fault (Baker et al., 1993; Yaminifar et al., 2007). The main recent fault (MRF) is an active NW–SE trending right lateral strike-slip fault which runs along the MZT (Berberian, 1995) and is observed northwest of the Kazerun fault (Tchalenko & Braud 1974; Ricou et al., 1977). At the present time, shortening across the Zagros accommodates about 50 percent of the 22 mm/yr convergence between Arabia and Eurasia (Vernant et al., 2004).

Of particular interest is whether or not the earthquake depths in this region show any evidence for intracontinental subduction. Based on a microearthquake survey and receiver functions for the eastern part of the central Zagros Hatzfeld et al.(2003) suggested, a crustal structure consisting of an 11 km thick sedimentary layer overlying an 35 km thick crystalline crust. The crystalline crust consists of an upper layer extending from 11 to 19 km depth and a lower layer extending from ٠19 km depth, to a crust-mantle interface at 46 km depth (Figure2). Moreover, in that region most well-located microearthquakes occurred between 10 and 14 km depth, with none occurring deeper than 20 km (Tatar et al., 2004). Using the waveform-modelled depths in the Zagros, Talebian & Jackson (2004) found no earthquakes deeper than 20 km anywhere except near the Oman Line in the extreme SE Zagros. Beneath the Zagros there is no evidence in the form of mantle earthquakes for present-day active subduction of continental crust, with shortening apparently accommodated entirely by crustal thickening and distributed deformation (Talebian & Jackson 2004). Finally, the Zagros has many earthquakes, but their magnitudes are all less than Mw 7.0 and nearly all the moment release occurs near the SW topographic edge (i.e. elevations between (500–1000m) of the belt(Talebian & Jackson, 2004). More than 2000 instrumentally recorded earthquakes occurring in the Iran region during the period 1918–2004 have been relocated and reassessed by Enghdahl, 2006 with special attention to focal depth, using an advanced technique for 1D earthquake location. The results indicate that most earthquakes in the Iranian continental lithosphere occur in the upper crust, with the crustal shortening produced by continental collision accommodated entirely by thickening and distributed deformation. In the Zagros Mountains nearly all earthquakes are confined to the upper crust (depths <20 km), and there is no evidence for a seismically active subducted slab dipping NE beneath central Iran (figure.1A).

## 3- Background

A valuable tool which can help to genetically link the observations made at different crustal levels and to understand the spatial and temporal variations of stress and strain during lithospheric compression or extension are numerical simulations. Such models use a quantitative description of the first-order physical processes acting in the lithosphere to calculate rock deformation on the basis of thermal and mechanical properties. Numerical simulations are based on the finite element (FE) method as this technique allows the calculation of stresses, strains and temperatures in compositionally heterogeneous models with non-linear rheologies and complex geometries. Thermal modeling is based on heat transport by conduction, while the mechanical simulations combine various material laws, e.g., brittle failure and temperature- and strain rate-dependent creep, to approximate the rheology of the lithosphere. Using cross-sections and two-dimensional approaches, several convergent zones have been studied (e.g. Bird,1978, 1998) taking into account the rheological complexity of the lithosphere with plastic and viscous constitutive laws. Models in three dimensions have also been presented, but due to their numerical complexity they have been used mainly to describe simple geometrical settings (e.g.chery, 2001). For deformation of a large area such as the Tibetan plateau, a rheological simplification of the lithosphere is used with a thin viscous sheet model (e.g., England, 1986, 1997). However, this model does not take into account the plastic (or frictional) deformation of the lithosphere in the upper crust and possibly the uppermost mantle. Therefore, the hypothesis of a strong upper crust driving the collision kinematics cannot be tested. The thin viscous sheet model has been applied to Iran by Jackson, 1984 and Sobouti and Arkani-Hamed, 1996. Both of these studies concluded that deformation of the Iranian lithosphere is determined by the shape of the rigid boundaries and the disposition of the rigid blocks within the collision zone. Vernant and Chery (et al., 2006) using a simple 2 layer numerical modeling with linear Maxwell rheology show that the mechanical behavior of the Iranian lithosphere seems to be partly controlled by the large strike slip faults. Bird (1978) using 2-D finite element modeling of lithospheric deformation in the Zagros with pure nonelastic rheology showed that shear stress concentrates on the suture zone for all slab–pull boundary condition whatever the densities and flow laws considered. To account for distributed deformation in the Zagros crustal wedge, the Neo-tethys oceanic slab must be detached there, the Arbia-Iran convergence now occurring in a compressive regime. In this study, we used a three-dimensional finite element model to account for the rheological properties of the Zagros lithosphere, with temperature dependent rheological law and fault discontinuities.

## 4- Model setup

Velocity sections are used to define the structural boundaries of the model. The crust is divided in to the upper and lower crust. This division is

# Preliminary Results of Long Term Strength of Lithosphere in the Zagros Mountains of Iran: Insight from Mechanical Modeling

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## Abstract

A three-dimensional lithosphere model with horizontal dimensions of 1500 km×600 km and a depth extent of 70 km for the Zagros is constructed from available geophysical data to find out strength of the outermost layers in this area. The structural boundaries of the model are based on the results from the deep seismic sounding profiles. First the finite element model for the temperature is solved in order to obtain initial temperature and the geotherm, after that structural viscoelastic problem is solved using the same mesh as in the thermal initial condition. Preliminary results for wet and dry rheology indicate that the depth of the BDT is about 8 km and 11 km for hot geotherm and 10.5 km to14 km for cold geotherm. The results are in good agreement with focal depth in the Zagros that most earthquakes occur in 8 to 15 km depth (Tatar et al., 2004 and Jackson et al., 2008), that the long-term strength of the continental lithosphere resided only in its upper part, which was contained wholly within the crust.

Key words: Rheology, Finite-Element Methods, Brittle, Geotherm, Iran

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## 1- INTRODUCTION

The strength of earth's outermost layers has been a topic of debate ever since the turn of the last century when Joseph Barrell first introduced the concept of a strong lithosphere that overlies a weak fluid asthenosphere (Barrell, 1914), for example, predict that strength increases with depth and then decreases in accordance with the brittle (e.g., Byerlee, 1978) and ductile deformation laws. In oceanic regions, the envelopes are approximately symmetric about the depth of the brittle-ductile transition (BDT), where the brittle-elastic and elastic-ductile layers contribute equally to the strength. In the continents, the strength envelopes are more complex, and there may be more than one brittle and ductile layer (Bird, 1993; Burov & Watts, 2006). Brittle-ductile transition (BDT) is a regime in the lithosphere, above which rocks exhibit brittle deformation and below which they fail by plastic (or ductile) flow. This layer is probably gradual, where the deformation mechanism changes slowly. This zone is quite often correlated with the maximum depth of crustal earthquakes, as the earthquakes are assumed to be consequences of brittle fracture. However, the real lower boundary of crustal earthquakes is connected to changes in frictional behaviour in the fault zone from stick-slip to stable sliding (from velocity weakening to velocity strengthening) (Scholz, 1990). Correlation between the critical temperature and the BDT depth, together with a known distribution of earthquake focal depths, can be used for analysis and verification of the rheological models. In this study we constructed a 3D numerical model for the Zagros and adjacent NE reigon

based on the known deep structure and seismic data (Hatzfeld et al.,2003; Paul et al., 2006; Kavinai et al., 2007; Yaminifard et al., 2006; Yaminifard et al., 2007) to estimate of stress level at which the crust of Zagros is failured. Constructed model is first used in the calculation of the initial temperature for two scenarios (cold and hot) in the model and secondly in analysing the deformational conditions with the obtained rheology for each of them. In this paper the structure of the collision zone and faults will be represented by contact surface with the coulomb friction law.

## 2- Zagros, Geology and Tectonic Setting

Among the present active belts, the Zagros mountain fold-and thrust belt, located in Southwest Iran, is one of the simplest and most seismically active in the world. It results from the collision of the Arabian plate with the continental blocks of Central Iran, beginning in Miocene time and continuing today (e.g. Stocklin, 1968). The main feature of the Zagros is a linear, asymmetrical folded belt, which forms a 200–300 km wide series of ranges extending for about 1500 km from eastern Turkey to the Strait of Hormoz (Figure.1B). The Zagros contains a thick and almost continuous sequence of shelf sediments deposited on the 1–2 km thick infra-Cambrian Hormoz Salt formation. The sediments, of Paleozoic to Late Tertiary age, are believed to be separated from the Precambrian metamorphic basement by this Hormoz salt (Alavi 1994; Berberian & King1981). The Main Zagros

values of micrite and δw of Eocene seawater of 0.85 SMOW, shows that temperature was around 39˚C. Cathodluminescence studies of carbonate cements illustrated dull luminescence, because these carbonates afftected by both meteoric and burial diagenesis, and thus typical yellow to orange luminescence and zonations, evidence of meteoric diagenesis, are absent. This statement is confirmed by isotropic evidences.

# مطالعه رفتار مکانیکی سنگ‌کره در منطقه زاگرس با استفاده از مدل‌سازی عددی سه‌بعدی مکانیکی

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### چکیده

مطالعه رفتار  سنگ‌کره(سنگ‌کره) به عنوان یک لایه مکانیکی قوی که شامل پوسته و گوشته بالایی است از جمله مباحث روز علوم ژئودینامیک است. در این مقاله با استفاده از مدل‌سازی عددی سه بعدی به روش المان محدود و استفاده از یک ریولوژی ویسکوز وابسته به دما (خزشی) در دو حالت مرطوب و خشک رفتار مکانیکی سنگ‌کره در زاگرس با استفاده از دو ژئوترم گرم و سرد مورد بررسی قرار می‌گیرد. ستبرای پوسته بالا و پایین و همچنین قسمت گوشته بالایی بر اساس نگاشت‌های زلزله‌شناسی (Yaminifar, 2007; Paul, 2006;  Hatezfeld, 2003) در مناطق مختلف زاگرس در نظرگرفته شده است. گسل‌ها به صورت المان‌های ویژه در مدل جاسازی شدند و از قانون کولمب پیروی می‌کند. نتایج نشان می‌دهد که مرز تغییر شکل (BDT) برای ژئوترم سرد و ریولوژی مرطوب در ژرفای ۵،۱۰کیلومتری و برای ژئوترم سرد و ریولوژی خشک در ژرفای ۱۴کیلومتری رخ می‌دهد. همچنین این مرز تغییر شکل(BDT) برای ژئوترم گرم به ترتیب در ژرفای  ۸ و ۱۱ کیلومتری رخ می‌دهد. این نتایج تطابق خوبی با ژرفای زمین‌لرزه‌هایی که در منطقه زاگرس رخ می‌دهد (Tatar, 2004 ;Jackson, 2008) دارد و گویای استحکام  سنگ‌کره در قسمت پوسته متمرکز است.

**کلید واژه‌ها:** ریولوژی، المان محدود، سنگ‌کره، ژئوترم، زاگرس

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