RELATIONSHIPS BETWEEN STRATIGRAPHY, DEFORMATION AND THERMAL HISTORY IN SEDIMENTARY BASINS. IMPACT OF GEODYNAMIC CONCEPTS IN PETROLEUM EXPLORATION

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The natural processes that generate petroleum accumulations in a sedimentary basin require several ingredients: (1) the petroleum system elements: source, reservoir, seal and overburden rocks, which are the result of sedimentation processes in a subsiding basin; (2) petroleum traps, which in many cases are the result of deformation and (3) heat to convert suitable organic matter into petroleum. Although these different phenomena are considered independent at the scale of an oil field, at the lithosphere scale (1) thermal phenomena, (2) vertical movements of the earth surface responsible for sedimentation and erosion; and (3) tectonic deformation are not independent phenomena, they are intimately related by physical quantitative laws. These mutual inter-relationships are useful in petroleum exploration to predict one factor having knowledge of the others. Applications of these concepts can contribute to understand the tectonic history of complex areas, such as the Colombian sedimentary basins, and reduce exploration risk.

Los procesos naturales que generan acumulaciones de petróleo en una cuenca sedimentaria requieren varios ingredientes: (1) los elementos de sistema de petrolífero: roca fuente, roca almacenadora, roca sello y rocas de sobrecarga, los cuales son el resultado de procesos de sedimentación en una cuenca que subside; (2) las trampas de petróleo, la cuales en muchos casos son el resultado de deformación y (3) el calor para convertir la materia orgánica apropiada en petróleo. Aunque estos diferentes fenómenos son considerados independientes a la escala de un campo de petróleo, a la escala de la litosfera (1) los fenómenos termales, (2) los movimientos verticales de la superficie terrestre responsables de la sedimentación y la erosión; y (3) la deformación tectónica no son fenómenos independientes. Ellos están íntimamente relacionado por leyes físicas cuantitativas. Estas interrelaciones mutuas son útiles en la exploración de petróleo para predecir un factor a partir del conocimiento de los otros. La aplicación de estos conceptos puede contribuir a entender la historia tectónica de áreas complejas, tales como las cuencas sedimentarias colombianas, y así reducir el riesgo exploratorio.

Keywords: *Geodynamics, petroleum systems, stratigraphy, lithosphere deformation, thermal history.*

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INTRODUCTION

The natural processes that generate a petroleum accumulation in a sedimentary basin require several ingredients (Figure 1): (1) a petroleum source rock, a reservoir rock, a seal rock and overburden rock (Magoon and Dow, 1994) which are part of the sedimentary fill of the subsiding basin and their understanding require knowledge of the stratigraphy; (2) a petroleum trap (i.e. Biddle and Wielchowsky, 1994) which in many cases resulted from deformation of the sedimentary basin fill and its understanding require knowledge of the structural geology and (3) heat to convert suitable organic matter into petroleum (i.e. Deming, 1994; Horsfield, and Rullkötter, 1994). These three different variables traditionally are considered independent at oil field scale (i.e. White, 1980; Allen and Allen, 1990). However, at the scale of the entire sedimentary basin (lithosphere scale) they are intimately related by important geodynamical processes (i.e. Turcotte and Schubert, 1982; Cloetingh *et al.*, 1993, 1994). From these relationships important information useful to petroleum exploration can be extracted. This paper focuses on these relationships. First we will examine what the lithosphere is. Second we will examine each one of the variables (1) heat input into the lithosphere, (2) subsidence/uplift of the lithosphere responsible for sedimentation/erosion and (3) deformation of the lithosphere. Finally we will examine the mutual relationships between these variables and highlight the use of geodynamics in petroleum exploration.

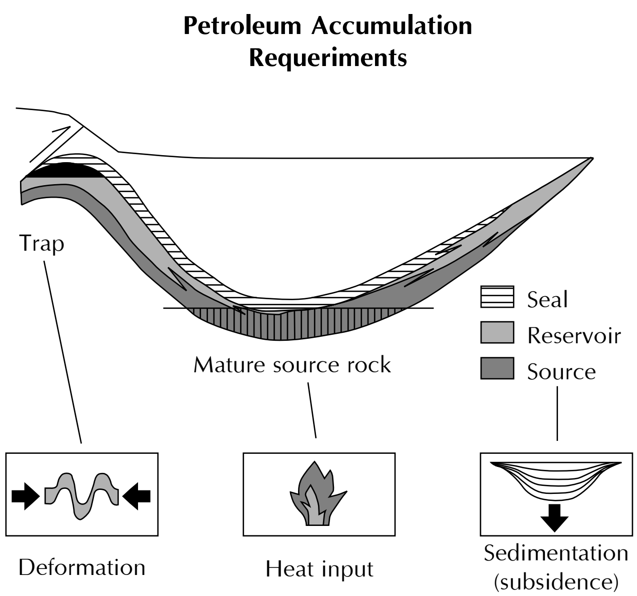

Figure 1. Petroleum accumulation requirements. The natural processes that generate a petroleum accumulation in a sedimentary basin require: (1) a petroleum source rock, a reservoir rock, a seal rock and overburden rock which are part of the sedimentary fill of the subsiding basin; (2) a petroleum trap which in many cases resulted from deformation of the sedimentary basin fill and (3) heat to convert suitable organic matter into petroleum

THE LITHOSPHERE

The lithosphere is the outer «rigid» layer of the earth, which is fragmented into a number of «rigid» tectonic plates floating on a viscous asthenosphere; the latter on geological time intervals behaves as a fluid. According to the plate-tectonics theory, thermally driven convection currents is one of the causes of the lithosphere plates relative movement (i.e. Ziegler, 1993) in such a way that the heat flow from the earth's interior is the ultimate source of energy for plate-tectonics movements. Contrary to the compositional subdivision between crust and mantle, the lithosphere concept (Figure 2) is defined on basis of its thermal and mechanical properties contrasting with those of the asthenosphere: lithosphere is a «rigid» cold layer contrasting with a viscous hot asthenosphere. The lower boundary of the lithosphere has been approximately considered at the isotherm 1333°C. Crust and mantle concepts are defined on the basis of lithological, mineralogical and chemical composition. The crust is made of silica rich rocks with an average composition similar to that of normal igneous rocks. Continental crust composition is similar to that of felsic to intermediate granitic rocks and oceanic crust composition is similar to that of mafic basaltic rocks. In contrast upper mantle rocks contain less silica and more iron and magnesium with a composition similar to ultramafic rocks. The lower boundary of the crust is the Mohorovicic discontinuity usually recognized by density and seismic wave velocity contrasts. Lithosphere includes the crust and the uppermost mantle (Figure 2).

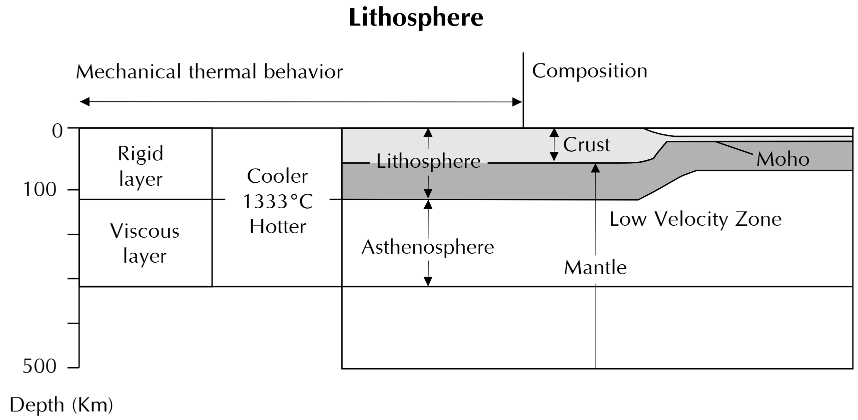

Figure 2. The lithosphere. Contrary to the compositional subdivision between crust and mantle, the lithosphere is defined on basis of its thermal and mechanical properties contrasting with those of the asthenosphere: lithosphere is a «rigid» cold layer contrasting with a viscous hot asthenosphere. The lower boundary of the lithosphere has been approximately considered at the isotherm 1333°C. Lithosphere includes the crust and the uppermost mantle

THERMAL. SUBSIDENCE/UPLIFT AND DEFORMATION PROCESSES IN THE LITHOSPHERE

Thermal phenomena

Many observations (i.e. deep wells) indicate that temperature increases with depth (i.e. Chapman, 1986). While heat convection is considered the main heat transport mechanism for the asthenosphere, heat conduction is considered the main heat transport mechanism for the lithosphere. For both oceanic and continental lithosphere the geotherms or temperature-depth curves are solutions for the Fourier's law of heat conduction

$$\rho C_p \frac{\partial T}{\partial t} = \nabla \cdot \left(k \nabla T\right) + A \qquad (1)$$

where T is temperature, t is time, and with the material parameters: ρ density, C_p heat capacity at constant pressure, k the coefficient of heat conduction, and A the radioactive heat production per volume unit. Material parameters are assumed to remain constant within the major crustal and subcrustal layers, except for the heat production A, for which in some cases an exponential decay with depth is usually assumed.

There are several processes that may affect the temperature distribution locally or regionally, including erosion and sedimentation (thermal blanketing), among others. We will discuss it latter. In general, it is the most recent thermal event that controls the thermal structure of the continental lithosphere.

If plate-tectonics is thermally driven by convective heat transfer from the earth's interior, heat has profound relationships with tectonics. In order to illustrate the effect of thermal phenomena in tectonic processes I will present two examples: the first is the generation of new lithosphere in an oceanic ridge and the second is in a rifted extensional basin.

Example 1 (Figure 3). An oceanic ridge axis is considered the place where new oceanic lithosphere is formed as «frozen» differentiated asthenosphere. The outward movement of the oceanic plates apart from the ridge axis allows the hot upwelling asthenosphere reach the sea bottom and freeze by rapid cooling indicated by an elevated heat flow to the sea bottom (i.e. Parsons and Sclater, 1977). As the new plate gradually moves as it cools down, it gradually displaces the 1333°C isotherm down decreasing its geothermal gradient and increasing lithosphere thickness. Also upward heat flow to the sea bottom decreases away from the ridge. Gradual cooling of the lithosphere leads to a gradual density increase responsible for gradual loss of buoyancy and a gradual decrease in elevation of the ridge submarine mountains relative to the ocean floor, a phenomenon known as thermal subsidence.

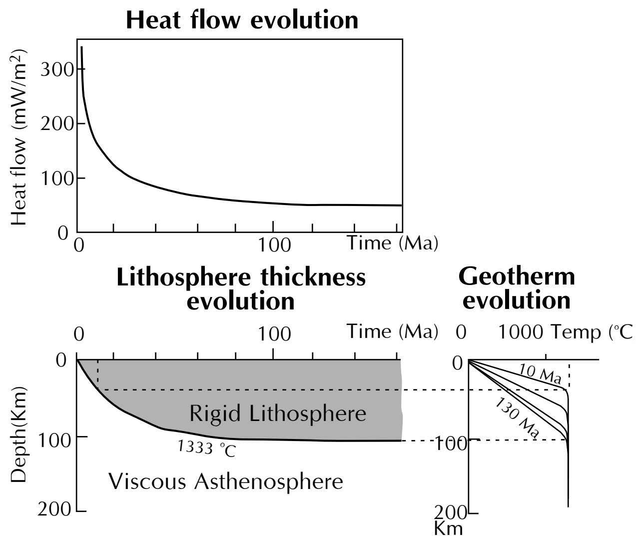

Figure 3. Thermal and thickness evolution of the oceanic lithosphere (after Parsons and Sclater, 1977). At an oceanic ridge axis new oceanic lithosphere is formed as «frozen» differentiated asthenosphere. As the new plate gradually moves as it cools down, it gradually displaces the 1333°C isotherm down decreasing its geothermal gradient and increasing lithosphere thickness. Gradual cooling of the lithosphere leads to a gradual density increase responsible for gradual loss of buoyancy and a gradual decrease in elevation of the ridge submarine mountains relative to the ocean floor, a phenomenon known as thermal subsidence

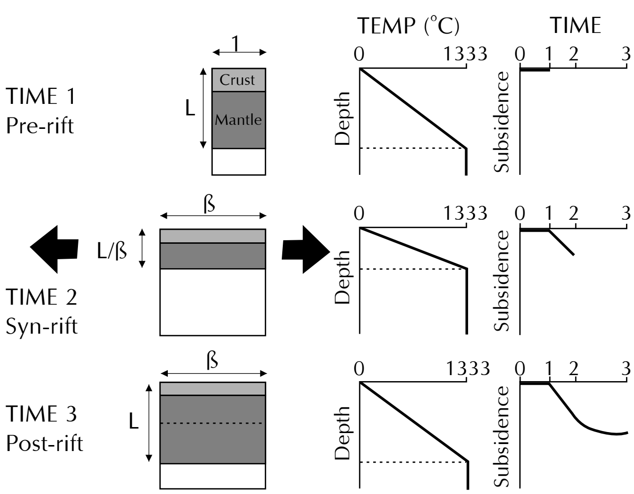

Figure 4. Instantaneous lithosphere stretching model (McKenzie, 1978). Assuming that the volume of lithosphere is preserved, if lithosphere extends horizontally by a factor β its thickness decreases by a factor 1/β. As the lower boundary of the lithosphere is the 1333°C isotherm such lithosphere thinning implies a shallowing of this isotherm and an increase in the geothermal gradient. The isostatic consequence of this rapid thinning of the lithosphere is fast subsidence during the rifting event. A consequence of this replacement of lithosphere material at depth by hotter asthenosphere is the elevated heat flow observed in rift basins. Then gradually thermal equilibrium is reached and heat dissipation leads to gradual cooling and gradually decreasing thermal subsidence

Example 2 (Figure 4). During lithosphere stretching extensional basins are generated by rifting. Assuming that the volume of lithosphere is preserved, if lithosphere extends horizontally by a factor β its thickness decreases by a factor 1/β (McKenzie, 1978). As the lower boundary of the lithosphere is the 1333°C isotherm such lithosphere thinning implies a shallowing of this isotherm and an increase in the geothermal gradient. The isostatic (or hydrostatic) consequence of this rapid thinning of the lithosphere is fast subsidence during the rifting event. A consequence of this replacement of lithosphere material at depth by hotter asthenosphere is the elevated heat flow observed in rift basins. Then gradually thermal equilibrium is reached and heat dissipation leads to gradual cooling and gradually decreasing thermal subsidence. Finally when the equilibrium is obtained the lithosphere thickness attains its original value, however crustal thickness does not increase. As density of the mantle lithosphere is greater than crust density the net effect is a density increase of the lithosphere.

Subsidence/uplift effects

Vertical movements of the lithosphere –subsidence or uplift– (Figure 5) lead to generation of (1) subsiding areas mostly known as sedimentary basins, which are usually filled by sediments or (2) uplifted areas (the most spectacular of them known as mountains), where usually erosion removes material away. If base level is defined as the surface that separates areas of dominant sedimentation located below the base level surface and areas of dominant erosion located above, the base level surface is located between sea level and earth surface. Base level surface mimics earth surface. Vertical movements of the lithosphere produce topographic differences of the earth surface. In conclusion interaction of the lithosphere with the hydrosphere, atmosphere and biosphere is responsible of sedimentation in areas dominated by subsidence and erosion in areas dominated by uplift. The sedimentary fill including unconformities recorded subsidence and uplift movements of the lithosphere. The stratigraphic record can be used to infer the history of vertical movements of the lithosphere. Fission track data, vitrinite reflectance data and pressure-temperature-time paths data from metamorphic rocks are also useful to infer the history of vertical movements of the lithosphere (i.e. Green *et al.*; England and Thompson, 1986).

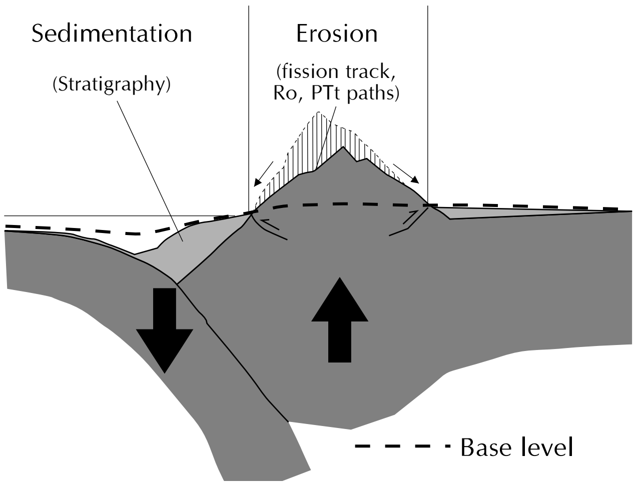

Figure 5. Subsidence (sedimentation) and Uplift (erosion). Vertical movements of the lithosphere lead to generation of (1) subsiding areas mostly known as sedimentary basins, which are usually filled by sediments or (2) uplifted areas, where usually erosion removes material away. Base level is defined as the surface that separates areas of dominant sedimentation located below the base level surface and areas of dominant erosion located above the base level surface

Lithosphere deformation

Deformation of the lithosphere is mainly the result of stresses induced by horizontal relative movements of adjacent lithosphere plates. Deformation styles (Figure 6) have been classified as dominantly (1) extensional, (2) compressional or (3) transcurrent, although any combination of them can occur in nature. Deformation is the consequence of stresses, however, the magnitude and type of deformation produced by the same stress is not the same in different materials, this is expressed in other words saying that different materials have different rheological behavior (Figure 7). Quantitative relationships between stress and strain for a particular material define the rheological behavior of that material. Two major types of rheological behavior can be defined: elastic and plastic depending if deformation is recoverable or not. When starting from zero a gradually increasing stress is applied to a rock resulting with gradually increasing strain that is developed where its magnitude is proportional to the applied stress. Then if the stress is eliminated strain also disappears; this behavior of recoverable deformation is called elastic. Transmission of seismic waves trough the rocks is an example of elastic behavior. Elastic behavior of rocks is exhibited at relatively low stress. However, if stresses are increased beyond a certain value, called the elastic limit, deformation is not completely recoverable, that is after eliminating the applied stress, strain is not completely eliminated and a permanent deformation has been produced (Figure 7). This is called inelastic or plastic behavior. Folds, fractures and faults are examples of inelastic permanent deformation. Among the

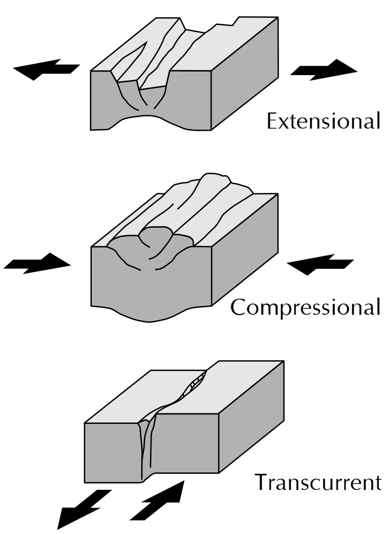

Figure 6. Deformation styles of the crust. Brittle deformation affects the upper crust and ductile deformation affects the lower crust.

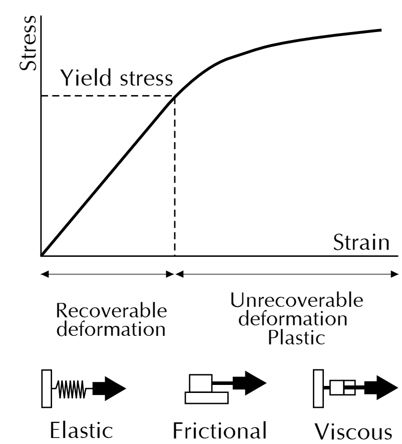

Figure 7. Strain stress relationship. Elastic vs. plastic deformation. An idealized elastic model is a spring; in this model stress is represented by the force applied to the spring and strain is represented by deformation of the spring. Two idealized plastic models are the frictional and viscous models. The frictional model is represented by a rigid body frictional displacement on a surface; this model is time independent. A loosely fitting piston in a liquid filled cylinder represents the viscous ideal model. Any slow displacement of the piston requires fluid flow between piston and cylinder walls. The force (i.e. stress) required to move the piston increases with the rate of piston movement (after Mandl, 1988)

several types of inelastic rheological behavior exhibited by lithosphere rocks brittle frictional fracture and ductile steady-state creep flow are particularly important. The rheological behavior of rocks that constitute the continental lithosphere is derived from laboratory experiments and subsequently extrapolated to geologically relevant times. The following summary about lithosphere rheology has been partially extracted from Beekman (1994) and Van Wees (1994).

Brittle frictional rheology. At low confining pressures and low temperatures, such as those of the upper crust, brittle frictional fracture (Figure 8) is predominant (strain-rate independent deformation). Fracturing is described by a Mohr-Coulomb criterion (i.e. Mandl, 1988). If at low confining pressures and low temperatures a rock sample is compressed in one direction shear stress is developed within the sample at a 45° angle relative to the applied principal maximum stress σ_{max}. Magnitude of this shear stress is proportional of the stress difference $\sigma_{max} - \sigma_{min}$. In an experiment done at the earth surface minimum stress σ_{min} is the atmospheric pressure. If the applied stress σ_{max} gradually is increased the stress difference $\sigma_{max} - \sigma_{min}$ also increases and the shear stress within the rock increases. This relationship between normal stress difference and shear stress is graphically represented by the Mohr circle, in this representation if the minimum normal stress remains constant and the applied maximum normal stress increases the circle representing the stress state increases and the maximum shear stress also increases as the radius of the circle (Figure 8). With gradually increasing shear stress brittle fracture is generated when the Mohr-Coulomb criterion is reached, which graphically is represented when the growing Mohr circle touches the stress envelope, described by

$$|\tau| = \tau_o + \sigma \tan \varphi \qquad (2)$$

where τ is the shear stress, σ is the normal effective stress, the angle φ is usually referred to as "the angle of internal friction" and τ_o is the «cohesive shear stress»;

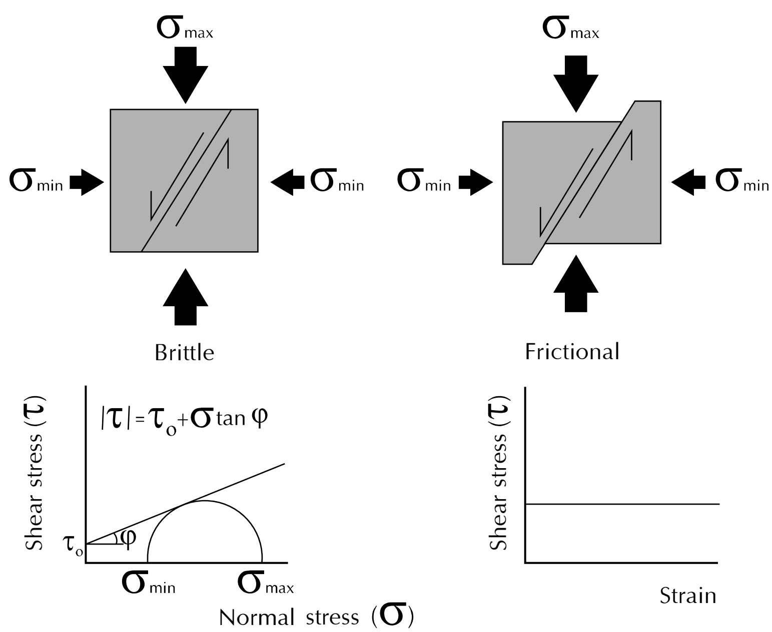

Figure 8. Mohr-Coulomb criterion of brittle-frictional behavior. If at low confining pressures and low temperatures a rock sample is compressed in one direction. Shear stress is developed within the sample at a 45° angle relative to the applied principal maximum stress σ_{max}. Magnitude of this shear stress is proportional of the stress difference $\sigma_{max} - \sigma_{min}$. This relationship between normal stress difference and shear stress is graphically represented by the Mohr circle, in this representation if the minimum normal stress remains constant and the applied maximum normal stress increases the circle representing the stress state increases and the maximum shear stress also increases as the radius of the circle. With gradually increasing shear stress brittle fracture is generated when the Mohr-Coulomb criterion is reached, which graphically is represented when the growing Mohr circle touches the stress envelope, described by the equation, where τ is the shear stress, σ is the normal effective stress, the angle φ is usually referred to as "the angle of internal friction" and τ_o is the «cohesive shear stress»; φ and τ_o are material parameters that characterize the rock. After this brittle fracture any further deformation will be frictional displacement along the fracture, which prevents further stress increase

φ and τ_o are material parameters that characterize the rock. After this brittle fracture any further deformation will be frictional displacement along the fracture, which prevents further stress increase. Following Byerlee (1978), the Mohr-Coulomb criterion (Figure 9) can be expressed in terms of effective principal stress difference, lithostatic overburden pressure, and pore fluid pressure (Ranalli, 1995) expressing the brittle strength of the rock $\sigma_{brittle}$ as:

$$\sigma_{brittle} = \sigma_1 - \sigma_3 = \alpha\rho gz(1-\lambda) \qquad (3)$$

where σ_1 and σ_3 are maximum and minimal principal stresses at the onset of faulting, ρ is density, g is gravitation acceleration, z is depth, and where α depends on the fault type and the sliding friction coefficient, λ is the ratio between the pore fluid pressure and the lithostatic pressure known as hydrostatic pore fluid factor. We should remember that ρgz is the lithostatic pressure (vertical downward stress resulting from the weight of the rock column), which increases with depth z, therefore at the low confining pressures and temperatures of the upper crust the rock brittle strength increases linearly with increasing depth (Figure 9).

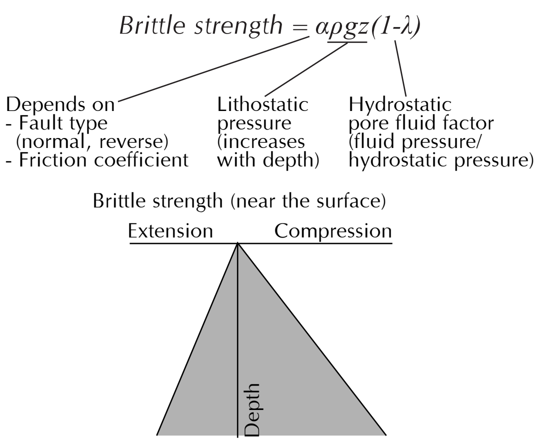

Figure 9. Brittle strength dependence on depth. The Mohr-Coulomb criterion can be expressed in terms of effective principal stress difference, lithostatic overburden pressure, and pore fluid pressure expressing the brittle strength of the rock $\sigma_{brittle}$ as:

$$\sigma_{brittle} = \sigma_1 - \sigma_3 = \alpha\rho gz(1-\lambda)$$

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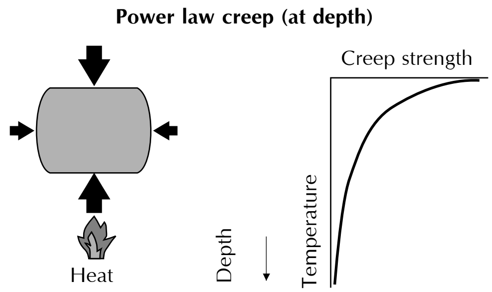
Ductile rheology. The ductile phenomena (Figure 10) are, by definition, negligible in the brittle range near the earth surface, but depending on the rock type, they become increasingly important at increasing depth at elevated temperatures, pressures at low strain rates of what is usually referred to as the ductile range. One particular type of ductile behavior known as steady-state creep (Figure 10, strain-rate dependent deformation) is believed to occur in the lower crust, that is described by a ductile flow law. In contrast to discontinuous deformation produced by brittle faulting this type of continuous deformation can be described as a flow, showing some similarities with a liquid flow. The transition zone between brittle and ductile behavior is primarily controlled by an increase in temperature which activates microscopic creep processes. Experiments with rock-forming minerals show that the critical principal differential stress necessary to maintain a given steady-state strain rate, is a function of a power of the strain rate and varies strongly with mineral composition and temperature. The ductile flow obeys the so called power-law creep (Kirby, 1983):

$$\sigma_{creep} = \sigma_1 - \sigma_3 = \left[\frac{\dot{\varepsilon}}{A_P}\right]^{\frac{1}{n}} exp\left[\frac{E_P}{nRT}\right] \qquad (4)$$

Where $\dot{\varepsilon}$ is strain rate A_P, n and E_P are flow parameters, T is absolute temperature and R is the gas constant. The above power-law creep indicates that critical differential principal stresses or rock strength decreases exponentially with increasing temperature. Flow parameters for various rock-forming minerals show that power-law creep flow parameters are strongly controlled by silica content (e.g, Carter and Tsenn, 1987). Felsic rocks (e.g, granite) show low critical differential stress values compared to mafic rocks (e.g, dunites), under similar conditions of strain rate and temperature. However, above a critical stress, the power-law creep breaks down in the so-called low-temperature plasticity, which is characterized by a nearly linear increase in stress with decreasing temperature (Dorn law, Goetze and Evans, 1979). In conclusion as temperature increases with depth, in the ductile range of the lower crust rock strength decreases with depth



$$\text{Creep strength} = \left(\frac{\text{strain rate}}{A_P}\right)^{1/n} exp\left(\frac{E_P}{nRT}\right)$$

Temperature

Figure 10. Ductile Power-law creep strength dependence on depth. The ductile flow obeys the so called power-law creep, where A_P, n and E_P are flow parameters, T is absolute temperature and R is the gas constant. The above power-law creep indicates that critical differential principal stresses or rock strength decreases exponentially with increasing temperature. As temperature increases with depth, in the ductile range of the lower crust rock strength decreases with depth

(Figure 10). Also ductile rock strength increases with silica decrease, therefore mantle rock strength should be larger than crust strength if other conditions are equal, as mantle silica content is lower than crust silica content.

Rheological profiles and integrated strength of a stratified lithosphere. Using the relationships expressed in equations (3) and (4) it is possible to calculate the brittle and ductile creep rock strength at any depth (Figures 11 and 12). For a given tectonic environment (thrusting, normal faulting or strike-slip faulting), at any depth, flow properties, temperature and strain rate, the lowest of the brittle and creep principal stress difference given by equations (3) and (4) respectively, gives a rheological rock strength (called the yield strength; i.e. Ranalli, 1995; Okaya *et al*., 1996; Cloetingh and Burov, 1996). Equation 3 shows that brittle strength is low at earth surface and increases with depth (as it is proportional to lithostatic pressure that increases with depth), Equation 4 shows that ductile creep strength decreases with depth (as it exponentially decreases with increasing temperature, which increases with depth). Near the surface brittle yield strength is minimum but ductile yield strength is maximum, therefore near the surface tectonic stresses easily overcome the brittle yield strength but not the ductile yield strength which is very high near the surface (Figure 11). This explains why brittle deformation prevails in the upper crust. Conversely at great depth brittle yield strength is maximum but ductile yield strength is minimum, therefore at great depth tectonic stresses at low strain rates as those affecting the lithosphere, easily overcome ductile yield strength but not brittle yield strength (Figure 11). This explains why ductile flow is the dominant deformation mechanism of the lower crust. For stress differences below this yield strength, plastic deformation at the imposed strain rate will not occur. The dependence of the principal stress differences in equations (3) and (4) on external conditions (pressure and temperature) and material properties (mineralogical composition), both varying with depth within the lithosphere, makes the yield strength also vary with depth, thus constituting a strength envelope (Figure 12).

Continental crust, with a thickness in the order of several tens of kilometers, is layered, with large variations between different tectonic provinces. A simple and representative crustal mineralogical model is a two layered crust with felsic rocks dominating the upper crust and mafic rocks dominating the lower crust, as indicated by seismic velocities. These crustal layers

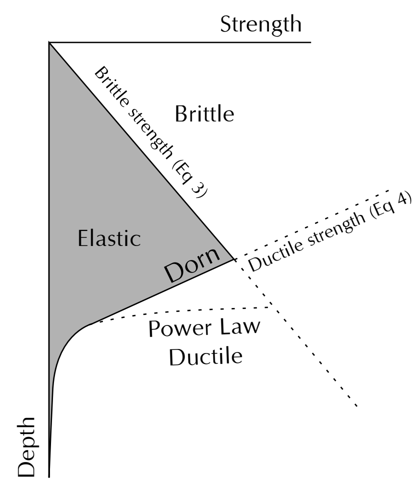


Figure 11. Brittle and Ductile power-law creep strength dependence on depth (after Ranalli, 1995). Near the surface brittle yield strength is minimum but ductile yield strength is maximum, therefore near the surface tectonic stresses easily overcome the brittle yield strength but not the ductile yield strength which is very high near the surface. This explains why brittle deformation prevails in the upper crust. Conversely at great depth brittle yield strength is maximum but ductile yield strength is minimum, therefore at great depth tectonic stresses at low strain rates as those affecting the lithosphere, easily overcome ductile yield strength but not brittle yield strength. This explains why ductile flow is the dominant deformation mechanism of the lower crust

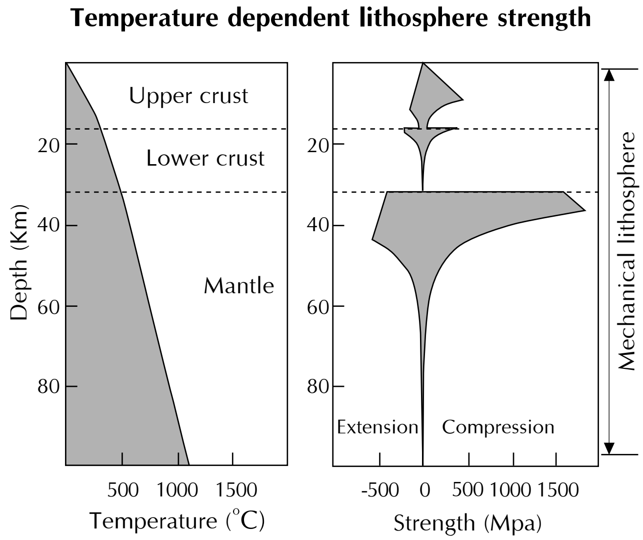


Figure 12. Temperature and depth dependent lithosphere strength profile (after Cloetingh and Burov, 1996). The profile shows a marked layering of a relatively strong (mostly brittle) quartzite upper crust, a weak (mostly ductile) diorite lower crust, and a strong (brittle and ductile) olivine subcrustal lithosphere mantle. The dependence of the principal stress differences in equations (3) and (4) on external conditions (pressure and temperature) and material properties (mineralogical composition), both varying with depth within the lithosphere, makes the yield strength also vary with depth

overly ultramafic upper mantle material composed primarily of olivine. For a typical continental lithosphere composed of a quartzite upper crust, a diorite lower crust, and an olivine mantle, the associated rheological layering is shown in Figure 12. Lithosphere strength envelopes typical for Phanerozoic extension and inversion settings show a marked layering of a relatively strong (mostly brittle) upper crust, a weak (mostly ductile) lower crust, and a strong (brittle and ductile) subcrustal lithosphere (Figure 12). This layering, predicted from extrapolation of rock mechanics, agrees quite well with interpretation of geophysical and geological data. This rheological layering is confirmed by the depth distribution of seismic activity, which is approximately confined to the crustal layers (e.g, Cloetingh and Banda, 1992), as well as by the presence of minima in seismic wave velocity and electrical resistivity coinciding with the ductile layers (Ranalli and Murphy, 1987). The rheological stratification may significantly affect the thermo-mechanical response of the continental lithosphere to stresses. For example, large normal or reverse faults in the upper crust tend to flatten out towards a detachment zone in the lower weak part of the upper crust (Braun and Beaumont, 1987).

HOW SUBSIDENCE/UPLIFT. DEFORMATION AND THERMAL PHENOMENA DO INTERACT AT LITHOSPHERE SCALE?

In this section we will examine with some examples the mutual relationships between subsidence/uplift, deformation and thermal phenomena (Figures 13 and 14). However, the selected examples are only to illustrate these relationships and they are not unique. Other different possible relationships not mentioned here may exist.

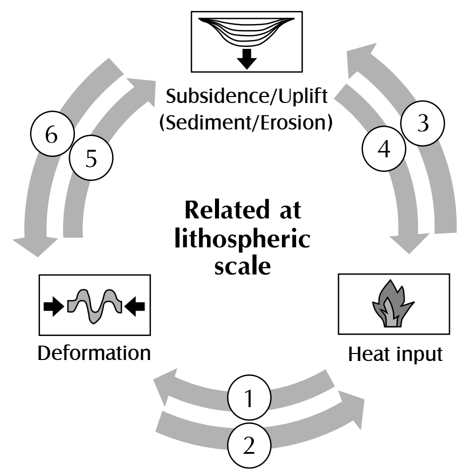


Figure 13. Lithosphere scale relationships between Thermal, Deformation and Subsidence (sedimentation)/Uplift (erosion) phenomena. The numbered arrows correspond to the numbered parts of figure 14. The numbered relationships are explained in the text paragraphs identified by the same numbers

(1) Thermal effects on deformation (Figure 14:1)

We have said that if plate-tectonics is thermally driven by convective heat transfer from the earth's interior, heat has profound relationships with tectonics. Additionally temperature plays an important role in the mechanical response of the lithosphere to stresses. Of particular importance is the variation of temperature with depth on the ductile strength of lithosphere rocks. When a metal is heated it becomes malleable, in a similar way heating of rocks reduces its strength. Heating of the lithosphere reduces the ductile strength of the lithosphere making it prone to deformation (i.e. Kooi, 1991; Kooi *et al*., 1992; Van Wees, 1994; Van Wees and Beekman, 2000). Conversely conductive cooling of the lithosphere over a relatively short period of a few mil-

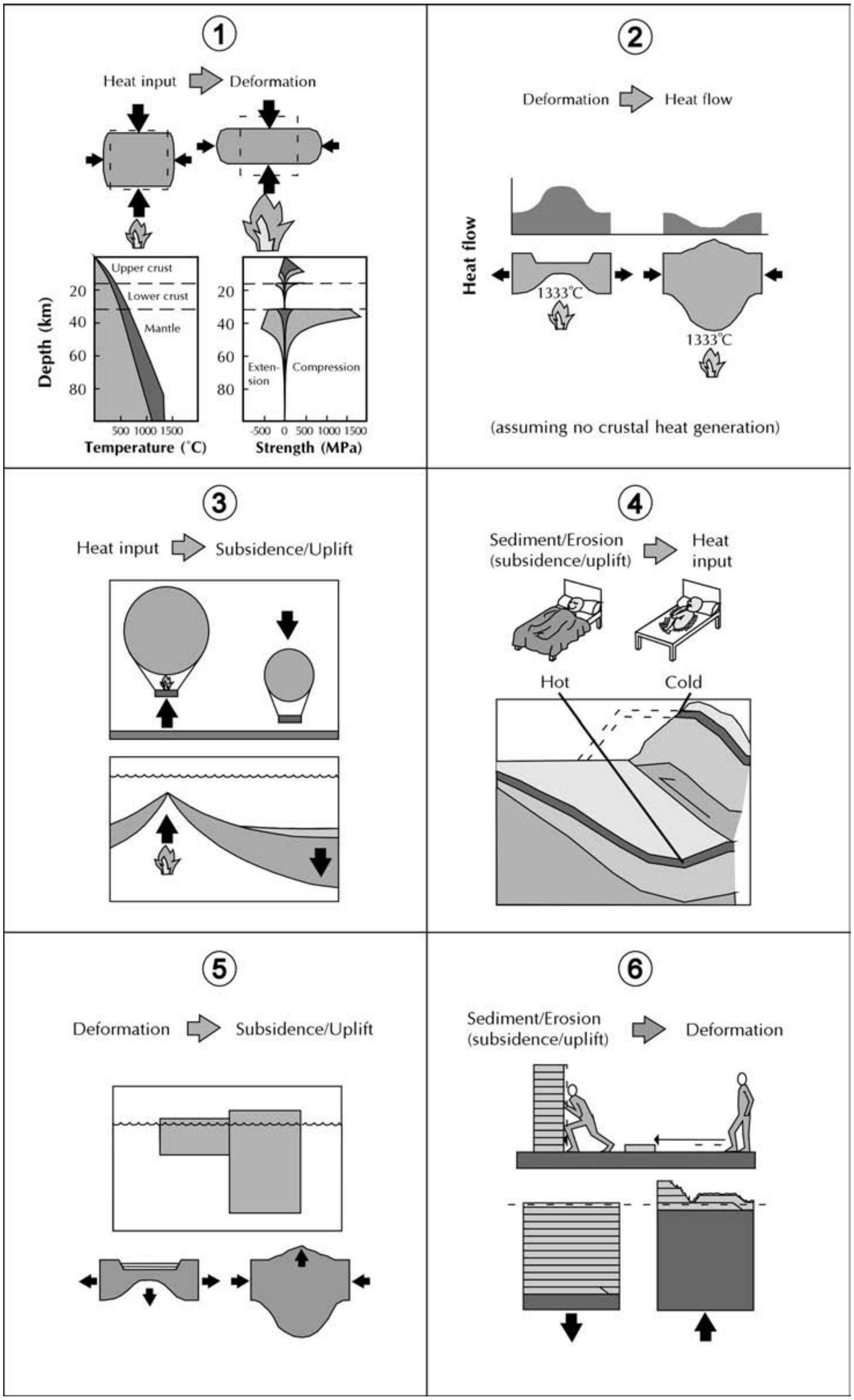


Figure 14. Lithosphere scale relationships between Thermal. Deformation and Subsidence (sedimentation)/Uplift (erosion) phenomena. The numbered parts of this figure correspond to the numbered arrows of Figure 13. The numbered relationships are explained in the text paragraphs identified by the same numbers.

lion years after the last heating event rapidly leads to a significant increase in the mechanical strength of the lithosphere. Over longer periods cooling also leads to thickening of the lithosphere, and of its mechanically strong sub-layers.

(2) Thermal effects of deformation (Figure 14:2)

Compressional shortening of the lithosphere mechanically increases its thickness as occurs at many mountain ranges (i.e. Kaila, 1981; Choukroune *et al*., 1989; Beaumont *et al*., 1994, 2000). Conversely extensional stretching of the lithosphere mechanically reduces its thickness as occurs in rifted extensional basins (i.e. Ziegler, 1994, Figure 14:2). These are examples that demonstrate that deformation can produce changes of thickness and the thermal structure of the lithosphere (i.e. Ter Voorde, 1996). We have seen the example of a rifted basin where thinning of the lithosphere implies a shallowing of its lower boundary: the 1333°C isotherm and an increase in the geothermal gradient. If the heat flow at the surface of the earth results from conductive heat transport from the hot astenosphere, mechanically driven changes of lithosphere thickness would change geothermal gradient and heat flow to the earth's surface. Additionally mechanical deformation can increase or decrease the thickness of radioactive heat generation upper crust layers changing the amount of radioactive heat generated within the crust. It is not surprising that areas of anomalous thick or thin lithosphere are also areas of anomalous heat flow. Thermal data can be an indicator of deep lithosphere phenomena (i.e. Sacks and Secor, 1990). At a local scale the effects of deformation on thermal history (and on organic matter maturity and petroleum generation) in thrusted terranes are well known: the burial of inmature sediments beneath a thrust sheet may result in sufficient heating to generate hydrocarbons (Angevine and Turcotte, 1983; Brower, 1981; Roure and Sassi, 1994; Sarmiento *et al*., 1997).

(3) Thermal effects on subsidence/uplift (Figure 14:3)

Heating of the gas in a balloon induces a vertical upward movement of the balloon because increase in temperature reduces the gas density increasing the buoyancy of the balloon. In a similar way heating of the lithosphere increases its buoyancy inducing uplift (i.e. Sleep, 1971). Conversely cooling reduces buoyancy of the lithosphere. We have mentioned the submarine mountain topography of an oceanic ridge system associated with young hot lithosphere and its gradual thermal subsidence away from the ridge axis produced by its gradual cooling. The uplift/subsidence signal extracted from the sedimentary record may reflect the thermal history of the lithosphere.

(4) Thermal effects of sedimentation (subsidence)/ erosion (uplift) (Figure 14:4)

While sleeping we keep warm covering our body with blankets, if we do not use blankets we get cold. In a similar way sedimentation prevents escape of heat to the earth's surface, reducing heat flow at the earth's surface: the so-called thermal blanketing effect in a sedimentary basin. Accumulation of thick sedimentary section favors keeping petroleum source rock units at elevated temperature favoring organic matter thermal maturity and petroleum generation (i.e. Deming, 1994; Horsfield and Rullkötter, 1994). These layers of sediments overlying petroleum source rocks and favoring petroleum generation are known as overburden rock, which is one of the elements of a petroleum system (Magoon and Dow, 1994). We have mentioned the higher thermal maturity in thrust belts may be related to the extra thickness of thrust sheets overburden rock (Angevine and Turcotte, 1983; Brower, 1981; Roure and Sassi, 1994; Sarmiento *et al*., 1997). Indicators of thermal history such as organic matter maturity indicators (i.e. vitrinite reflectance), of common use in the oil industry, are useful to estimate thickness of overburden rock in the past. The sedimentary record can be used to infer information about the thermal history of sedimentary basins. Conversely indicators of the thermal history of a rock sample (vitrinite reflectance, fission track; i.e. Green *et al*., 1995) or pressure-temperature-time paths of metamorphic rocks (England and Thompson, 1986) can be used not only to infer the thermal history, but also to estimate uplift and exhumation.

(5) Effects of deformation on subsidence/uplift (Figure 14:5)

If lithosphere volume remains approximately constant shortening or extension produced by horizontal deformation should produce changes in its thickness. We have seen how deformation can produce changes of thickness of the lithosphere: examples of this are lithosphere thickening during orogenic collisional compression or lithosphere thinning during lithosphere extensional stretching (i.e. Ziegler *et al*., 1995, 1998). In a similar way that the height of an iceberg above sea level is an indication of its depth below sea level: the higher the top of the iceberg the deeper its bottom, isostasy applies to the lithosphere. Changes of lithosphere thickness produced by deformation isostatically induce vertical movements of the earth's surface (i.e. Cloos, 1993). This explains why subsidence/sedimentation is common in areas of thinned lithosphere such as basins produced by lithosphere extensional stretching, and uplift/erosion is common in areas of thickened lithosphere such as mountain belts produced by lithosphere collisonal compression. Additionally at the local scale the effects of deformation on local uplift/erosion and subsidence/sedimentation in piggyback sedimentary basins are well known (i.e. Zoetemeijer, 1993; Zoetemeijer *et al*., 1993).

(6) Effects of sedimentation (subsidence)/erosion (uplift) on deformation (Figure 14:6)

Sedimentation (subsidence) or erosion (uplift) can have an important effect on deformation. I will illustrate this with an example. If we accumulate a lot of boxes in a big pile, the pile becomes very heavy and it is very difficult to slide the pile on the floor. However if we remove boxes from the pile it becomes so light that with little effort we can slide it on the floor. Similarly accumulation of a thick section by sedimentation in subsiding areas prevents development of horizontal thrust faults near its bottom by compressional deformation. Conversely removal of rocks by erosion in uplifting areas favors compressional horizontal thrusting (England and Molnar, 1990; Dahlen, 1990; Sanders, 1998). There are other cases in which uplift can generate large slopes and normal fault sliding. Extensional collapse deformation of mountain belts can be a consequence of gravitational instability generated by uplift (i.e. Dewey, 1988). This represents another example of deformation induced by vertical movements of the earth's surface.

In addition, sedimentary processes are responsible for stratigraphic piles with layered variations in strength of different kinds of sedimentary rocks, so that some sedimentary layers are weaker than others. This favors thrust detachment at specific stratigraphic positions, sliding planes along specific beds. This is an important factor in the geometry of thrust belts (Boyer and Elliott, 1982; Mandl, 1988).

CONCLUSIONS

- At the lithosphere scale (1) thermal phenomena, (2) vertical movements of the earth surface responsible for sedimentation and erosion and (3) tectonic deformation are not independent phenomena, they are intimately related by physical laws than can be studied in a quantitative basis (Figures 13 and 14). The mutual interdependence of sedimentation/erosion, deformation and thermal processes make the isolated study of traditional fields of earth sciences i.e. stratigraphy, geomorphology and structural geology obsolete. New approaches in earth sciences are evolving to integrate these traditional fields into a more general earth science looking at the earth as a single system.

- The natural processes that produce petroleum accumulations in a sedimentary basin require several ingredients: (1) the petroleum system elements: source, reservoir, seal, and overburden rocks, which are the result of sedimentation processes in a subsiding basin; (2) petroleum traps, which in many cases are the result of deformation processes of the sedimentary fill of the basin and (3) heat to convert suitable organic matter into petroleum (Figure 1). Although these three different phenomena are considered independent at the scale on an oil field, their mutual quantitative inter-relationships (Figures 13 and 14) may allow predicting one factor having knowledge of the others. For example the sedimentary record of a sedimentary basin can be used to better understand its deformational and thermal –thermo-tectonic- history. Conversely the thermal history recorded in rocks can be used to reconstruct the past uplifting and erosion events. Applications of these concepts can contribute to understand complex hydrocarbon exploration areas as the Colombian sedimentary basins (i.e. Sarmiento-Rojas, 2001) and reduce petroleum exploration risk.

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