A PLASTIC RHEOLOGY PHENOMENOLOGICAL MODEL THAT EXPLAINS THE ANDES EVOLUTION IN NORTHERN ARGENTINA

Antonio INTROCASO, Mario GIMÉNEZ, M. Patricia MARTÍNEZ & Francisco RUIZ

Boletín del Instituto de Fisiografía y Geología

Introcaso A., Giménez M., Martínez M.P. & Ruiz F., 2012. A plastic rheology phenomenological model that explains the Andes evolution in northern Argentina. *Boletín del Instituto de Fisiografía y Geología* **82**: 1-8. Rosario, 28-12-2012. ISSN 1666-115X.

Abstract.- A plastic rheology, partially phenomenological model is presented to explain the isostatically compensated Andean relief formation. This model considers a combination of lithospheric heating with long period relaxation and successive crustal shortenings on a north section of Argentina located at 24°S latitude. The present size of the Andean root –related to the Andes construction– was obtained by inverting regionalized Bouguer anomalies, also consistent with geoid undulations. The relief of 1520 km^2 is balanced both with a thermal root of 70.63 km located at the lithospheric mantle and with an evolutionary crustal root originated by shortenings. A pretended contradiction must be noted: lateral compressive shortenings and side moving tending to squash and to widen the high relief. For better understanding the problem, a linear model is first presented. Then, a nonlinear model of plastic rheology is considered with a stability coefficient depending on width and elevation. We have assumed that isostatic balance keeps on the whole orogenic construction. We also impose an empirical stability relationship. If this relationship does not hold, there will be an extra relief and therefore an extra root that will plastically flow until reaching stability.

Key-words: Andean relief; crustal shortening; Andean root; bouguer anomalies; isostatic balance; northern Argentina.

Resumen.- ***Modelo fenomenológico de reología plástica que explica la evolución de los Andes en el norte de Argentina.*** En este trabajo se presenta un modelo parcialmente fenomenológico de reología plástica para explicar la formación del relieve andino compensado isostáticamente. Este modelo considera una combinación de calentamiento litosférico con período largo de relajación y sucesivos acortamientos corticales en una sección del norte de Argentina, localizado en latitud 24°S. El tamaño actual de la raíz andina, relacionado con la construcción de los Andes, fué obtenido invirtiendo anomalías de Bouguer regionalizadas, consistentes con las ondulaciones del geoide. El relieve de 1520 km^2 es balanceado con una raíz térmica de 70.63 km localizada en el manto litosférico y con una corteza evolutiva originada por acortamientos. Debe ser notada una supuesta contradicción: acortamientos compresivos laterales y movimientos laterales tendiendo a aplanar y ensanchar el relieve alto. Para una mejor comprensión del problema se presenta primeramente un modelo lineal. Luego, es considerado un modelo plástico reológico con un coeficiente de estabilidad dependiendo del ancho y de la elevación. Asumimos aquí que se mantiene un balance isostático sobre el total de la construcción orogénica. Además imponemos una relación de estabilidad empírica. Si esta relación no se mantiene, habría un relieve extra y así una raíz extra que fluiría plásticamente hasta adquirir estabilidad.

Palabras clave: relieve andino; acortamiento cortical; raíz andina; anomalías de Bouguer; balance isostático; norte Argentino.

Adresses of the authors:

Antonio Introcaso [geofisic@fceia.unr.edu.ar]: *Facultad de Ciencias Exactas, Ingeniería y Agrimensura, Universidad Nacional de Rosario & Instituto de Física Rosario (CONICET), Pellegrini 250, 2000 Rosario, Argentina.*

Mario Giménez, M. Patricia Martínez, Francisco Ruiz: *Instituto Geofísico y Sismológico "Fernando Volponi", Universidad Nacional de San Juan, Argentina.*

Received: 06/10/2012; accepted: 09/11/2012

INTRODUCTION

Two main processes could explain the Andean built: subduction of the young oceanic crust under western South America, and the continuous advance of the South America Plate in the opposite direction (Fant et al. 1996). It is well known that the Nazca Plate makes frontal pressure on the western zone of the continental plate at 6.5 to 7 cm per year, according GPS observations (Barrientos 2010). The result is the built of the Andean belt. In order to analyze the original deformation it is necessary to know the geometry of the subduction contact, for example the oceanic plate dip and the convergence velocity.

Subducted oceanic plate dips are very changeable (Isacks & Barazangui, 1989). The weakest plates are located under the Andes by which subduction there causes significant tectonic compression producing shortenings of the South American margin. The compression stress σ_{xx} is about 1 kbar. During the orogenic building relief produces a constant growing overweight. The process stops at $\sigma_{xx} = \sigma_{zz}$. Nevertheless, vertical elevation forces could still act. They can be caused by thermal expansion or –as we assume here– by an initial excess (ascending growth) due to strong compression of weak materials.

A young plate with little dip and strong mechanical coupling has been classified as Chilean type. It presents great seismic activity and strong compressive deformation. On the contrary, a plate with a large dip and weak coupling has been classified as Mariana type.

It is well known that transitory elastic deformation dissipates in presence of significant earthquakes (Ramos et al. 2002), but the permanent deformations which we are concerned with, is the one caused by orogenic shortenings. In this case a minor elevation mechanism is also present: thermal expansion that makes the lithosphere ductile too (Introcaso 1997, Isacks 1988). In this paper we assume that crustal materials are deformed plastically from horizontal forces originated by the subduction contact.

It is not simple to explain this deformation because it involves a complex combination of different pushes. Different variations of structural style are tightly associated to the basement rheology and to the previous tectonic history (Allmendinger et al. 1983, Ramos et al. 1996, Key et al. 1999). Artyushcov (1983) showed that the continental crust shortening is easier if the bottom crust rests on a ductile deformable layer. Gratton (1989) has presented a dimensional analysis of the orogenic triangular growth from shortenings, isostatic compensation and plastic collapse. As we do in this work, he gave main importance to the crustal root, responsible for the relief built. Andes evolution requieres either relief (h) or width (a) increasing. Considering $\theta = a/h$ constant leads to a simple linear case not verifying distensive relief with lateral crust compression.

In this paper we propose a plastic rheololgical model explaining the paradox: slipping in high topography and, at the same time, lateral horizontal compression. Dewey & Lamb (1992) have analyzed the deformation of the whole Andean belt from tensors of the seismic moment. They have found parallel thrust at both sides of the Cordillera, and also detachment mainly on the western slope. An important conclusion they obtained is that shortening is originated by the force transfer from the subduction contact. Of course, this force is partially used for slipping part of the crust along the trench. Sobolev et al. (2006) have proposed a thermo-mechanical model, assuming: (1) a high thrust rate, (2) a thick crust, and (3) a relatively high coefficient of friction in the subduction channel.

Before going farther, we need to mention the brittle-ductile behaviour of the Andean tectonism. The brittle behaviour involves a very simple law. It is well known that the main part of the fault is entailed to an old fracture. This hypothesis is reasonable considering that the upper part of the lithosphere is fractured. Besides, the slipping on pre-existent planes has demanded less energy than the required for new fracture formation. Then the crust deformation concentrates over pre-existent discontinuities. The law for estimation of the quantity of shear necessary to produce fracture is nearly linear (Byerlee law). When the upper crust resistance is exceeded, usually brittle fractures are produced. But other alternative can be considered to explain that: materials behaviour is initially elastic and after some time the deformation is plastic, as observed in folds. Moreover at large depths, temperature increases and make materials soft (ductile behaviour).

In the continental crust the transition brittle to ductile is usually produced at depths of 15 or 20 km. Temperature there is 350 °C to 450 °C. So, if temperature should increase the transition would be located in the upper lithosphere (less resistance). In the Andean Cordillera 5000 m high (crust 70 km in thickness) width increases because yet enough energy exists to laterally accumulate material. Then, in Bolivian Andes, it reaches 800 km width. Crust there is thick and lithosphere is thin. A similar case can be observed under Tibet. Tension analysis usually assumes the Andes as a cylindrical belt (or a 2D-structure). Stress is originated by (a) topography produced by an overweight (σ_z) and by a light root (ascendant push, Arquímedes) and (b) border forces (σ_h). The deviator stress will be $\sigma_z - \sigma_h$. Sagging and extension (shipping) will take place if σ_h decreases or it is negligible.

We must point out that an opposition between floating forces flattening the crust (compressed vertically) and resistance of viscous forces is present in this orogenic process. Buoyancy effects can be described by a non-dimensional parameter: the Argand, Ar (England & McKenzie 1982). The Argand number (see also Gratton 1989) can be interpreted as a relationship between two processes: floating forces producing relaxation in vertical sense and plastic flow in horizontal sense. If t_c is the time of the first process and t_p the time of the second one, the Argand number is $Ar = \tau_c/\tau_p$.

If $t_p \sim t_c$ floating forces and plastic resistance are balanced.

If $t_p >> t_c$ lithosphere is rigid and relaxation process will be slow.

If $t_p << t_c$ lithosphere is soften and, in this case, relief is controlled by vertical forces. When t_P increases, also increases flatten.

England & McKenzie (1982) have pointed out that a low Argand number agrees with high viscosity, whereas a high one agrees with low viscosity (flatten, deformation). Argand number is relevant in geological processes (Willet 1999). We have assumed Ar = 1000 for our model. Vilotte et al. (1986) assumed values of 200, 500, 700 and 900 for a heterogeneous lithosphere.

The strong crustal shortening model we propose can justify the Andes of Argentina and Chile relief (Isacks 1988, Introcaso & Pacino 1988). With these ideas, 40 km thick crust and lithospheric mantle models were carried out using seismic and gravity data. We have found (a) a thermal root located in the low middle half of the lithosphere, which can justify 0.63 km of elevation, and (b) a crustal root 27 km thick which can justify 3.37 km of elevation (Introcaso et al. 1997). We assume that thermal relaxation time is very long compared with lithosphere plastic relaxation. In this case, viscosity mean value does not depend on temperature.

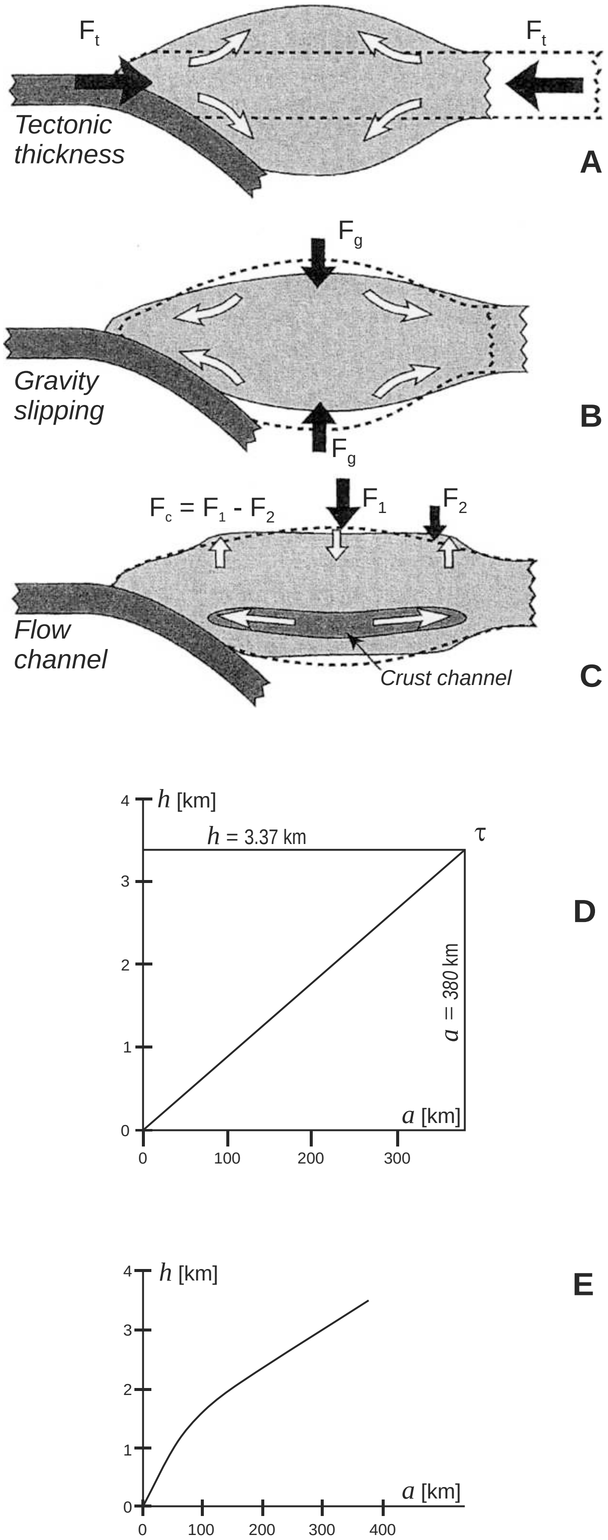

Figure 1. A: Tectonic forces produce crust shortening and thickening. **B:** Gravity and buoyancy forces slipping laterally the relief and the root. **C:** A channel of partial melting materials originates the whole relief flatten. Modified from Medvedev et al. (2006). **D:** width *a* versus relief elevation *h* for stability coefficient **θ** = 112.76 (constant). **E:** *a* versus *h* for **θ** increasing during evolution.

Table 1. Values for *a, h, θ'* (= *ah*) and S_h (= 0.20625 *θ'*) being ε (temporal evolution). *θ* constant throughout the whole evolution in a linear model.

ε	θ constant		θ	S_h
	a	*h*		
-	-	-	-	-
0.2	143.62	1.27	182.94	37.63
0.4	203.11	1.80	365.88	75.46
0.6	248.76	2.21	548.82	113.19
0.8	287.25	2.55	731.67	150.93
1.0	321.15	2.85	914.71	188.65
1.2	351.69	3.12	1097.65	226.39
1.4	380	3.37	1280.60	264.12

It is interesting to analyze different models considering different rheologies and weighs with very simple calculus variations (Vilotte et al. 1986). It is important to point out that Scandinavia rheology arises from a crustal elevation of 1 cm per year. This velocity corresponds to a viscosity about $4(10^{22})$ Pa and a relaxation time of 10 000 years (Vening & Meinesz 1937). Ranalli & Schloessin (1989) have proposed a non-linear intermittent episodic behaviour (creep) for heterogeneous materials in mantle deformation (change to plastic state).

We show in this paper the repeated process in a phenomenological way. We have assumed that in a time *t*, cycles are repeated in steps Δ*t*. Every cycle deformation velocities *ε* increases, while stress σ decreases. At the same time, deformation resistance (creep) increases. We consider there is a cinematic recover.

We propose Andean evolution in a cross section located at 24.5°S, involving: (1) a lithosphere (and crust) present model obtained from seismic and gravimetrical observations, with $\theta = 112.75$ (stability rate), (2) an evolution of width *a* and elevation *h* in time using plastic flux laws, with *a*, *h* and *θ* proportional to $t^{2/3}$, $t^{1/3}$ and $t^{1/3}$ respectively. As evolution of ductile deformation takes place from the crustal root, the isostatic balance in this model can be justified by the plastic elevation of the topography. It is possible that relief involve mobility on reactivated old fractures. Assuming plastic rheology the model explains shortening and relief balance evolution in time, for a stability θ_S verifying $\theta_S \geq A'' t^{1/3}$. If $\theta < \theta_S$ the high relief will collapse, laterally slipping, increasing *a* and decreasing *h*. The process stop when $\theta = \theta_S$. This model agrees with younger orogenies or plateaus like the Tibet (Gratton 1989, Molnar & Tapponier 1978).

Medvedev et al. (2006) model, partially shown slightly modified in Fig. 1A, is clearly a phenomenological model, whereas the model presented herein is mathematical. On the other hand, the model presented herein is different from other models, mainly because it is a crustal model, not lithospherical, and because it explains the apparent paradox: compression (in total crust)-tension (on Andean top) by means of mathematical plastic rheology.

In summary, we assume that changes in deep materials (see Ampferer 1906 among others), particularly orogenic roots plastic flattens (Gratton 1989), are responsible for relief deformations. Alternatively, a flow channel in deep crust (Medvedev et al. 2006) can justify Andean relief deformations. In Fig. 1A is shown the single orogenic process.

Table 2. ε: steps for the temporary evolution, h: column that indicates the growth of relief (Eq. 4), a: Andean wide growth, $\theta = a/h$ (Eq. 6). h, a and θ values correspond to instability except for the end $\varepsilon = 1.4$. This situation anticipates the collapse. The column θ' indicates invariant areas, as also S_h (Eq. 3). The two last columns h_S and a_S correspond to stability reached in the relief in each pulse, and reproduce values of Tab. 1 for comparison.

ε	Instability			Common parameters		Stability	
	h	a	θ	θ'	S_h	h_S	a_S
0	-	-	-	-	-	-	-
0.05	1.12	40.61	36.06	45.74	9.43	0.63	71.81
0.10	1.42	64.46	45.42	91.47	18.87	0.90	101.56
0.20	1.79	102.32	57.23	182.94	37.63	1.27	143.62
0.40	2.25	162.43	72.19	365.80	75.45	1.80	203.09
0.60	2.58	212.84	82.56	548.71	113.17	2.21	248.77
0.80	2.84	257.84	90.85	731.74	150.93	2.55	287.24
1.00	3.06	299.20	97.87	914.65	188.64	2.85	321.14
1.20	3.25	337.87	103.96	1097.58	226.29	3.12	351.80
1.40	3.42	374.44	109.48	1280.60	264.12	3.37	380.00

Horizontal tectonic forces shorten and thicken the Andean crust. In Fig. 1B it is shown the upper crust and the root slipping from gravity and buoyancy forces. As shown in Fig. 1C the channel with partially melt materials in middle and low crust is responsible for the relief flatten from the Miocene. The absolute value of F_c associated with the channel flow is the result of lithostatic loads at the bottom of the lower crust. So the low viscosity channel flow originated by thermal radioactivity produces differential load in upper crust. Channel effects are significant and if partial melting with viscosities of 10^{18} - 10^{20} Pa exist, an upper plateau appears. The effect is not significant (10^{22} Pa) outside the channel.

Considerations on θ and θ_1.- Stability coefficients $\theta = a/h$ and $\theta_1 = a/R$, in which a is width, h is relief elevation and R is crustal root, are associated with the relief and the root. Gratton (1989) has assumed a triangular model with shear stress in relief and root $\tau = 2g\sigma h$ and $\tau' = 2g\sigma R$, with g, σ, h, R, θ and θ_1 being gravity, density, relief high, crustal root and stability coefficients, respectively. So, the higher α and α' or the smaller θ and θ_1, the taller and more unstable the geological structure will be. Moreover τ and τ' can be dynamically expressed as functions of velocity or of orogenic time.

THE MODEL (IN 24°S LATITUDE)

From $\theta = a/h$ (stability coefficient) and $\theta' = ah$ (2D-relief), we have:

$$h = \sqrt{\theta'/\theta} \qquad \textbf{(Equation 1)}$$

$$a = \sqrt{\theta\theta'} \qquad \textbf{(Equation 2)}$$

Considering that $\theta' = 1520$ km^2 (Introcaso et al. 2000), that the plateau is 4 km high (Froideveaux & Isacks 1984), and that 0.63 km of the 4 km are produced by thermal elevation (Introcaso et al. 1997), let us consider the whole elevation (4 km): $a = 380$ km, then $\theta = 95$. The 2D-area corresponding to the thermal effects is (0.63 km)(380 km) = 239.4 km^2 by which the new value of θ' will be 1280.6 km^2 with $h = 3.37$ km. Thus, $\theta = 112.75$ km. Shortening building the relief will be

$$S_h = \frac{ah + 7.25ah}{T(=40)} = 0.20625\theta \text{ km} \qquad \textbf{(Equation 3)}$$

T is the thickness of an undeformed crust. Root R is $7.25h$ (Introcaso et al. 1997).

LINEAR MODEL

If the model to understand evolution is linear, although simple it would not agree with the compression-traction process observed at the Andes. If θ constant, the relationships a/h or h/a will keep invariant throughout the whole evolution (Fig. 1D and Tab. 1, in which the column ε points out the advance of the deformation in time and anticipates what is shown Tab. 2 and Figs. 2-3). Eight values from 0 to 1.4 were considered separated by steps of 0.2; or seven equally distributed in time (1.4 per 1000). There are obtained steps of 189.94 km^2. Then with $\theta = 112.75$ (constant) and θ' (2D-relief, successively increasing), and from Eqs. 1-3 we obtain h, a and S_h for every step.

As we have pointed out, the model involves lateral compression that shortens part of the crust. This shortening originates pulses that make the relief grow up to stability. Thus, the plastic model explains the slipping of the relief.

An Andean model from the gravity inversion at 24°S has already been presented by Introcaso et al. (2000: 25, fig. 8). This model confirms the isostatic Airy behaviour: $Rh = 6.675$, being R the root, h the elevation.

PLASTIC MODEL

It is well known that the oceanic plate thickness is proportional to $t^{1/2}$ (Le Pichon et al. 1973). Orogenic building also involves different times: relief building, for example $t_0 = 10$ at 15000 Ma; isostatic balance t_p, 1000 times smaller, for example 10000 to 15000. Gratton (1989) found $\theta = k(t_0/t_p)^{1/2}$, with t_0/t_p about 1.46(10^3) times.

Vilotte et al. (1987) have indicated Argand numbers of 300, 500 and 900 for heterogeneous lithosphere. The relationship t_0/t_p can be considered as the number of times the isostatic balance takes place in the orogenic time t_0. For example $\theta = k(t_0/t_p)^{1/3}$ with k = 10 will lead to $\theta = 10(1000)^{1/3} \approx 100\,\theta$, a and h depend on $t^{1/n}$. For θ and h we have assumed $n = 3$ and for a, $n = 2/3$ with t increasing from 0.2 in 0.2(10^3) up to 1.4(10^3). As Molnar & Tapponier (1978) have pointed out, assuming a static model, at the end of a structure evolution the principal stresses are $\sigma_{xx} = \sigma_{zz}$. Relief building stops but material accumulates in both lateral borders.

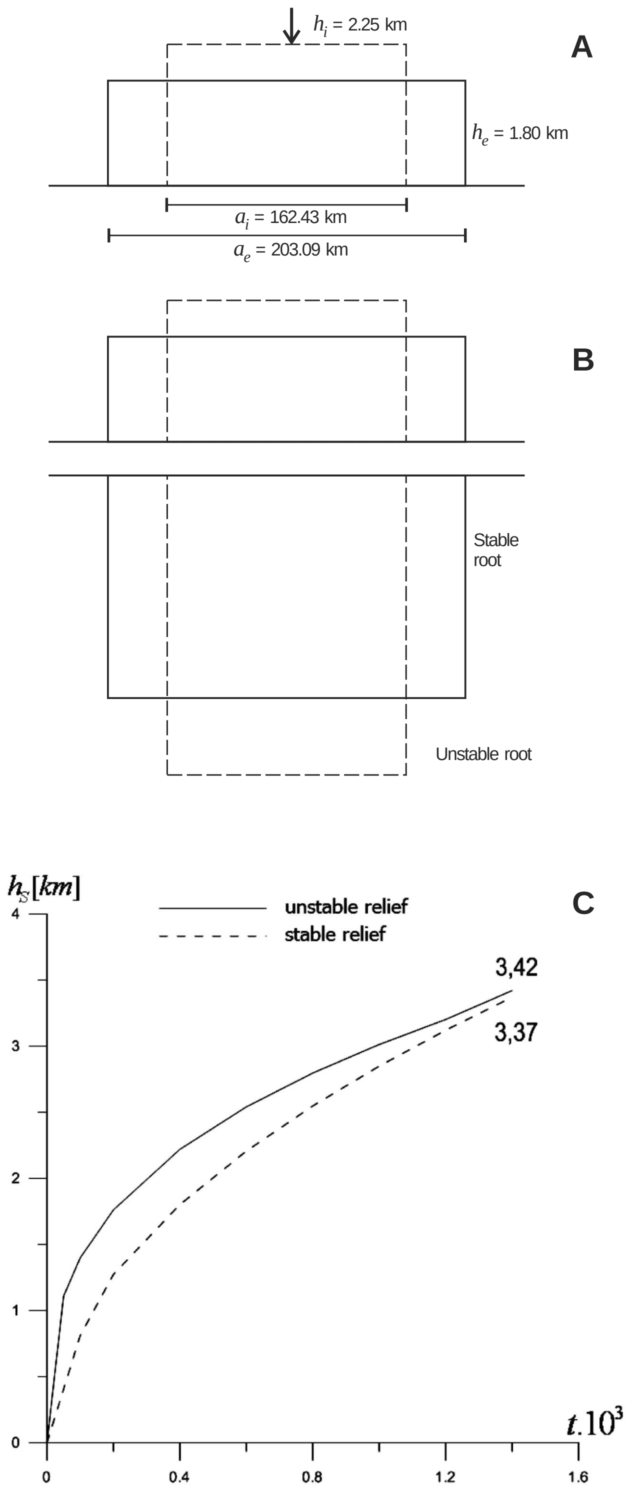

Figure 2. A: Plastic pulse in 0.4 shortens the crust (unstable width a) and increases h (unstable) as shown by dashed lines rectangle. There is instability θ_i (unstable) < θ_s (stable). Then the rectangle of invariant area is flattened. Relief decreases and width increases until reaching stability (θ = 112.76). **B:** From Archimedes forces, the unstable root is compressed vertically and it widens. The invariant area of the root will be $\theta = a_i R_i$ = 2652.05 km, and the stability root $Rs = 7.25 h_S$ = 13.05 km. **C:** Non-linear variation of the relief (stable) for the pulses from stability values (Tab. 2, column 2). Solid line for unstable relief (θ < 112.76), dashed line for stable relief (θ = 112.76).

Table 3. Thicknesses of unstable and stable roots for the pulses indicated in Tab. 2.

ε	Unstable R_i	Stable R_s
0.00		
0.05	8.12	4.56
0.10	10.30	6.53
0.20	12.98	9.21
0.40	16.31	13.05
0.60	18.70	16.02
0.80	20.60	18.48
1.00	22.18	20.66
1.20	23.56	22.62
1.40	24.80	24.43

We will now see the expressions used here, involving θ, a and h for the adopted plastic model. Considering Eqs. 27 and 28 in Gratton (1989), we have fitted the terms to obtain a = 380 km and h = 3.37 km. We will assume that shortenings make the relief increase 50 m additionally, *i.e.* the maxima reached height will be 3.42 km. Thus we have deduced A in Eq. 4. As the 2D-relief area (ah = 1280.60 km^2) remains unchanged, the corresponding width will be 374.44 km. Thus we have obtained A' in Eq. 5 and finally A'' in Eq. 6 from Eqs. 4 and 5:

$$h = A\tau^{\frac{1}{3}} \qquad \textbf{(Equation 4)}$$

with A = 0.305715 km and $\tau = \varepsilon(10^3)$, ε = 0.2, 0.4, ..., 1.4, and

$$a = A'.\tau^{\frac{2}{3}} \qquad \textbf{(Equation 5)}$$

with A' = 2.9920149 km and $\tau = \varepsilon(10^3)$, ε = 0.2, 0.4, ..., 1.4, and

$$\theta = A''.\tau^{\frac{1}{3}} \qquad \textbf{(Equation 6)}$$

with A'' = 9.7869418 and $\tau = \varepsilon(10^3)$, ε = 0.2, 0.4, ..., 1.4. Coefficients A, A', and A" have been calculated as $A = (3.42/1400^{1/3})$ km, $A' = (374.44/1400^{1/6})$ km, $A'' = (3.42/1400^{1/3})$ km.

While τ, θ and A'' are dimensionless, A, A' and Sh are expressed in km. The meaning of A, A' and A'', A is the rate of unstable relief h plastic growth rate in time. If A increases, h increases inversely. A' is width A plastic growth rate in time. Finally A'' is the instability coefficient related to the plastic process. Gratton (1989) pointed out that A, A' and A'' depend on density, gravity and shear stress in relief and root, to the powers 1/3, 1/3 and 2/3 respectively. Our model holds A and A' constant during the whole process. Tab. 2 shows the steps ε for temporal evolution and the values of θ, a, h, θ' and Sh for the unstable growth pulses, which became stable.

To clarify the process we can take values from Tab. 2 for ε = 0.4. Fig. 2 shows this case. Numerical values corresponding to the unstable plastic process, from Eqs. 4-6 for ε = 0.4, are $h = 0.305715(400)^{1/3}$ = 2.25 km, $a = 2.992015(400)^{2/3}$ = 162.43 km, θ = 72.2. Note that the 2D-surface of the relief (θ'

$= a_i h_i = 365.8$ km^2) is invariant. In the stability process, the value $a/h = 112.76$ is assumed as stable. From Eqs. 1 and 2 the new values will be $h = (\theta'/\theta)^{1/2} = 1.80$ km, $a = (\theta'\theta)^{1/2} = 203.09$ km.

Let us consider the root from which orogeny is handled. In the unstable course, the width a_i and the thickness by the unstable root R_i will be $a_i = 162.43$ km. And $R_i = 7.25(2.25)$ km = 16.3125 km. Then, for the pulses indicated in Tab. 2 it can be obtained the thicknesses of unstable and stable roots indicated in Tab. 3. Plastic slidings (relief and width increasing) leading to stability are expressed by:

$$-\Delta h = -At^{1/3} + \sqrt{\frac{\theta'}{\theta}} \qquad \textbf{(Equation 7)}$$

$$\Delta a = -A't^{2/3} + \sqrt{\theta\theta'} \qquad \textbf{(Equation 8)}$$

$$\Delta\theta = 112.76 - A''t^{1/3} \qquad \textbf{(Equation 9)}$$

In summary, increasing of width a and high h of the relief takes place by successive pulses of plastic deformation (Eqs. 3-4, 6). When θ is assumed lower than the stability limit ($\theta_S = 112.75$), the process is unstable. Soon, collapses will follow one another (h decreases up to h_S and a decreases up to a_S) in order to reach the stability. h_S and a_S were obtained from the Eqs. 1 and 2 using $\theta = \theta_S$ and θ' in the fifth column of Tab. 2.

Willet (1999) has pointed out that it is frequent to observe simultaneous compression and extension. Since the flow time in orogenic times is instantaneous, we will assume that the Andean growth (today 0.05 km in excess) and the collapses that follow are produced at the same time.

Fig. 2C shows the nonlinear variation of the relief (stable) for the pulses of Tab. 2 from instability values (Tab. 2, column 2). Fig. 3, with values of Tab. 2, explains the crustal shortening process. Final values must be increased in 0.63 (thermal expansion). On the right it can be seen the deformation front that, according to Ramos (2001), nearly agrees with the inner boundary of the Wadatti-Benioff zone. If $\theta = Bt^{1/2}$ type of rheology would be adopted, a and h could be $a = B't^{2/3}$ and $h = B''t^{1/6}$. Then $\theta = (B'/B'')t^{1/2} = 112.75$ with coefficients $B' = 3$ km, $B'' = 0.1$ km and $B = 33$ (dimensionless).

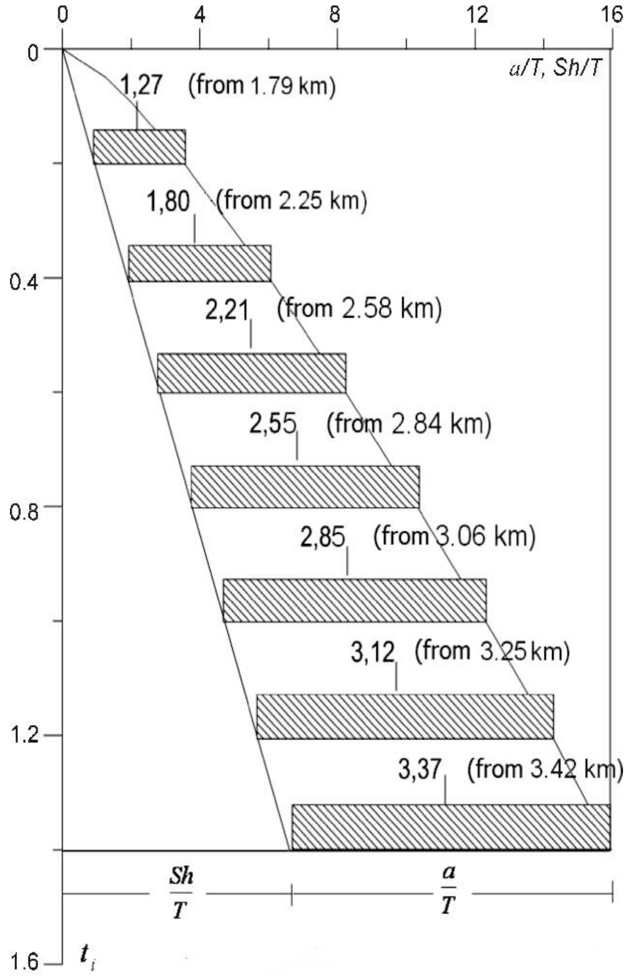

Figure 3. Evolution of the Andes in a 24°S cross section, according to the proposed model of plastic rheology. Only the relief originated by shortenings is shown. Vertical axes is ε (10^{-1}) with ε from 0 to 1.4 in steps of 0.2. Horizontal axes is shortenings S_h/T and width a/T. Assuming that thermal expansion holds during evolution, it would add 0.63 km to the relief (see Tab. 2). The deformation front increases with time towards the foreland. See Fig. 1B for a tentative model that explains the Andean built.

DISCUSSION AND CONCLUSION

The explanation of the formation of the Andes requieres the consideration of strong horizontal compressive forces, which indeed exist. They are of the order of up to 1 kbar in the Altiplano Plateau (Froideveaux & Isacks 1984). These forces are originated in the subduction contact and produce the shortenings in western South America.

It is well known that, at a certain depth, for example at the crust bottom, temperature increases and the materials are ductile. For the middle and lower crust a channelled flow of partially molten materials has been proposed (Sobolev et al. 2006 among others). This proposal is plausible because under the Andes of northern Argentina, there are recorded high anomalous lithospheric temperatures (Froideveaux & Isacks 1984, Introcaso & Pacino 1988, Isacks 1988, Introcaso et al. 1997). This recognition is important because we assume that plastic deformation of the root drives the relief. Thus, it is neither possible to consider a Newtonian rheology because the required melting temperatures are not observed, nor a rigid rheology is convincing because it could not explain the compressional-tensional stresses found step by step.

Our model adopts a plastic rheology with a deformation effect that explains the unstable relief. The deformation effect is carried out by (a) a stability coefficient of $\theta \approx \text{Ø}t^{1/3}$, (b) a width $a \approx \text{Ø}t^{1/3}$ and (c) an altitude $h \approx \text{Ø}t^{1/3}$ with $t = 1000$, that expresses the dimensionless rate of orogenic time/plastic flow time. When recognizing the long Andean heating in northern Argentina, we have adopted a soft lithospere to explain the formation of the mountain range. Agreeing with the shortenings, the instantaneous isostatic compensation in orogenic scale takes place. In addition, a slenderness coefficient must be imposed in order to guarantee stability. They contribute to the deformation progress, the temperature increasing and the growth of crustal root (Wdowinsky & Boock 1994). While in the beginning the high temperatures reduces viscosity (increasing the Argand number) and hold it, the crustal thickens replaces a resistant upper mantle by weak crust.

Having defined *a, h* and θ for different time intervals, we confirm that θ_i keeps under θ_S = 112.75, *i.e.* in an unstable relief zone. Auto regulation takes place, entailing relief decreasing and width increasing reaching stability (θ = 112.75). In the deformation process by sagging, both shortening and areas (relief and root) keep invariant in each pulse. Finally values must be increased in 0.63 km (thermal expansion).

The intrinsic characteristics of parameters such as A, A' and A'' and shear stresses remain to be shown.

Acknowledgements.- The authors wish to thank G. González Bonorino (Centro Austral de Investigaciones Científicas, Ushuaia) and a further anonymous reviewer for their valuable comments as reviewers of the journal. This work has been granted by I ING 261 (Universidad Nacional de Rosario, 2008-2012).

REFERENCES

Allmendinger R., Ramos V., Jordan T., Palma T.E. & Isacks B., 1983. Paleogeography and Andean structural geometry, northwest Argentina. *Tectonics* **2**: 1-16.

Ampferer O., 1906. Über das Bewegungs Bild von Faltengebirge. *Jahrbuch Geologisches R. A. Viena* **56**: 265-272.

Argand E., 1924. La tectonique de l'Asie. *Proceedings of the 13rd International Geological Congress* **1**: 170-372.

Artyushcov E.V., 1973. Streses in the litosphere caused by crustal thickness in homogeneties. *Journal of Geophysical Research* **78**: 7675-7708.

Artyushcov E.V., 1993. Geodynamics. Elsevier, 312 p.

Barrientos S., 2010. Terremoto (M = 8.8) del 27 de febrero de 2010 en Chile. *Revista de la Asociación Geológica Argentina* **67**: 412-420.

Dewey J.E & Lamb S., 1992. Active tectonics in the Andes. *Tectonophysics* **205**: 79-95.

England P.C & McKenzie D.P., 1982. A thin viscous sheet model for continual deformation. Geophysical Journal Reserch (Astronomical Section) **70**: 294-321.

Fan G., Wallace T.C., Beck S. & Chase C., 1996. Gravity anomalies and flexural model: constraints on the structure beneath the Perunian Andes. *Tectonophysics* **255**: 99-109.

Gratton J., 1989. Crustal shortening root spreading, isostasy, and the growth of orogening belts: adimensional analysis. *Journal of Geophysical Reserch* **94**: 15627-15634.

Introcaso A., 1994. Gravimetría. Sus aplicaciones tectónicas. Instituto de Física Rosario, 210 p.

Introcaso A. & Pacino M.C., 1988. Gravity andean model associated with subduction near 24º25' south latitude. *Revista de Geofisica* **44**: 24-47.

Introcaso A., Pacino M.C. & Guspi F., 1997. A gravity chart to quickly evaluate the probable contribution of lithospheric heating to isostatic balance in the central Andes. *Bollettino di Geofisica Teorica ed Applicata* **38**: 41-54.

Introcaso A., Pacino M.C. & Guspi F., 2000. The Andean of Argentina and Chile: crustal configuration, isostacy, shortening and tectonic features from gravity data. *Temas de Geociencia* **5**: 1-31.

Isacks B. & Barazangui M., 1981. Geometry of Benioff zones, lateral segmentation and downord bending of the subduction lithosphere. Maurice Ewing Series 1, **AGU 99**: 1-114.

Isacks B., 1988. Uplift of the Central Andes Plateau and bending of the Bolivian Orocline. *Journal of Geophysical Reserch* **93**: 3211-3231.

Kley J., Monaldi C. & Salfity J., 1999. A long-Strike segmentation of the andean foreland cauces and consecuences. *Tectonophysics* **301**: 75-94.

Le Pichon X., Francheteau J. & Bonin J., 1993. Plate tectonics. Elsevier, 300 p.

Medvedev S. Podladchicov Y., Haudy M. & Sheuber E., 2006. Control on the deformation of the central and souther Andes (10º - 35º S). *In*: O. Oncken, G. Chong,, G. Franz, P. Giese, P., H.-J. Götze, V.A. Ramos, M.R. Strecker & P. Wigger (*Eds*.): The Andes. Springer, p. 474-494.

Molnar P. & Tapponier P., 1978. Active tectonism of Tibet. *Memories d' Ostram* **86**: 1-310.

Ramos V., Cegarra M. & Cristallini E., 1996. Cenozoic tectonism of the hight Andes of west-central Argentina (30º-36º latitude). *Tectonophysics* **259**: 185-200.

Ramos V., Cristallini E. & Perez D., 2002. The pampean flat-slab of the central Andes. *Journal of South American Earth Sciences* **15**: 59-78.

Ranalli G. & Schloessin H., 1989. Role of episodic creep in gobal mantle deformation. *Geophysical Monograph* **49**: 55-63.

Sobolev S., Babeyko A., Kaulacov I. & Oncaen O., 2006. Mechanism of the Andean orogenic. *In*: O. Oncken, G. Chong,, G. Franz, P. Giese, P., H.-J. Götze, V.A. Ramos, M.R. Strecker & P. Wigger (*Eds*.): The Andes. Springer, p. 513-535.

Suarez G.P., Molnar P. & Burchfield B.C., 1983. Sismicity, fault plane solutions, depth of faulting an active tectonics of the Andes of Peru, Ecuador and Souther Colombia. *Journal Geophysical Research* **88**: 10403-10428.

Vening-Meinesz F.A., 1937. The determination of the Earth's plasticity from the post glacial uplift of Scandinavia: Isostatic adjustement. *Proceedings of the Koninklijke Nederlandse Akademie van Wetenschappen.* **B40**: 654.

Vilotte J.P., Madariaga R., Daignieres M. & Zienkiewicz O., 1986. Numerical study of continental collision: influence of buoyancy forces and initial stiff inclusion. *Geophysical Journal of the Royal Astronomical Society* **84**: 210-226.

Wdowinsky S. & Book Y., 1994. The evolution of deformation and topography of high elevated plateau. Application of central Andes. *Journal of Geophysical Research* **99**: 12245-12256.

Willet S. D., 1999. Rheological dependence of extension in wedge models of convergent orogens. *Tectonophysics* **305**: 419-435.