MAGYAR ÁLLAMI
EÖTVÖS LORÁND
GEOFIZIKAI INTÉZET

# GEOFIZIKAI
KÖZLEMÉNYEK

ВЕНГЕРСКИЙ
ГЕОФИЗИЧЕСКИЙ
ИНСТИТУТ
ИМ Л. ЭТВЕША

# ГЕОФИЗИЧЕСКИЙ
БЮЛЛЕТЕНЬ

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CONTENTS



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BUDAPEST

## TARTALOMJEGYZÉK



## СОДЕРЖАНИЕ

MAEGS 8

EVOLUTION OF INTRAMONTANE BASINS

# 8th Meeting of the Association of European Geological Societies

Topic:
EVOLUTION
OF INTRAMONTANE BASINS
ON THE EXAMPLE OF THE PANNONIAN BASIN WITH PARTICULAR EMPHASIS ON SEQUENCE STRATIGRAPHY AND NEOTECTONICS

Schedule:

| | | |
|---|---|---|
| September | 18 (Sat) | |
| | 19-20 (Sun-Mon) | (A) Pre-Meeting Field Trip: Marginal Facies of the Pannonian Basin (including lignite deposits) |
| | 21 (Tue) | Registration and Sightseeing in Budapest, with a visit to the Hungarian Geological Survey (founded in 1869) |
| | 22-24 (Wed-Fri) | Technical Sessions (including a Poster Session) |
| | 25-26 (Sat-Sun) | Post-Meeting Field Trips: |
| | | (B) Geology, agriculture, environment and urban engineering. Geology in the Pannonian Basin. |
| | | (C) Oil and gas, underground water, and geothermy in the Pannonian Basin. |

Language: English

During the Meeting,
an All-European Regional Meeting of the International Geological Correlation Programme (IGCP)
will also be held

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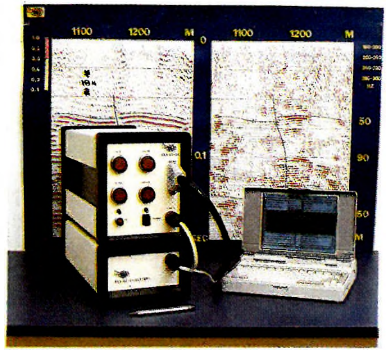

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# FORMATION OF THE CRUST-MANTLE BOUNDARY IN THE PREVIOUS UPPER MANTLE

Károly POSGAY*

Structural units extending over the present crust-mantle boundary are to be seen on the Pannonian Geotraverse deep seismic section [POSGAY et al. 1990]. Studying these units one can conclude that the present crust-mantle boundary lies in the previous upper mantle. During the depth-ward drift of this boundary the fabric of the depth range of the crust-mantle boundary was not destroyed thus it can be studied on the seismic section.

It can be assumed that the above appearance can be linked with the formation of the Pannonian Basin. Due to tensile stress in the lithosphere, partial melting resulted from the intrusions penetrating into the crust-mantle boundary and from the elevated asthenosphere. The melted material settled down in nearly horizontal bodies. The position of the new crust-mantle boundary is assumed on the basis of the lowest reflection having outstanding amplitude obtained from these bodies.

The repetition of the described process can be supposed on the basis of the boundaries below each other.

**Keywords: Mohorovičić discontinuity, basin development, Pannonian Geotraverse, upper mantle reflections**

## 1. Introduction

Results of seismic measurements along the Pannonian Geotraverse [POSGAY et al. 1991a] show that the crust-mantle boundary moved towards the depth during the development of the lithosphere. Certain parts of the section allow us to conclude that the present crust-mantle boundary lies in a domain that previously belonged to the upper mantle.

The novelty of the assumption accounts for studying whether it is in accordance with the facts and theories described in the literature. There are

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references [MATTHEWS and the BIRPS Group 1990, LIE et al. 1990] that mention those three or four places where a continuously reflecting layer from the upper mantle has been able to be determined. Apart from the Pannonian Basin, all the other determinations have been obtained by offshore measurements [BREWER et al. 1983, CLOWES et al. 1987, LIE et al. 1990]. It is likely that the good penetration results are due to utilizing 2-10 Hz frequency domain [POSGAY et al. 1990]. The low frequency experiments [DOHR, FUCHS 1967] ceased probably because at the time of the 1961-64 measurements there was not yet adequate computer processing facilities for enhancing reflections and suppressing surface waves. At the time of computerized geophysical measurements, however, low frequency measurements lost their importance as vibroseis measurements are easier to carry out in densely populated or industrialized areas.

## 2. Interpretation of reflecting horizons obtained from the vicinity of the crust-mantle boundary

The enclosure shows the 110 km long part of the seismic profile measured along the Pannonian Geotraverse. After true-amplitude recovery the strongest amplitudes were marked with violescent colours, then red-pink-yellow-green-blue colours were used to differentiate the amplitudes. Light blue marked the smallest amplitudes [POSGAY et al. 1991a]. The continuous black lines indicate the assumed deep fault structure zones, upthrusting and piling up fractures in the depth domain of the crust-mantle boundary, the bottom of the Neogene basement and the crust-mantle boundary in accordance with the assumption that constitutes the basis of the interpretation.

In general the crust-mantle boundary coincides with the deepest lower crustal reflecting horizons over the domain of the upper mantle that seems to be transparent when the usual frequency band is involved [MOONEY, BRAILE 1989]. If low frequencies are applied, reflecting layers can be determined in the upper mantle as well. These layers usually appear as weaker amplitudes than those coming from the crust. In that case the deepest reflections having relatively stronger amplitude with short interruptions can be marked as the crust-mantle boundary. Such reflection series often have a characteristic difference from that of reflections observed in its surroundings [POSGAY et al. 1981]; the horizon of these series can more precisely be followed: its amplitude is higher, its dip is smaller it 'strikes through' the background reflection picture. The overall impression is the following: a previous fabric can be concluded from the background, the traces of which have already been dimmed by now, and the assumably present crust-mantle boundary is a younger formation than the 'background'.

The crust-mantle boundary on the Pannonian Geotraverse seismic section has been marked on the basis of the above mentioned principles, and this boundary correlates well with the boundary having 8.1 km/s velocity previously determined mainly by refraction and wide angle reflection measurements [POSGAY et al. 1991b].

The following conclusion can be drawn from the picture appearing in the depth domain of the crust-mantle boundary in the vicinity of upthrusts [POSGAY et al. 1991a, POSGAY, SZENTGYÖRGYI 1991]: both the crust-mantle boundary before the upthrusts and the boundaries having been formed since then give reflections. The highly reflecting layer at the 84 profile km around 8.4 s rising from south can be interpreted as an upthrusted, previous crust-mantle boundary. Under the assumed thrust plane to the left from the 8.8 s of 84 profile km there is a layer which is similarly highly reflecting; this layer was probably the continuation of the upper layer before the overthrust. The nearly horizontal reflection series starting to the right, on the southern side of the upthrust plane may indicate the present crust-mantle boundary. The reflection picture is very similar to those in the vicinity of the assumed thrust plane south of the 8.8 s region of the 99 profile km.

Assuming that the wave picture between 8 and 8.5 s of the 66 profile km reflects a structure similar to that which more clearly can be seen around 8.4 s of 84 profile km, a faster movement of the crust-mantle boundary towards the depth along the northern side of the strike-slip zone can be concluded. The horizontal reflection appearing markedly at 8.7-8.8 s of 59 profile km may be the residue of the crust-mantle boundary after upthrusting. The 'present boundary' can be concluded at 9.5-9.6 s and a just forming surface at 10.3-10.4 s. This conclusion is backed by the fact that the reflection series at 9.5-9.6 s can better be followed and has stronger amplitudes.

This reflection series intersects reflecting horizons that have seismic characteristics similar to those horizons — located beneath the presumed recent crust-mantle boundary — south of the main wrench fault zone, between 90 and 100 profile kms at 9.8-10.3 seconds.

According to the outlined assumption the recent crust-mantle boundary can be found both to the north and south of the main wrench fault zone, in a depth range that previously belonged to the upper mantle. It follows from this conception that in the northern region the boundary crosses such horizons that presumably are below the present boundary, south of the major structural line. Consequently the boundary moved downward on both sides of the main fault zone although its subsidence might have been stronger on the north side of the zone. Fractured structure and tectonic stress seem to be more increased in the northern part. Possibly a relationship may be sought between the considerably fractured feature and the faster downward shifting of the crust-mantle boundary.

## 3. Formation and drift of the crust–mantle boundary

The crust of the Earth is a density differentiation product of the mantle. This process is still going on even today and the crust displays a thickening tendency [MEISSNER 1986]. Detectable reflecting lower boundary of the young crust coming into being on oceanic ridges can be formed as a result of intrusions penetrating into the base of the crust from the mantle during a period of about 100 000 years [BARTON 1986], and its thickening trend is similar to those of older crust segments. On the basis of data collected about the crusts of the Earth and the Moon, Meissner inferred that the crust's thickness can be evaluated by the formula:

$$z = 25 \log t - 32$$

($z$ is given in km). As tectonic events disturb greatly the crust structure, $t$ (given in million years) represents the tectono-thermal age of the crust. It must be reckoned from the last tectonic event resulting in metamorphosis in almost the whole depth range of the crust [MEISSNER et al. 1987]. The formula given above is not valid for intensive tectonic belts: regions of mountain systems and rifts. During their formations 'anomalous' crustal thickening beneath the mountains and crustal thinning in rift zones and also in intramontane basins occurred. In both cases significant time (presumably of the order of $10^7$–$10^8$ my.) is necessary for the thickness anomaly to regress [MEISSNER et al. 1987, MCKENZIE 1978] and to get 'into accordance' with its tectono-thermal age. The above formula shows the tendency of this process. Conditions of the crustal formation cause the standard deviation of data [MOONEY, BRAILE 1989].

One interpretation of the thickening tendency — by MEISSNER [1986] — suggests that more material gets into the crust from the mantle (e.g. as intrusions) than from the crust into the mantle (e.g. by means of subduction).

Examples supporting the variability of the crust-mantle boundary even by seismic profiles can be found in the literature. New crust-mantle boundaries can take form through complex crust-mantle structures [POSGAY et al. 1981, BROWN et al. 1986, KLEMPERER et al. 1986, BARTON 1986, HALL et al. 1990]. On published sections there are dipping reflecting horizons through which the reflections of the crust-mantle boundary pass nearly horizontally. Thus a similar characteristic visible on the Pannonian Geotraverse seismic profile cannot be regarded as a particular and unique occurrence. On the seismic section in the upper and lower vicinity of the boundary, observable parts — from which piling up, upthrust, strike-slip and a more ductile transition zone in the lower lithosphere can be concluded according to the above assumption — should be supported by supplementary information making the novel attempt in interpretation reasonable.

Velocity data from the depth range of the crust-mantle boundary may offer a basis for interpreting this region's formation. Along the shores of Norway on the bottom of the continental crust ZEHNDER et al. [1990], during their reflection and refraction studies, specified a 'layer' of 0-3 km thickness and 7.5 km/s velocity which 'could represent a region of melt liberation and underplating that accompanied the Jurassic phase of extension'. SMITHSON [1989] considering a layer approximately 3 km thick having a seismic velocity of 7.4-7.8 km/s located on the bottom of the crust, in the area of Basin and Range, assumes that it 'could consist of a residuum from partial melting or a cumulate zone from crustal underplating' while with regard to the crust-mantle boundary 'the Moho zone is strongly layered and discontinuous because of intershearing of mantle and crustal rocks or because of intrusion of basaltic melt from the mantle along the mantle-crust transition'.

Velocity analysis in the Pannonian Basin area suggests a 5-7 km thick sequence on the bottom of the crust, characterized by 7.2-7.7 km/s seismic velocity [POSGAY 1975, POSGAY et al. 1980, 1981].

Relying on the results of the Pannonian Geotraverse PGT-1 seismic profile an interpretation can be given having the following new feature: the moments associated with the formation of the Pannonian Basin [STEGENA 1967, HORVÁTH et al. 1988, ROYDEN, DÖVÉNYI 1988] are assumed to have played a great part in the development of the crust's lower sequence with 7.2-7.7 km/s seismic velocity. During the stretching of the lithosphere in the initial phase — that can be characterized by block fractures, subsidence and uplifting of the asthenosphere [MCKENZIE 1978] — basaltic intrusions penetrating into the lower crust [GRIFFIN, O'REILLY 1986] increased the velocity of this sequence. The pushing up of more acid melt parts into upper region of the crust might have contributed to the velocity increase [FOUNTAIN 1986].

The plasticity difference between the lower crust and the uppermost domain of the mantle could have been reduced as a result of the heating effect of intrusions penetrating into the crustal bottom [FOUNTAIN 1989]. This process probably promoted the intrusions intruding into the mentioned uppermost mantle region also. Assuming basaltic intrusions [DAWSON 1987] their penetration might have decreased the velocity of the depth range in question thus forming a 'layer' of the upper mantle the velocity of which can be regarded as a transition between the crust and the mantle (e.g. 7.7 km/s on the area of the Pannonian Basin, 7.4-8.1 km/s in the Basin and Range province; [SMITHSON 1989, MOONEY, BRAILE 1989]).

It can be assumed that the observable piling up structures at the depth range of the crust-mantle boundary occurred in the course of Eocene-Oligocene strike-slip movement [POSGAY, SZENTGYÖRGYI 1991], viz. they are older than the extension structures [HORVÁTH et al. 1988]. Since these older structures are clearly visible on the seismic section, presumably the intrusion process, the developed heating effect and also the metamorphosis pervaded only partially this depth range. This hypothesis is reinforced by the theoretical studies of FURLONG, FOUNTAIN [1986] according to which relatively smaller amounts of intrusions (with smaller heat content) are formed from shallow asthenosphere than from a deeper position. Partial melting as a result of the heating effect might have affected only those parts of rock components having lower melting point. The rock-frame could mostly preserve the earlier structure of the depth range (which is verifiable seismically) but because of the partial melting it could become so permeable in smaller domains that arrangement of melt parts into horizontal bodies was possible owing to gravitational and thermal fields. They jointly constitute the new reflecting crust-mantle boundary whose depth cannot differ essentially from the basic/ultrabasic boundary detectable by the refraction method. During the extension period both deformations and intrusion (also volcanic) activities are likely to be divided into several time phases.

Repetition of the described phenomena can be assumed. 'Crust-mantle boundaries' forming in this way are probably situated under each other as, because of the deepening of the asthenosphere and thermal decrease of the lithosphere, isotherms moved towards the depth. On the basis of seismic results [POSGAY et al. 1981, 1986, 1991a] it can be supposed that some deep faults pass transversely the whole lithosphere or a significant part of it. Taking into consideration the elevated position of the asthenosphere during the extension period, further the repeated reactivating of the faulting [POSGAY, SZENTGYÖRGYI 1991] it is likely that these fractures played a great part in the outlined processes. It could give a reason for the fact that on strongly tectonized parts of the section the displacement of the crust-mantle boundary towards the depth might have been greater than the average.

## 4. Discussion/Conclusions

The composition and structure of the crust-mantle boundary may depend on the geological history of the lithosphere [HALE, THOMPSON 1982]. On the basis of deep seismic investigations along the Pannonian Geotraverse the interpretation of displacement of the boundary towards the depth beneath the studied part of the Great Hungarian Plain seems to be reasonable. Presumably the recent crust-mantle boundary is situated in the former (before the basin formation) upper mantle. From locations of reflecting horizons that correspond to the earlier crust-mantle boundary and from the neighbouring structural elements, piling up and upthrust can be concluded. The recent boundary is assumed to extend beneath the former boundary/boundaries. Its formation may be connected with the evolution of Pannonian basins. In this time range the lithosphere could have had a main extension of approximately EW direction [ROYDEN 1988] which is almost perpendicular to the Pannonian Geotraverse.

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## A KÉREG-KÖPENY HATÁR KIALAKULÁSA A KORÁBBI FELSŐKÖPENYBEN

POSGAY Károly

A Pannon Geotraverz szeizmikus szelvényén a kéreg-köpeny határon átnyúló szerkezeti elemek láthatók [POSGAY et al. 1990]. Vizsgálatukból arra lehet következtetni, hogy a jelenlegi kéreg-köpeny határ a korábbbi felsőköpenyben helyezkedik el. E határfelület mélység felé történő vándorlásakor a kéreg-köpeny határ mélységtartományának szerkezete nem semmisült meg, az a szeizmikus szelvényeken tanulmányozható.

Feltételezhető, hogy a fenti kép a Pannon-medence kialakulásával hozható összefüggésbe. A litoszféra extenziós igénybevétele következtében a kéreg-köpeny határtartományba hatoló intrúziók és a kis mélységben elhelyezkedő asztenoszféra hatására részleges olvadás következett be. Ezek együtteséről kapott legalsó, kiemelkedő amplitúdójú reflexió alapján valószínűsíthető az új kéreg-köpeny határ.

A leírt folyamat ismétlődését lehet feltételezni az egymás alatt megfigyelhető határokból.

## ОБРАЗОВАНИЕ ГРАНИЦЫ КОРА-МАНТИЯ ВНУТРИ ДРЕВНЕЙ ВЕРХНЕЙ МАНТИИ

Карой ПОЖГАИ

На сейсмическом разрезе Паннонского Геотраверса наблюдаются структурные элементы, пересекающие границу кора-мантия [POSGAY et al. 1990]. Предпологается, что современная граница кора-мантия находится внутри древней верхней мантии. При перемещении вниз этой границы не уничтожились ранее существующие структуры в окрестности границы кора-мантия, поэтому их можно исследовать по сейсмическим разрезам.

Можно предпологать, что выявленная ситуация связана с происхождением Паннонского бассейна. Под влиянием интрузий, внедрявщихся в граничную область кора-мантия вследствие экстенсионной загрузки литосферы, а также неглубокого расположения астеносферы, произошло частичное расплавление. Расположение новой границы кора-мантия, вероятно, прослеживается по самым глубоким сейсмическим отражениям большой амплитуды от данного комплекса. Судя по выявленным друг под другом границам, предпологается, что описанный процесс повторялся несколько раз.

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# COMPARISON OF SUBDUCTION ZONES VERSUS THE GLOBAL TECTONIC PATTERN: A POSSIBLE EXPLANATION FOR THE ALPS-CARPATHIAN SYSTEM*

Carlo DOGLIONI**

Apparently, plates are not moving randomly over the Earth's surface but rather following a common flow. Moreover the sum of plate motions detected in the hot-spot reference frame indicates that plates are moving 'westward' relative to the mantle. A reference point within the mantle should then move 'eastward' passing beneath different sections of the lithosphere, enabling a mantle source to be located at different times under continental or oceanic lithosphere. This can also explain the major differences between thrust belts related to subductions following or opposing the relative 'eastward or northeastward' mantle counter flow. These statements could provide a tool for the Alps-Carpathian puzzle. The Alps are a thrust belt mainly related to an E-dipping subduction following the 'eastward' mantle flow. They have high elevation, shallow foredeep, deep crustal rocks involved and no back-arc basin. In contrast, the Carpathians are related to a subduction opposing the 'eastward or northeastward' mantle flow, they have low elevation, deep foredeep, shallow rocks involved and the Pannonian back-arc basin. The Central-Eastern Alps are characterized by dextral transpression, the Western Carpathians by sinistral transpression. The Vienna Basin is located in the area of transfer between the two systems controlled by different subduction polarity.

**Keywords: tectonics, subduction zones, thrust belts, Alps-Carpathian system**

## 1. Introduction

The relative motion vectors between lithosphere and underlying mantle appear to follow global flow lines (*Fig. 1*) which can be constructed by linking axes of extension and compression over the Earth's surface [DOGLIONI 1990]. The flow lines of the last 40 Ma are generally WNW-ESE (E-W), with an about 15 000 km wavelength undulation showing gradual and progressive

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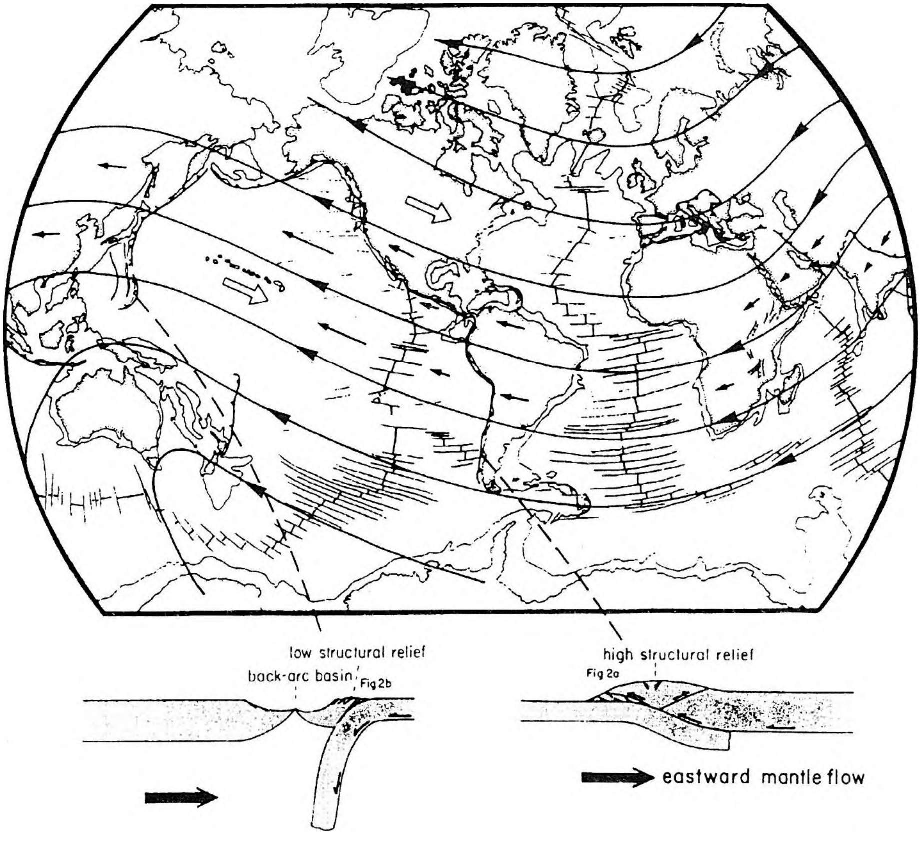


*Fig. 1.* This map shows the inferred flow lines along which plates move relative to the underlying mantle. These flow lines also indicate the westward polarity of plate motions (small black arrows) relative to the eastward mantle flow (big white arrows). Note below that this global tectonic polarity is responsible for the different dip of the subduction zones. The locations of Fig. 2a and Fig. 2b are shown. [Modified after DOGLIONI 1990]

*1. ábra.* A térkép a köpenyhez képest mozgó lemezek feltételezett áramlási vonalait mutatja. Ezek az áramlási vonalak a lemezmozgások nyugati polaritását is jelzik (kis fekete nyilak) a keleti irányú köpenyáramláshoz képest (nagy fehér nyilak). Figyeljük meg a térkép alatti két rajzon, hogy ez a globális tektonikai polaritás okozza a szubdukciós zónák eltérő lehajlását. A 2a és 2b ábrák helyeit feltüntettük. [DOGLIONI 1990 nyomán]

*Рис. 1.* На карте показано предпологаемое направление движения плит относительно мантии. Линии течения показывают также и западное движение плит (мелкие черные стрелки) относительно к восточному мантийному течению (большие белые стрелки). По рисункам, расположенным под картой, можно выявить, что эта противоположная глобальная тектоническая полярность вызывает противоположное направление субдукции. Отмечены участки, показанные на Рис. 2a и 2b. [По DOGLIONI 1990] variation in orientation. Plate tectonics can be analysed taking into account the net westward s.l. drift of the lithosphere relative to the mantle [LE PICHON 1968, BOSTROM 1971, NELSON, TEMPLE 1972], which is not an artifact of a particular hot-spot reference frame but a real physical observation which can be produced in a toroidal field by the lateral heterogeneities both in the mantle and in the lithosphere [RICARD et al. 1991]. On this basis plate tectonics may be considered as a consequence of variable decoupling at the lithosphere base as a function of the mantle anisotropies [DZIEWONSKI 1984, RICARD, VIGNY 1989, DOGLIONI 1990]. Simply stated, when there is compression or transpression between two plates, it is the eastern plate which moves faster westwards relative to the underlying sublithospheric mantle. If there is extension or transtension, it is the western plate that moves faster westwards or, more precisely, along the undulate global flow of the lithosphere. Lithospheric subduction [DEWEY 1981, UYEDA 1982, VON HUENE, LALLEMAND 1990], especially if it dips westward, produces an obstacle to the eastward flow of the mantle [NELSON, TEMPLE 1972, UYEDA, KANAMORI 1979]. Plates move following a well defined global tectonic pattern (Fig. 1) showing a major undulation of motion between eastern Africa and western Pacific [MINSTER, JORDAN 1978, GORDON, JURDY 1986, DOGLIONI 1990]. The same plate motion directions have been pointed out by geodetic satellite analysis [SMITH et al. 1990]. This paper tries to compare the general features of thrust belts with the subduction zones following (E- or NE-dipping) or contrasting (W- or SW-dipping) the mantle flow (Fig. 1). This subdivision seems to be more important with respect to the thickness and type of lithosphere involved in the collisional processes.

An E-dipping subduction can start only if there is an original thinner lithosphere to the west relative to a thicker lithosphere to the east. In contrast, we need to activate a W-dipping subduction in the presence of a thinner lithosphere to the east relative to a thicker one to the west. This is independent of the nature of the subducted lithosphere (both oceanic and continental). However the thinner and more oceanic lithosphere subducts more easily. Lateral thickness variations are a first order factor in controlling the amount and style of subduction. Moreover, longitudinal lithospheric variations in thickness and composition are able to produce strong asymmetry along the subduction zones, e.g. the Apennines which are characterized by a thin crust in the Ionian Sea and a relatively thicker crust in the Adriatic Sea to the north along the W-dipping subducting slab.

## 2. Subductions versus thrust belts

E- (or NE-) dipping subductions follow the mantle flow (Fig. 1). They are associated with thrust belts with huge exposures of basement rocks (also lower crust), high structural and morphologic reliefs in contrast to limited and usually shallow foredeeps (e.g. American Cordilleras, Western Alps, Dinarides, Zagros, Himalayas, *Fig. 2a*). The area of active compression may be very wide (hundreds of km), with extensional isostatic collapses in the internal core [PLATT 1986, DEWEY 1988]. Only in this kind of thrust belt have coesite-pyrope-bearing assemblages and eclogites been found [CHOPIN 1984, WANG et al. 1989], confirming that thrust sheets have detachment planes connected with depths ranging between 20 and 30 kbars (almost the lithosphere base). Collision should continue until the vertical lithostatic stress exceeds the horizontal shear values.

W- (or SW-) dipping subductions contrast with the mantle flow. They are instead associated with thrust belts involving high layers of the crust, back-arc basins and very consistent foredeeps generated by the roll-back of the subduction hinge (e.g. West Pacific accretionary wedges, Barbados, Apennines, Carpathians, *Fig. 2b*). W-dipping subductions are characterized by an eastward migrating tectonic wave (back-arc extension to the west and compression to the east). The extension continuously propagates and cross-cuts into the previously formed thrust belt. The area of active compression in this kind of thrust belt is very narrow (a few tens of km). Deep rocks may be involved in this type of accretionary wedge only if they were in a high structural and morphologic position before the onset of the W-dipping subduction (e.g. the granulite rocks of Calabria which emplaced during an earlier Eoalpine E-dipping subduction). The accretionary wedge is mainly formed by stacking of upper layers of the lower plate.

Note that in the E-dipping subductions the basal and intralithospheric decollements of the eastern actively thrusting plate are transmitted at the surface and provide a mechanism for bringing deep crustal levels to the surface (Fig. 1), whereas in the W-dipping subductions the base plate detachment is never connected to the surface, but it is rather folded and subducted itself. In fact the W-dipping case produces more superficial detachments in the accretionary wedge spatially followed by an eastward migrating extensional wave.

The two types of thrust belts should be considered as two end members. Oblique and lateral subductions (e.g. the Gibraltar arc, Betics and Maghrebides, etc.) have to be further distinguished in between. In fact, thrust belts parallel or slightly oblique to the mantle flow (Chaman transform zone, Central Alps, Pyrenees, Maghrebides etc.) are in general connected to body forces and/or second order rotations of blocks. Further distinctions have to be made on the basis of the relative thickness and viscosity contrasts between colliding plates, e.g. oceanic and continental (Andes) or continental collision (Himalayas): both are related to subductions following the mantle flow, but they are associated with different pre-existing lateral heterogeneities in the lithosphere which control the amount and kind of subduction.

A third type of subduction and related thrust belt is that produced by localized rotation of microplates, e.g. the counterclockwise rotation of Spain which produced the Pyrenees. This kind of subduction does not generally follow the mantle flow and is associated with very low or absent volcanism.

The origin of foreland basins in the two different thrust belts could consequently be controlled by different geodynamic factors: In the E-dipping case with the load of the thrust sheets [QUINLAN, BEAUMONT 1984], whereas in the W-dipping subduction, where the topographic load is insufficient to generate deep foredeeps basins [ROYDEN, KARNER 1984], the subsidence is allowed by the eastward retreating of the subduction hinge, due to the eastward push of the mantle. Dextral transpressive subductions characterize E-W oriented chains like the Central-Eastern Alps, the Betics and Maghrebides chains. Then a third type of foredeep is that produced at the transpressive subductions (e.g. the Venetian foredeep in front of the Southern Alps), or in front of thrust belts produced by rotation of microplates (e.g. the Pyrenees).

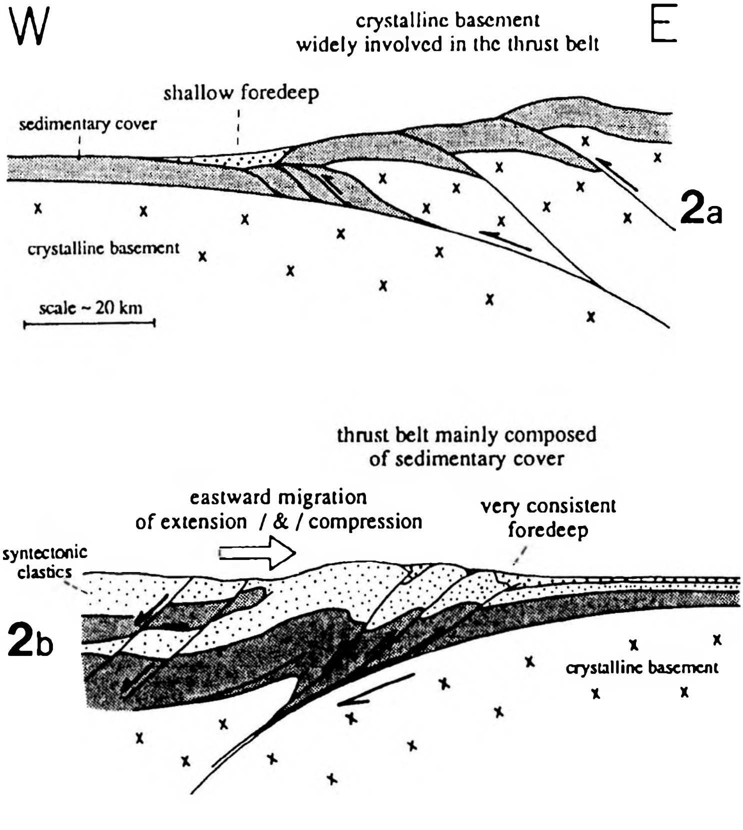


*Fig. 2.* Main features and structural differences between thrust belts associated with E-dipping (2a) and W-dipping (2b) subductions

*2. ábra.* A kelet felé (2a) és a nyugat felé (2b) hajló szubdukciókhoz kapcsolódó gyűrődéses övek fő vonásai és a közöttük lévő szerkezeti különbségek

*Рис. 2.* Характеристика и главные структурные различия складчатых поясов, связанных с восточной (2a) и западной (2b) субдукцией

## 3. Discussion

E-dipping subductions almost everywhere activate a backthrusting accretionary wedge, like all the southern and northern American Cordilleras. Consequently these E-vergent thrust belts with metamorphic core complex (e.g. Rocky Mountains) are related to E-dipping subductions and have to be distinguished from E-vergent thrust belts associated with W-dipping subductions characterized by the shallow rocks involved [e.g. the Apennines, BALLY et al. 1986]. We can also note that in W-dipping subductions the tangent to a pre-deformation marker descends into the trench, while in E-dipping subductions the same marker would rise towards the hinterland (Fig. 2). Moreover from the lithology of old thrust belts we could reconstruct whether they were formed during E- or W-dipping subductions: the deep rocks of the Caledonides should then have been formed with a subduction following the mantle flow. From the present orogenic belts and from their associated subduction we can predict the pre-existing shape and geographical position of the lithospheric stretching: this is a new tool in the reconstruction of Tethyan basins. Another consequence of the model is that magma sources positioned in the mantle might continuously move along the flow lines, i. e. bringing the continental mantle beneath the oceanic crust (*Fig. 3*).

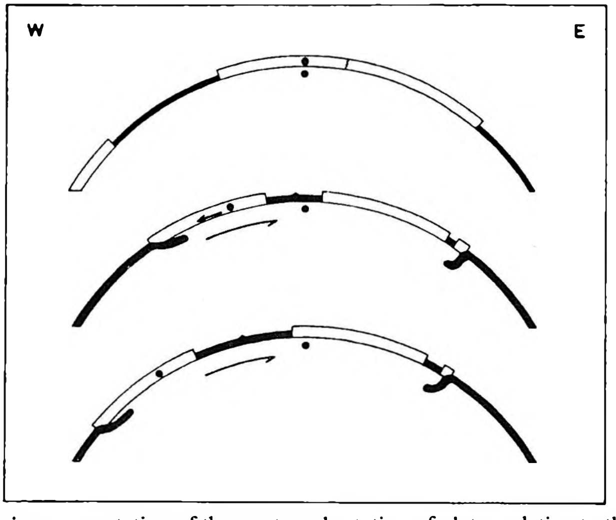


*Fig. 3.* Schematic representation of the westward rotation of plates relative to the underlying mantle. Note that the reference point in the mantle is continuously moving below a different lithosphere

*3. ábra.* A lemezek köpenyhez viszonyított nyugati irányú forgásának vázlata. Figyeljük meg, hogy a köpenybeli referenciapont mindig más litoszféra alá mozdul el

*Рис. 3.* Схема западной ротации плит относительно к мантии. Наблюдается, что мантийная точка относительности движется под разными литосферными блоками

## 4. Regional application of the model

We now try to investigate the Alps-Carpathian system in terms of the global tectonic pattern. For a recent review of the two main features, see ROYDEN, HORVÁTH [1988] and COWARD et al. [1989] and references therein. In the Carpathian-Pannonian basin, subsidence starts at the Badenian (16.5 Ma), and thrusting in latest Oligocene-early Miocene. ROYDEN et al. [1982] described an eastward migration of the volcanic activity, a current vertical slab, estimated by fault plane solutions, and a downbending of the subducted plate. The estimated extension (75 km west Carpathians —100 km east) is comparable to the synchronous crustal shortening and the foredeep is about 8 km deep [ROYDEN, KARNER 1984].

Farther west, the NE-dipping Dinaric subduction started at least in the Late Cretaceous times and partly stopped in the Miocene [BURCHFIEL 1980]. The dipping of the slab is gentle and the foredeep depth ranges between 3-4 km. Due to the lack of data, no estimation of the shortening has been published. The Variscan basement widely crops out in the internal zone.

On the other side of the Adriatic Sea the Northern Apennines, characterized by a W-dipping subduction, display a completely different geometry. The foredeep is quite deep like in the Carpathians, up to 8 km at the Messinian reflector, and the strong flexure of the subducting plate cannot be explained by the topographic load of the chain [ROYDEN, KARNER 1984]. Even if the basement has to be involved in depth by the thrusting, it never crops out (apart from limited internal zones, e.g. Apuane Alps, Tuscany), and the main decollement level is the top of the Trias and the thrust sheets are formed by the sedimentary cover.

Recent deep seismic data of the Alps allowed new structural interpretations of the chain. In the Alpine section, W-vergent thrusting cutting throughout the crust into the mantle supports the 'obduction' of the Adriatic plate on to the European foreland. The Alps are the product of the Late Cretaceous to the present closure of a lithosphere thinned during Late Permian-early Middle Mesozoic. In spite of several paleogeographic problems, we shall try to analyse alpine tectonics in terms of relative activity of the decollement at the base of the lithosphere. During the initial phases of extension the western European continent should have had a plane of detachment more active with respect to the eastern Adriatic plate in order to produce the Tethyan rifting. The Cretaceous inversion [DAL PIAZ et al. 1972, POLINO et al. 1990] would correspond to the inversion of the relative velocity which should become greater at that moment beneath the Adriatic plate. Kinematic indicators in the Central Alps give a dominant E-W or NW-SE sense of relative motion from Late Cretaceous time [PLATT et al. 1989] suggesting that the motions of Adria and Africa were more or less independent from that time and characterized by different amounts of decoupling at their base. The plate margins evolved through time as a function of differences in plate velocity, e.g. the northern Adriatic plate margin was represented by the Austroalpine units during extensional phases and Eoalpine and probably Mesoalpine inversion, while it is now represented at the surface by the Insubric Lineament which is the present boundary of relative plate velocity contrast. In order to get dextral transpression along the Insubric fault zone we need a greater detachment at the Adriatic plate base with respect to the European one. Extension occurred within the Alpine edifice and around it at different stages of the orogenic evolution [e.g. the Oligocene phase, LAUBSCHER 1983]. This extension [DAL PIAZ 1976] has also been interpreted as due to isostatic re-equilibrations [PLATT 1986, DEWEY 1988]. Another possibility is that the load and the thickening of the lithospheric doubling in the E-dipping subductions would produce a disactivation (or decrease) of the detachment at the base of the eastern plate, generating a relatively greater westward drift of the western plate, which is responsible for extension. In other words, the collision should pass alternating phases of compression or extension as a function of the activity of the basal (or intra) lithospheric detachments. *Fig. 4* presents a schematic picture of the Neogene to recent Alps-Carpathian system pointing out that the main frame might be analysed in terms of different westward decoupling of the lithosphere relative to the mantle.

The Mediterranean region might be interpreted in terms of differences of the base lithosphere decoupling [DOGLIONI et al. 1991]. The detachment occurs at the base of an anisotropic and segmented lithosphere which overlies an eastward (northeastward) directed mantle flow. This is in agreement with the general eastward rejuvenation of extension and magmatism observed in the Central-Western Mediterranean. The region is located within the global undulation of the flow lines interpreted as generated by the instability of the rotation axis. In fact the main motion of the mantle relative to the lithosphere seems to be E-W directed in the western Mediterranean, gradually changing to ENE-WSW trends [MANTOVANI et al. 1987] in the eastern regions. Regional variations in the general trend may be interpreted as due to body forces like transpressions or transtensions (e.g. the Central-Eastern Alps, the Betics and Maghrebides, or the Southern Carpathians) or to second order tectonic fields produced by rotations of microplates (e.g. the Pyrenees) which could generate for instance N-S compressions. The Mediterranean tectonics is a good example of lateral variations in lithosphere thicknesses and compositions, which are fundamental factors in controlling the rate of subduction and the relative velocities among plates.

## 5. Concluding remarks

The model described in the previous sections is applicable to the Mediterranean where the eastward or northeastward relative migration of the underlying mantle with respect to an inhomogeneous disrupted lithosphere could explain the puzzling tectonic evolution of this area [DOGLIONI et al. 1991]. The geodynamics is complicated by N-S compressions resulting from second order rotations (e.g. Spain and Adriatic plates) and transpressions induced by body

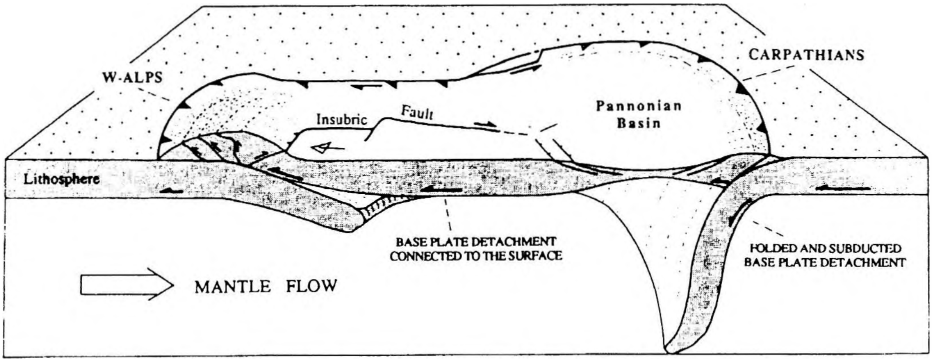


*Fig. 4.* Schematic picture of the Neogene to recent Alps-Carpathian system analysed in terms of different relative westward velocities of the lithosphere relative to an eastward-northeastward directed mantle flow. Note that the back-arc extension in the Pannonian Basin is a direct consequence of the eastward retreating of the W-dipping subduction hinge zone due to the eastward push generated by the mantle flow. Note that the Vienna Basin is the area of separation between the relative plate motions producing dextral transpression in the Central-Eastern Alps and sinistral transpression in the Western Carpathians. [After DOGLIONI et al. 1991]

*4. ábra.* A keleti-északkeleti irányú köpenyáramláshoz képest nyugat felé mozgó litoszféralemezek sebességkülönbségei alapján megrajzolt vázlat az Alpok-Kárpátok rendszer kialakulásáról a neogéntől napjainkig. Figyeljük meg, hogy az ív mögötti extenzió a Pannon medencében közvetlen következménye a nyugat felé hajló szubdukciós zóna keleti irányú hátrálásának, amelyet a köpenyáramlás keleti irányú tolóhatása eredményez. Látható, hogy a Bécsi medence a Közép-Kelet-Alpokban jobbos, a Nyugati-Kárpátokban pedig balos transzpressziót okozó relatív lemezmozgásokat elválasztó terület. [DOGLIONI et al. 1991 nyomán]

*Рис. 4.* Схема происхождения Альпийско-Карпатской системы с неогена до современности, построенная по разности скорости западного движения литосферных плит относительно мантийному течению. Обнаруживается, что задужная экстенсия в Паннонском бассейне является непосредственным следствием перемещения на восток зоны субдукции с западным склонением, что связано с восточным движением мантийного течения. Показано, что Венский бассейн является участком, разделяющим Центральные-Восточные Альпы, где относительное движение плит вызывает правую транспрессию, от Западных Карпат, где оно приводит к левой транспрессии. [По DOGLIONI et al. 1991]

forces. The present Mediterranean or Alpine-Carpathian shape (Fig. 4) is a direct function of the inherited Mesozoic lateral heterogeneities of the lithosphere, which are, in turn the key point for relative differences in velocity and subduction. LAUBSCHER [1988] and ROYDEN and BURCHFIEL [1989] pointed out the main differences between 'Alpine' thrust belts (push arc) and 'Apenninic or Carpathian' thrust belts (pull arc). These differences may be interpreted as a consequence of the different position and behaviour of the detachment planes in west or east dipping subductions.

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# A SZUBDUKCIÓS ZÓNÁK ÉS A GLOBÁLIS TEKTONIKA ÖSSZEHASONLÍTÁSA: AZ ALPOK-KÁRPÁTOK RENDSZER KIALAKULÁSÁNAK EGY LEHETSÉGES MAGYARÁZATA

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A lemezek nyilvánvalóan nem rendezetlenül mozognak a Föld felszínén, hanem egy közös áramlási irányt követnek. Sőt a hot-spot vonatkoztatási rendszerben megfigyelt lemezmozgások összege azt mutatja, hogy a lemezek a köpenyhez képest „nyugat felé" mozognak. Így egy köpenybeli vonatkoztatási pontnak a litoszféra különböző részei alatt elhaladva „kelet felé" kell mozognia, így ugyanaz a köpenybeli forrás különböző időpontokban egyszer a kontinentális, máskor az óceáni litoszféra alatt található. A fenti elmélet megmagyarázhatja a „keleti vagy északkeleti" köpenyáramlással azonos, illetve ellentétes irányú szubdukciókhoz kapcsolódó gyűrődéses övek fő különbségeit. Ezen állítások segítségével közelebb juthatunk az Alpok-Kárpátok rendszer megértéséhez. Az Alpok egy olyan gyűrődéses öv, amely a „keleti" köpenyáramlással azonos irányú, kelet felé hajló szubdukcióhoz kötődik. Nagy tengerszint feletti magasság, sekély előtéri medence, a kéreg mélyebb részeiből származó kőzetek és az ív mögötti medence hiánya jellemzik. A Kárpátok ezzel ellentétben a „keleti-északkeleti" köpenyáramlással szemben végbement szubdukcióhoz kapcsolódik, így kisebb tengerszint feletti magasság, mély előtéri medence, kis mélységből származó kőzetek és a Pannon, ív mögötti medence jellemzik. A Közép-Kelet-Alpokban jobbos transzpresszió, míg a Nyugati-Kárpátokban balos transzpresszió figyelhető meg. A Bécsi medence e két különböző irányítottságú szubdukciós rendszer által kijelölt átmeneti területen helyezkedik el.

# СОПОСТАВЛЕНИЕ ЗОН СУБДУКЦИИ И ГЛОБАЛЬНОЙ ТЕКТОНИКИ: ВОЗМОЖНОЕ ОБЪЯСНЕНИЕ ПРОИСХОЖДЕНИЯ АЛЬПИЙСКО-КАРПАТСКОЙ СИСТЕМЫ

Карло ДОЛЪЕНИ

Движение плит на поверхности Земли явно не является беспорядочным, а соответствует некоторому общему направлению. Суммирование наблюдаемых движений коры показывает, что плиты движутся в западное направление относительно мантии. Следовательно, если точку относительности разместить в мантию, то она должна двигаться на восток под разными частями литосферы, и поэтому тот же источник в разное время может находиться как и под океанической, так и под континентальной литосферой. Такая гипотеза может объяснить различия между складчатыми поясами, связанными с субдукцией по направлению согласной и противоположной восточному или северо-восточному мантийному течению. Применение такой гипотезии способствует интерпретации Альпийско-Карпатской системы. Альпийский складчатый пояс связан с субдукцией в восточном направлении, т.е. в направлении мантийных течений.

Он характеризуется значительной высотой над уровнем моря, неглубокой предгорной впадиной, наличием пород, образовавшихся в глубоких зонах коры и отсутствием задужного бассейна. Наоборот, Карпаты связаны с субдукцией в противоположном мантийным течениям направлении и характеризуются меньшей высотой, глубокой предгорной впадиной, породами, образовавшимися в небольшой глубине, и наличием Паннонского — задужного — бассейна. В Центральных-Восточных Альпах наблюдается правая, а в Западных Карпатах—левая транспрессия. Венский бассейн находится на промежуточной территории между двумя системами.

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# SEISMIC IMAGING OF POROSITY AND HYDROCARBON IN CONSOLIDATED FORMATIONS

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Consolidated reservoir formations are defined here as those formations of significant thickness having acoustic impedance (density-velocity product) which is consistently greater than or equal to that of similarly aged, normally pressured shales. For such formations, porosity and hydrocarbon presence can be effectively imaged via seismic velocity changes over a substantial range of depths (to 16,000-plus feet, to 4800 m). Here we shall note consolidated sand and carbonate reservoirs from East Texas, the North Sea and Oklahoma. Plays of current interest encompassing such reservoirs include the Austin Chalk, Alaskan formations, Niobrara and Frio-Vicksburg, as well as others.

Starting with a brief summary of requirements for definitive seismic imaging of consolidated reservoirs we move on to the technology for interpreting, calibrating, and using such displays. The effectiveness of spatially dense seismic velocity measurements and color inversion sections scaled in velocity is clearly demonstrated by the variety of successful examples reviewed. These approaches and the positive results obtained suggest that future seismic techniques will be even more definitive in addressing consolidated reservoirs for both exploration and production applications.

**Keywords: consolidated formations, hydrocarbon detection, sand/shale reflectivity, seismic imaging, velocity**

## 1. Introduction

Our approach to hydrocarbon exploration and reservoir definition rests principally on sound basic geology and refined seismic methods. There are two themes which are applied in parallel to the seismic data. Significantly improved and definitive seismic imaging constitutes one of these directions. Use of a unified framework, an organized sequence of high technology procedures, to identify and qualify key anomalies and subsurface characteristics is the second theme.

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## 2. Enhanced seismic imaging and display

It was recognized more than 10 years ago that as much as 80% of the stratigraphic information contained in seismic data presented in conventional black and white displays was lost due to inappropriate processing and limitations in visual dynamic range [NEIDELL, BEARD 1984 and NEIDELL, BEARD 1985]. Indeed, properly conditioned data displayed in a suitable color scheme provides a 20-fold increase in detail over black and white displays. Moreover, when the data is rescaled (semi-quantitatively) as velocity, porosity and hydrocarbon presence may be detected and confirmed under a wide range of circumstances. Velocity scaling of the seismic data also allows definitive correlations and parameter predictions when used along with well logs.

To take full advantage of the extended visual range of color displays our processing sequence seeks to produce optimally imaged data. The presentation of relative amplitudes, collapse of the wavelet to a consistent zero phase character and full prestack migration are essential. Central to our method also is our intensive analysis of the velocity field using high resolution velocity spectra which are of proprietary design. These analyses perform well at reflection times where other analyses lose their resolution (typically below about 2.0 s for land data and 3.0 s for marine).

Analyses are computed 16 per mile for land data or 12 per mile for marine surveys on fully prestack migrated CDP gathers. When interpreted with due consideration for the geological setting, these produce a velocity field which not only yields an optimum stacked section, but can be used to derive a velocity model for the final poststack migration and the background velocity trend for the acoustic impedance section.

The color seismic inversion displays are derived exclusively from seismic inputs. Correlative studies with logs are performed later and only where appropriate.

Since seismic data are band limited, trace inversion provides only partial information about changes in acoustic impedance. Interval velocities calculated from the high resolution velocity field between carefully chosen horizons generally show good correspondence to the gross features of a sonic log where it should. This allows us to use the stacking velocity not present in the individual traces. Taken alone the low frequency trend gives us valid information on a large scale of relative changes along the profile. It may also be approximately 'matched' to well information for better correspondence in terms of magnitude.

The seismic data is inverted, combined with the trend and scaled to velocity. It is plotted in contrasting colors at 400 ft/s (120 m/s) velocity steps, and at scales of 10 in/s to 20 in/s (0.25 m/s to 0.5 m/s) depending on the objective of the project. We now have a section which displays for the interpreter the full potential of the seismic data.

## 3. Unified interpretive procedure

Interpretive overlays noting anomalies and key features are prepared as a final product from the inversion color displays. Existing maps, log correlations and all prior knowledge are incorporated into the overlay and final reporting. Anomalies are examined within the framework of geometric considerations described first and quite completely in the EXXON series of papers published in AAPG Memoir [PEYTON 1977].

Of course the velocity dimension inherent in the data is used fully as well. Intensive study of all the moveout velocities unambiguously identify lithologies such as salt, plastic shales, carbonates and sand-shale sequences. Also such studies identify lateral gradients which are critical for establishing true structural geometries and making effective time-to-depth conversion. The added visibility of the stratigraphic changes which result from the improved data processing and enhanced visual dynamic range display, in conjunction with the use of both geometry and velocity information, assure that no anomalies of significance will be overlooked.

As a further component of this second theme we suggest that the existing hydrocarbon reservoirs we recognize be subdivided into two categories: sands-sandstones, and carbonates and others (i.e. fractured lavas, shales, coal seam methane etc.). Our basic studies of reservoirs on a global scale following this philosophy have provided some extraordinary results and insights.

For example, in the first category of sand-sandstone reservoirs [NEIDELL, BERRY 1989] we have studied and documented the relationships of acoustic parameters for sands and shales and found it necessary to consider depth and geologic age as independent variables. The 'crossover' region of more or less equal sand-shale velocity and its special properties and problems have been considered in detail — this is of course the Zone II, which separates the Bright Spot world from the more consolidated Zone III lithologies below.

The Zone terminology was introduced in the 1986 SEG Distinguished Lecture (unpublished) to characterize Sand/Shale reflectivity relations essentially in all basins. Young unconsolidated sands should they be present, when water filled and under normal pressure conditions are lower in velocity and density than contemporary shales. This is termed Zone I regime. With age and burial the sands compact through the shale levels for velocity and density, becoming consolidated and now being the higher acoustic impedance (velocity-density product) component under analogous conditions. This is the Zone III environment for reflectivity. Where the transitioning in velocity and density occurs, the Zone II terminology is applied.

We see that the reduction in seismic measured velocities for hydrocarbons in the bright-spot world of Zone I is quite pronounced (1000-4000 ft/s, 300-1200 m/s). In this young, unconsolidated environment, where sand acoustic impedance is less than that of shale, the contrasts in reflection coefficient which result from gas presence can be two-fold, three-fold, even five-fold.

Hence we can see the signatures directly on conventional seismic displays of reflectivity where relative amplitudes have been preserved.

It is revealing that in Zone III below the crossover Zone II, the sands and sandstones have compacted and consolidated to acoustic impedance values greater than those of the shales. Hydrocarbons again drop velocities as seen by the seismic method almost as much as in Zone I (1000-2500 ft/s, 300-760 m/s), but now the changes in reflection coefficient owing to the generally higher acoustic impedance values only approach 10-15%. This level of change can not be seen on conventional seismic displays. Fortunately, when detailed data processing preserves the information, such velocity changes are clearly observed on the color displays scaled in velocity. These displays have the capability to present such changes most clearly and put their geometry and continuity in a structural context. Hence, hydrocarbons in deeper, older formations can be detected more readily than previously believed.

It is principally in Zone II for sand reservoirs that our technology is generally least effective. Here, we can not even use synthetic seismograms for correlative purposes with any reliability. While it is true that our success in finding and defining reservoirs in Zone II is diminished, the approach described here remains the single most effective technique for working in this regime. Substantial thicknesses of Zone II (1000-10 000 ft, 300-3000 m) are frequently encountered and the economic potential matches what we find for Zone I or III.

Carbonate reservoirs in terms of measured seismic parameters for porosity and hydrocarbons behave much like the Zone III consolidated sands. Clearly the geometries and geologic origins are quite different for these lithologies as compared to sands and sandstones.

Once the interpretation has been formulated with all appropriate information and assistance, we address the verification phase of anomalies as a separate matter. Indicated velocity drops in structurally favorable settings and within potential reservoir units can of course relate to reservoir presence, but may also be caused by bed thinning with subsequent tuning and detuning or by lithologic change. Instantaneous frequency analysis is sometimes an effective means for recognizing tuning effects. Model studies can separate tuning effect from effects caused by parameter changes relating to lithology or hydrocarbon presence. In Zone I sand, hydrocarbon presence causes increased negative reflection amplitude-with-offset as distinguished from lithology change. For Zone III, amplitudes with offset decrease with hydrocarbon presence as opposed to lithology change which cause no comparable effect. RUTHERFORD and WILLIAMS [1989] have documented that amplitude-with-offset studies are largely ineffective in Zone II.

Catalogues are developed which relate porosity and pore fluid to the magnitude of the velocity anomalies which are observed. Such techniques enable calibration studies to be used for predictive purpose, and often with remarkable accuracy. Research and practical study are blended effectively in such day-to-day operational interaction.

The ability to image our objectives, qualify them and establish their quantitative significance from conventional seismic data sets these procedures apart from the standard 'piece-meal' and less enlightened practices. Various levels of informational synergy are given opportunities to contribute to our knowledge in this way.

We see in this procedure a unity of approach which draws together most of the newer techniques into an organized framework for validation and confidence building. The same methods, of course, most fully define reservoirs using seismic data. Taken in concert with our calibration studies and fundamental work on the relation of seismic data to well logs, an approach is continuing to evolve which in the industry is one of the most effective means for using available seismic data.

## 4. Examples

North Sea Line 12 is shown in conventional seismic display in *Fig. 1.* We shall concentrate on only one formation for purposes of this discussion, the Bunter Sands which are marked. Faulting above the Bunter Sands is believed to relate to evaporate dissolution. Also, a structure is seen quite clearly in the Platten Dolomite which we note for reference as well.

The more or less counterpart color acoustic impedance display is also included in Fig. 1. While the horizontal scales in the two cases are not quite the same, the differences are dramatic. Here the Bunter Sandstone is clearly seen to be a transgressive sequence. Two of the barrier bars in particular appear to have hydrocarbon potential in terms of indicated velocity reduction. Likely porosity with hydrocarbons is seen in fact at the center of two of what are likely large Barrier Bar sand deposits. Note the white and dark blue areas at the center respectively of two of the barrier bars. Also, consider that this is an older consolidated sand formation.

We see in this case both the geologic time lines and the lithology change. The conventional black and white display has been driven well beyond the capabilities of its visual dynamic range. Well penetrations on both ends of the line beyond the portion shown indicate the member units of the Bunter sand are greater than 100 m in thickness, hence what we see is clearly not a thin bed effect. In fact, a quite similar situation is exhibited by the West Sak sands of the Kuparuk field in Alaska.

For our next example we consider the Austin Chalk in Dimmit and Zavalla counties, Texas supplied courtesy of Alliance Geophysical Co. (refer to *Figs. 2* and *3*). In general, Austin Chalk fracture zones are recognized by decreases in the seismic derived interval velocity within an appropriate member or unit of the Chalk as it is traced laterally. A velocity decrease between 400 to 800 ft/s (120 to 240 m/s) is considered diagnostic in terms of fracture development and oil saturation. Commercial gas accumulation would yield greater contrasts usually approaching and even exceeding 1000 ft/s (300 m/s).

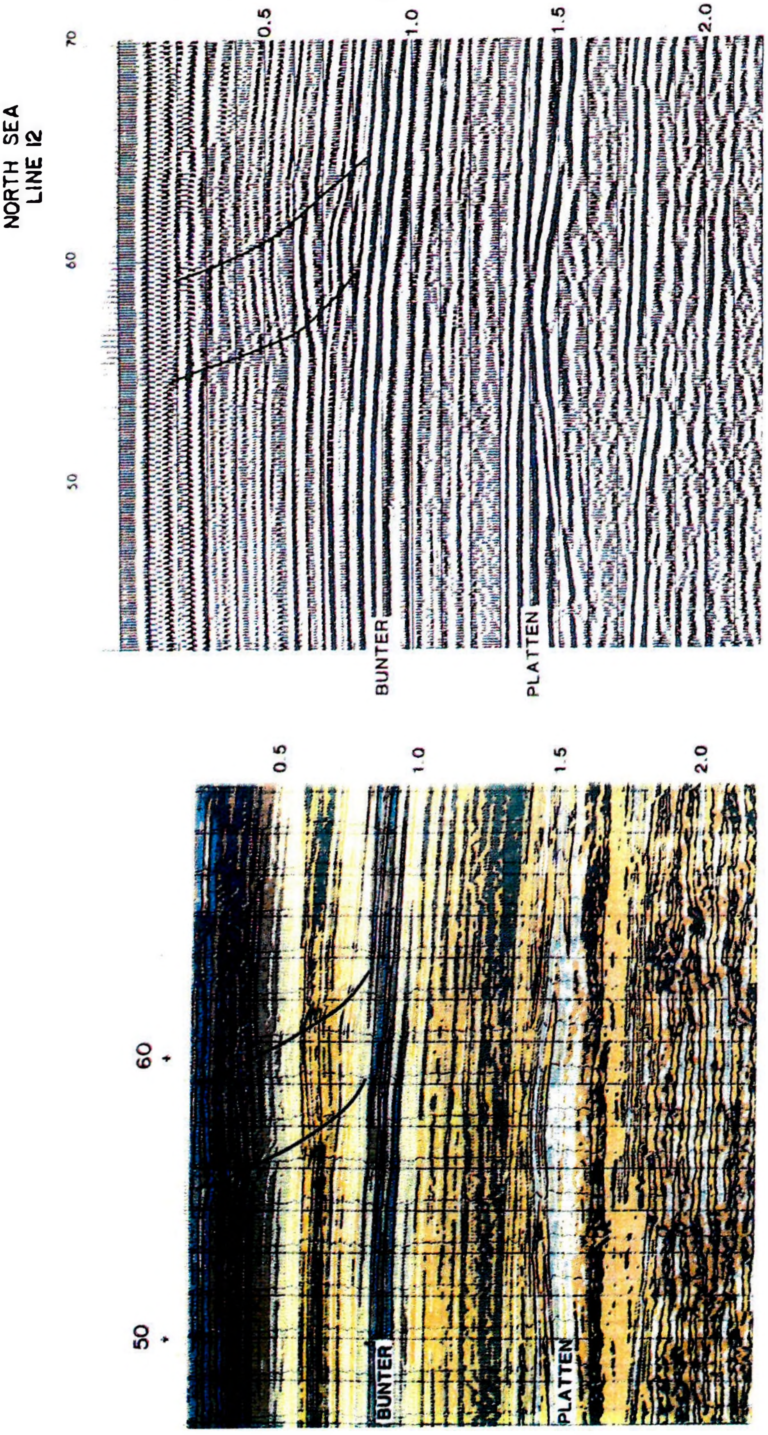


*Fig. 1.* Color inversion compared to conventional section — scales not quite equal

*1. ábra.* A színes inverz és a hagyományos szelvény összehasonlítása — a léptékek nem teljesen azonosak

*Рис. 1.* Сопоставление традиционного и цветного обратного разреза (масштабы немного различаются)

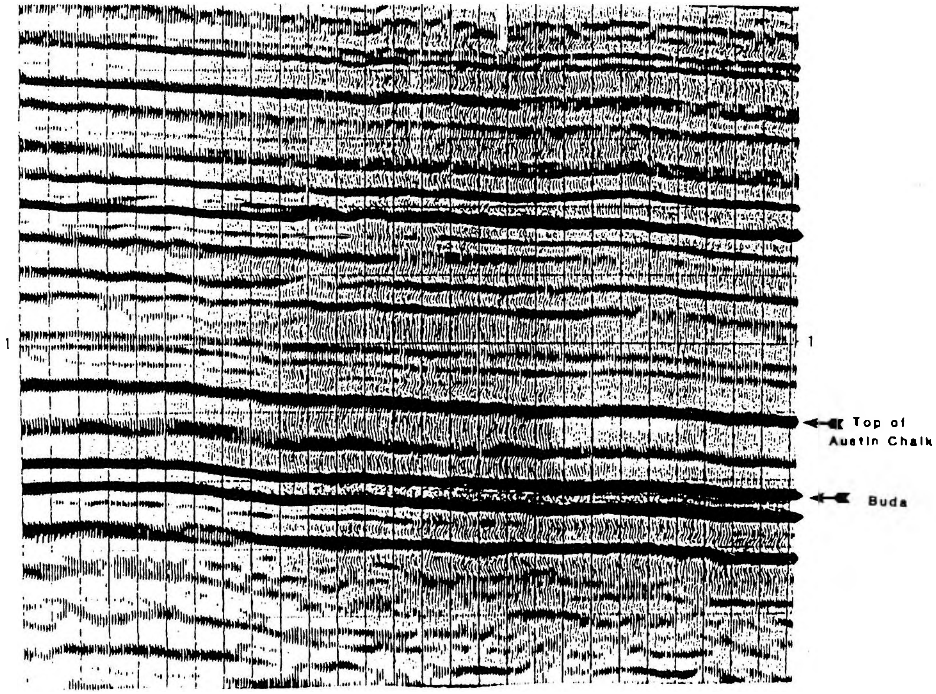


*Fig. 2.* Alliance Geophysical Co. Austin Chalk line — conventional display
*2. ábra.* Alliance Geophysical Co. Austin kréta vonal — hagyományos megjelenítés
*Рис. 2.* Профиль Аустинский мел Alliance Geophysical Co. – традиционное изображение

Within the Chalk as seen on the color inversion display of Fig. 3 we identified changes in acoustic impedance due to fractures and noted fracture zones within three rather distinct Chalk members. These were identified as an upper, middle, and lower member and correspond to similar groupings which might be made on the basis of well logs. Analogous interpretation is not possible using the conventional data display of Fig. 2 although anomalies are in fact indicated by the changes in amplitude which we can observe in this intensively processed seismic data. Of course relating the changes we see to porosity development and determining in which member they occur is quite another matter.

In general terms, the middle zone appears to have the lowest seismic velocity within the Chalk section (about 14 000 ft/s, 4200 m/s — light blue color — likely indicating the most shale streaking and ash beds). Also, the deeper member at about 16 600 ft/s (4900 m/s) in many places seems to be about 400 ft/s (120 m/s) higher in velocity than the upper zone (pumpkin-ma-

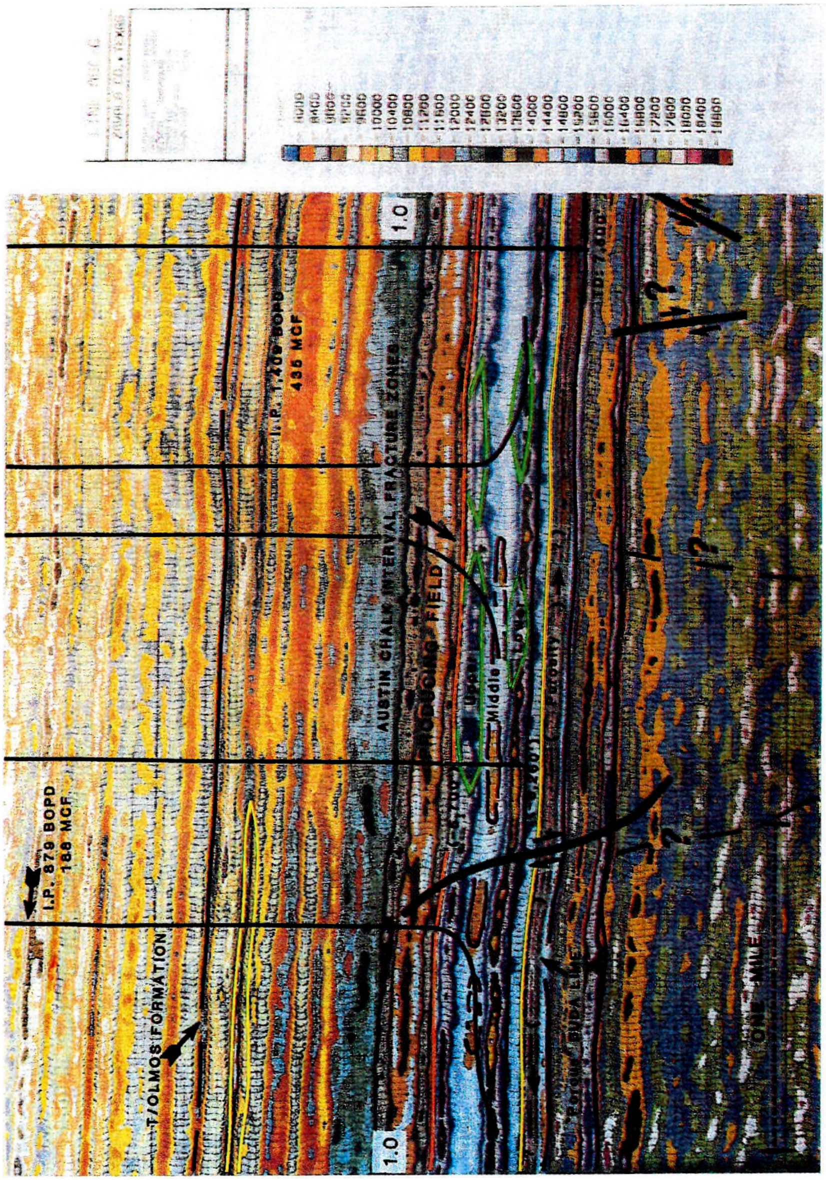


*Fig. 3.* Alliance Geophysical Co. Austin Chalk line color inversion display with interpretive overlay

3. *ábra.* Alliance Geophysical Co. Az Austin kréta vonal színes inverz megjelenítése és értelmezése

*Рис.3.* Цветное обратное изображение профиля Аустинский мел Alliance Geophysical Co. и его интерпретация roon color as opposed to maroon-purple color). These lithological distinction within the key chalk horizons and their respective properties explain why they can independently exhibit quite different fracture behavior.

The most frequently observed decrease in velocity noted on the color display for an oil filled fracture zone ranges from 400 to 800 ft/s (120 to 240 m/s). The lower value (400 ft/s, 120 m/s) is what we would expect for porosity indications within average carbonates and particularly for oil with no gas saturation. This observation is of course empirical, and in line with our prior experiences. Higher contrasts (above 1 000 ft/s, 300 m/s) were most often found to indicate commercial gas accumulations or extremes of porosity (25% or better). We saw the largest velocity reductions (1200 ft/s, 360 m/s) in association with the deepest fracture zone on a portion of the line not shown here at about 1.1 seconds where tight chalk was indicated by seismic interval velocities of about 16 600 ft/s (5060 m/s). Wells in porosity zones in that area showing the large velocity contrast had extraordinary production rates (many thousand of bbl per day).

The fracture zones in the middle Chalk member are seen most readily. Pumpkin color zones isolated within the light blue are quite clear. Contiguous zones of color contrast representing lowered velocity in the upper and lower Chalk members which depart from systematic regional variations similarly denote fracture zones. Again the accompanying velocity drops are about 400-600 ft/s (120-180 m/s).

Well control superimposed on the color section was provided after the fracture interpretation. There is in fact a remarkable correlation between the initial flow rates (IP's) and the quality and size of the fracture zones. Note that direct imaging of the fracture zones enables their 'ranking' and well designs to test them can be very specific in terms of location, horizontal reach and in identifying which Chalk member is to be penetrated. This translates clearly to great cost savings.

Projections of several well-bores into the plane of this seismic profile have been noted. By further comparing production information and in particular Initial Potentials (IP's) with the color inversion display anomalies, the most highly productive fracture systems in this area could be seen to occur within the middle and lower zones of the chalk (lower 400 feet (120 m) of chalk section). Such semi-quantitative information again has important impact on the economics of this play.

Finally, we can look at an area in Oklahoma where the Mississippian Carbonates and Bromide (Wilcox) Sand developments constitute the likely prospects. We shall note again how the careful processing and inversion display clearly identify a low risk additional well location. Seismic inversion and color display of one of the seismic profiles, Line No. 6, was performed in order that detailed velocities derived from seismic moveout could amongst other objectives aid in imaging the fractured Limestone reservoir. Low velocity anomalies are observed on Line No. 6 (color display, *Fig. 4*) within the fractured Limestone interval, from 0.915 to 0.950 seconds reflection time. These exhibit decrease of interval velocity up to 1200 ft/s (365 m/s) between shotpoints 270

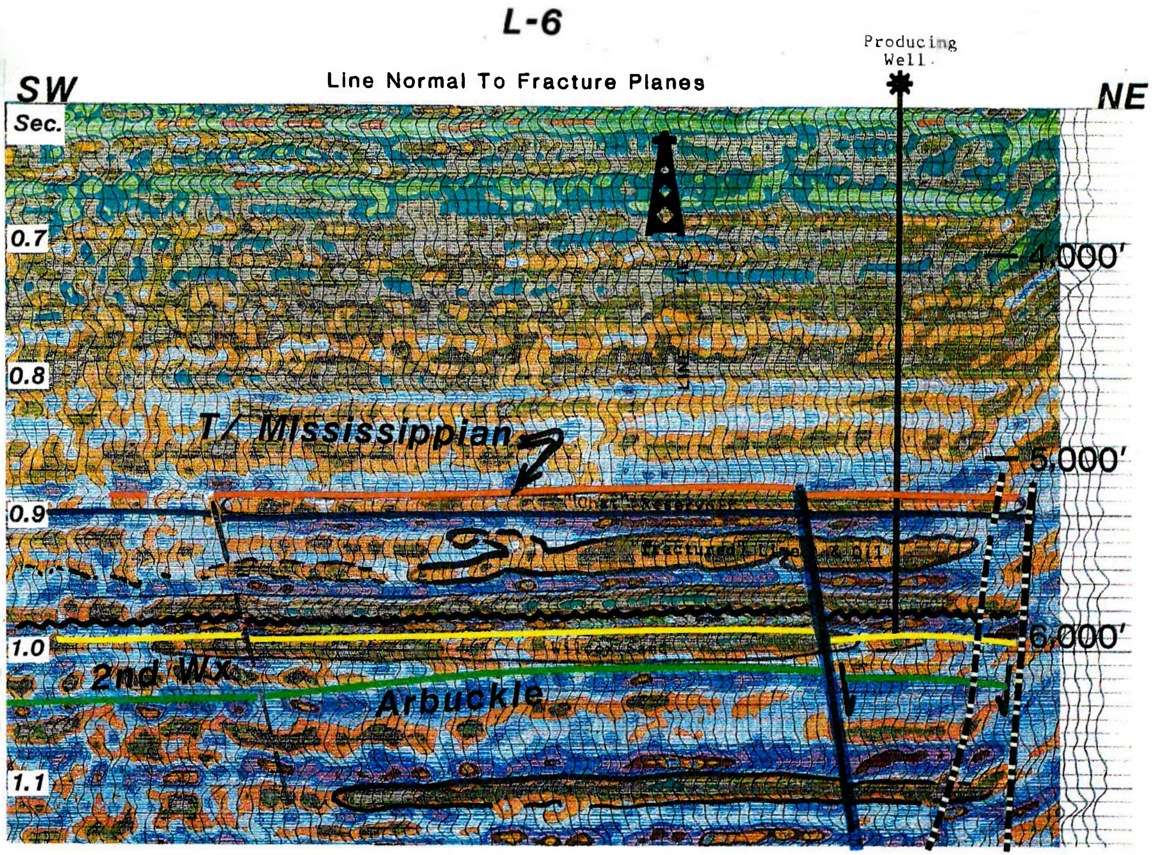


*Fig. 4.* Oklahoma case study showing well location and prospect — color inversion display
*4. ábra.* Az oklahomai esettanulmány a meglévő és tervezett fúrásokkal — színes inverz megjelenítés
*Рис. 4.* Разрез Оклахома с пробуренными и проектными скважинами (цветное обратное изображение)

and 335. A portion of the counterpart black and white section is included here as *Fig. 5* for inspection and comparison purposes.

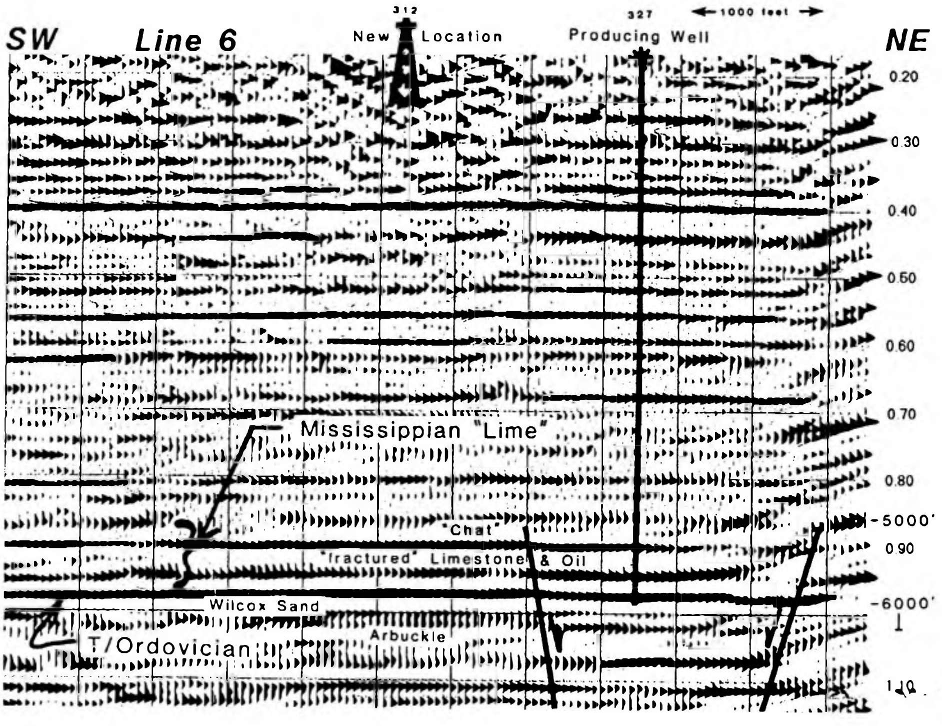


*Fig. 5.* Oklahoma case study — conventional display of specially processed data
*5. ábra.* Az oklahomai esettanulmány — speciálisan feldolgozott adatok hagyományos megjelenítése
*Рис. 5.* Разрез Оклахома – традиционное изображение данных, обработанных специальными методами

In order to unambiguously identify the reflection intervals and relate them to depth, a synthetic seismic trace was computed. The synthetic was derived from density data observed in a Neutron Density Log. Matrix velocity was established by sonic log control from area well locations. The resultant lithology/time tie to the seismic data was excellent, and the fractured (oil producing) zone can be correlated to mid-section of the Osage Lime.

Looking once again at the color display of Fig. 4 we should point out that the best fracture zone has brown coloring within the orange coloring indicating an additional 400 ft/s (120 m/s) velocity reduction. The best porosity is thus indicated by the largest continuous brown areas within the orange zones. While the orange zones represent a 400 ft/s (120 m/s) velocity reduction from the blue color, they are less likely to be productive or would not produce as well as the zones having even lower velocity (brown color). Of course the exact situation would have to be established by drilling. Nevertheless, with the color display, one is able to drill the best or at least most likely wells first. This is clearly not possible using the conventional data display of Fig. 5 alone.

## 5. Conclusions

It is difficult in a short exposition with a few illustrations to demonstrate all of the information which can be accessed in seismic data using the methods described. Further references [NEIDELL et al. 1985, NEIDELL et al. 1984] can provide some added dctail. Nevertheless, using three diverse cases, all of consolidated nature we show successful imaging of hydrocarbon filled porosity. Two of the case studies in fact show oil reservoirs. Gas reservoirs are easier to see owing to the substantially greater velocity drops. Success in this type of endeavor rests with the appropriate detailed processing of the seismic data, display of that data with enhanced visibility color displays, and of course a knowledgeable interpretive approach. Such results are indeed an excellent platform for application of even more advanced techniques to further qualify and even quantify the nature of hydrocarbon reservoirs in advance of the drill both in exploration and production environments.

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## KONSZOLIDÁLT FORMÁCIÓK POROZITÁSÁNAK ÉS SZÉNHIDROGÉN TARTALMÁNAK SZEIZMIKUS LEKÉPEZÉSE

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Konszolidált rezervoár formációk alatt olyan jelentős vastagságú formációkat értünk, amelyeknek akusztikus impedanciája (sűrűség-sebesség szorzata) a hasonló korú, normális nyomás alatt álló agyagpalákénál nagyobb vagy megegyezik azokéval. Ilyen formációkra a porozitás és a szénhidrogén tartalom a szeizmikus sebességváltozások alapján jelentős mélységintervallumra (mintegy 4800 m-ig) hatékonyan képezhető le. Az alábbiakban megemlítjük Kelet Texas, az Északi-tenger és Oklahoma néhány konszolidált homok és karbonát tárolóját. A szóban forgó tárolókat tartalmazó, napjainkban reménybeli képződmények az Austin kréta, a Niobrara és Frio-Vicksburg Alaszkai formációk és mások.

A konszolidált tárolók szeizmikus leképezéséhez szükséges követelmények rövid összefoglalásából kiindulva eljutunk a kapott képek értelmezésének, kalibrációjának és használatának módszertanáig. A térbelileg sűrű szeizmikus sebességmérések és a sebesség léptékű színes inverz szelvények hatékonyságát a sikeres példák egész sora bizonyítja. Ezek a megközelítések és a kapott pozitív eredmények azt sugallják, hogy a konszolidált rezervoárok vizsgálatában alkalmazott szeizmikus módszerek a jövőben még meghatározóbbak lesznek, mind a kutatási, mind a termelési alkalmazásokban.

## СЕЙСМИЧЕСКОЕ ОТОБРАЖЕНИЕ ПОРИСТОСТИ И СОДЕРЖАНИЯ УГЛЕВОДОРОДОВ КОНСОЛИДИРОВАННЫХ ФОРМАЦИЙ

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Под консолидированной резервуарной формацией понимаются такие формации значительной мощности, которые обладают таким же или высшим значением акустического импеданса (произведение плотности и скорости), чем глинистые сланцы такого же возраста при нормальном давлении. Для таких формаций пористость и наличие углеводородов можно эффективно отобразить по изменениям сейсмической скорости для значительного интервала глубины (до 4800 м). Приводятся примеры песчанных и карбонатных резервуаров из Восточного Техаса, Оклахомы и участка Северного моря. Записи этих резервуаров содержат формации Аустинский мел, Ниобрара и Фрио-Виксбург Аляска, и др.

Помимо краткого описания необходимых условий для выполнения сейсмического отображения консолидированных резервуаров, статья также занимается и интерпретацией, калибрацией и применением полученных отображений. Эффективность применения густой сети измерения скорости сейсмических волн и изображения в масштабе скорости цветных обратных разрезов доказываются целым рядом удачных примеров. Судя по положительным результатам, примененные сейсмические методы в исследовании консолидированных резервуаров в будущем станут еще более определяющими как и в разведочных, так и в добывающих областях.

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# 'PULL UP' AND 'PUSH DOWN' EFFECTS IN SEISMIC REFLECTION: A USEFUL CONSTRAINT

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'Pull up' and 'push down' effects are commonly observed in seismic sections. These fictitious deformations of seismic reflections under local velocity anomalies can, if they are not identified, lead to misinterpretation of seismic sections. However, in some cases and using simplifying assumptions, these effects can provide a useful constraint in the estimation of the seismic velocities and therefore help interpretation in complex areas. Two examples are presented: the first displays a distinct 'pull up' effect related to the high seismic velocities associated with a salt diapir, the second shows a significant 'push down' effect related to the low velocity of sedimentary basins.

**Keywords: seismic reflection, velocity anomalies, pull up, push down**

## 1. Introduction

The 'pull up' and 'push down' effects, commonly observed in seismic sections, engender fictitious deformation of seismic reflections beneath a local anomaly of high or low velocity (*Fig. 1*). Such an anomaly may be related, for example, to salt diapirism, coral reefs, volcanic intrusions or young sedimentary basins, and has been known since the earlier development of reflection seismology. 'Pull up' and 'push down' effects can, usually, be recognized when they affect distinct and relatively flat underlying reflectors. The shape of the reflections in the seismic section follows the geometry of the high or low velocity anomalous body. A good knowledge of the seismic velocities is important in correcting such effects on the depth sections.

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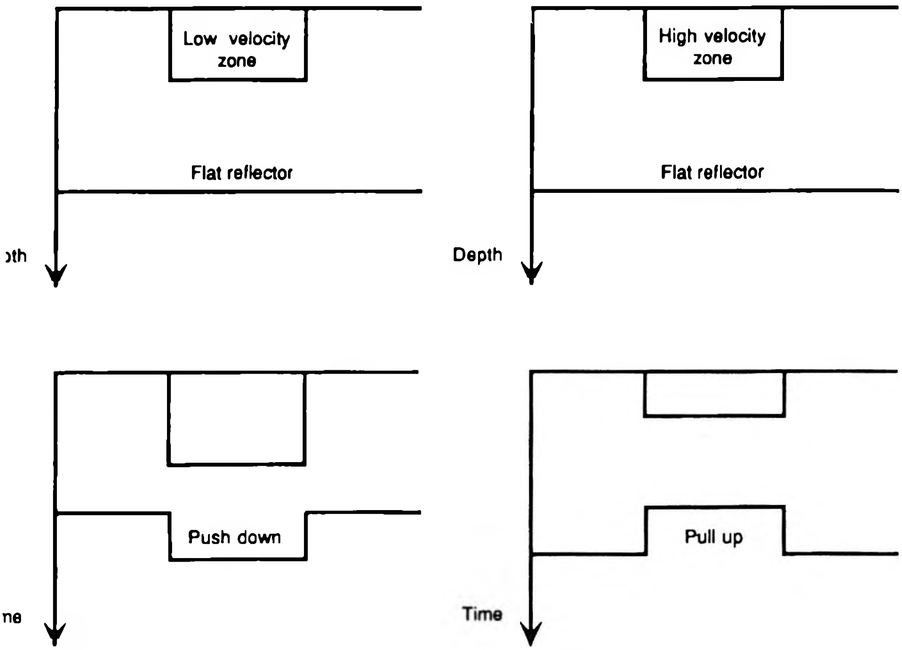



*Fig. 1.* 'Pull up' and 'push down' effects: a fictitious deformation of seismic reflectors under a local anomaly of velocity

*1. ábra.* Felboltozódási („pull up") és bemélyedési („push down") hatások: a szeizmikus reflektáló felületek helyi sebességanomáliák alatt létrejövő látszólagos deformációi

*Рис. 1.* Своды („pull up") и впадины („push down"): кажущиеся деформации сейсмических отражений в связи с локальными скоростными аномалиями

Nevertheless, 'pull up' and 'push down' effects are often considered as an inconvenience which can affect the interpretation of the seismic sections. In salt diapiric zones, for example, the seismic velocities are usually poorly defined and these effects, if not efficiently corrected on the depth sections, can be interpreted as geological features.

This paper intends to show that, in some cases and under simplifying assumptions, 'pull up' and 'push down' effects can provide a useful constraint in estimating the seismic velocities and, thus, help in interpreting complex areas. We present a simple method to perform this kind of analysis, and apply it to two examples. The first, taken from conventional exploration profiles, shows a study over a salt diapir. The second, from deep seismic reflection data, is used to obtain an estimation of the seismic velocity in the upper crust.

## 2. Modelling of 'pull up' and 'push down' effects: an approach to velocities

The conventional method used to interpret seismic reflection sections is based on two kinds of information: the time section itself, improved by more or less sophisticated processes, and the information on velocity, usually provided by the velocity analysis technique on multichannel experiments. On the basis of this information, one can migrate the section if required by the slope of the reflections and, finally, depth convert it. Nevertheless, problems may be encountered in the application of this method, particularly if the seismic velocities are poorly determined. The velocity analysis technique is not effective for areas which include strong velocity contrasts, numerous scatterers, dipping reflectors and/or complex geometrical patterns. Also, this technique cannot be employed for deep reflections because of the weak sensitivity of the CDP (common-depth-point) hyperbolae to the seismic velocity.

We propose a different way to help the interpretation of seismic sections in such difficult areas. This method, which is quite rough and can only provide an approximation of the seismic velocities, is based on the observation of 'pull up' and/or 'push down' effects on underlying reflections. Such 'pull up' or 'push down' effects, related to high or low velocity anomalies, are often observed in areas where velocity analyses are not suitable. The method consists of an a priori assumption of the geometry of the underlying reflectors, from geological evidence and/or analysis of the data which are not affected by the 'pull up' or 'push down' effect. Such an a priori assumption leads to the calculation, at several locations, of estimated velocities in the anomalous body using the classical relation

*Interval velocity* = 2 * *Thickness / Interval vertical time*

Such a formulation is valid only in the case of a relatively simple a priori assumption on the geometry of the anomalous body and underlying reflectors. A more accurate formulation, which is not required in the following examples, should take into account the effect of more complex geometry on the ray paths. The geological plausibility of the resulting velocities governs the acceptance or rejection of the a priori assumption. Realistic velocities a posteriori confirm the validity of the model, while unrealistic velocities lead to its rejection.

In order to check the applicability of such a method, we analysed the sensitivity of the result (the velocity) to variations of the input parameters (the thickness). A too high sensitivity means a weak probability of obtaining reasonable velocity values, even with a correct geometrical model; a too low sensitivity could lead to plausible velocity values, even with a false model. We included uncertainties on the interval time measurements in the calculations. In the following examples, we obtain a 30% variation (1.2 km/s) for the interval velocity in the diapir assuming a 20% variation (0.4 km) for the thickness of the diapir. We also obtain a 23% variation (1.38 km/s) for the velocity of the upper crust assuming a 20% variation (5 km) for the lower crustal reflectors. The calculations suppose an uncertainty on the time measurement which reaches 0.05 s. Thus, the sensitivity analysis shows that significant variations of the geometrical model lead to departures of the velocities beyond the range of plausible values, and demonstrates the applicability of the method. The two examples mentioned above present applications of this method to a 'pull up' effect observed beneath a salt diapir and 'push down' effects observed beneath sedimentary basins.

## 3. Lateral variations of the seismic velocity in a salt diapir

We have used well data and conventional seismic profiles to study a 20 km$^2$ area in the southern Rhine Graben (France), at the northern border of the Mulhouse potassic salt mine district. These data (*Fig. 2*) show a thick Tertiary sequence underlain by Jurassic sediments. A Bajocian-Bathonian oolithic limestone, known as the 'Grande Oolithe', is particularly clear from the seismic data (*Fig. 3*) and closely underlies a major discordance, corresponding to Upper Jurassic and Cretaceous [MDPA 1983]. This discordance, known from the well data, is not associated with a pronounced angular discontinuity and probably corresponds to a hiatus of deposition rather than to a tectonically related erosion. The evaporitic layers were deposited at Sannoisian and are related to the Oligocene subsidence of the Rhine Graben [BLANC-VALLERON, GANNAT 1985]. This subsidence continued during the Upper Oligocene, leading to the deposition of detritic Stampian and Chattian sediments. Diapirism occurred during and after the late Oligocene and roughly followed the N–S orientation of the Rhine Graben normal faults [LARROQUE, ANSART 1985].

Figure 3 displays a seismic section in this area. The strong reflection (noted '*O*') close to the base of the diapir corresponds to the top of the 'Grande Oolithe' layer, as evidenced by the well data. The other marked reflection ('*T*') represents the top of the diapir. The complex seismic pattern observed in the diapir is a typical characteristic of salt diapirs [see for example NELY 1980, JENYON 1986]. We note that reflection *O* seems to reflect the shape of the diapir itself. A similar observation can be made from other profiles across the diapir. We suggest that the observed shape of reflection *O* is related to a 'pull up' effect due to the high seismic velocity in the salt.

Using velocity analyses at several locations along the profiles we calculated, using the DIX [1955] formula, the approximate interval velocities of the near surface layer (from surface to reflector *T*) and of the diapir (between reflectors *T* and *O*) at each location. The resulting interval velocities of the diapir range between 3.0 and 6.0 km/s. We find no relation between these velocities and the thickness of the diapir (*Fig. 4a*). This large dispersion of the interval velocities may be related to the poor suitability of the classical velocity

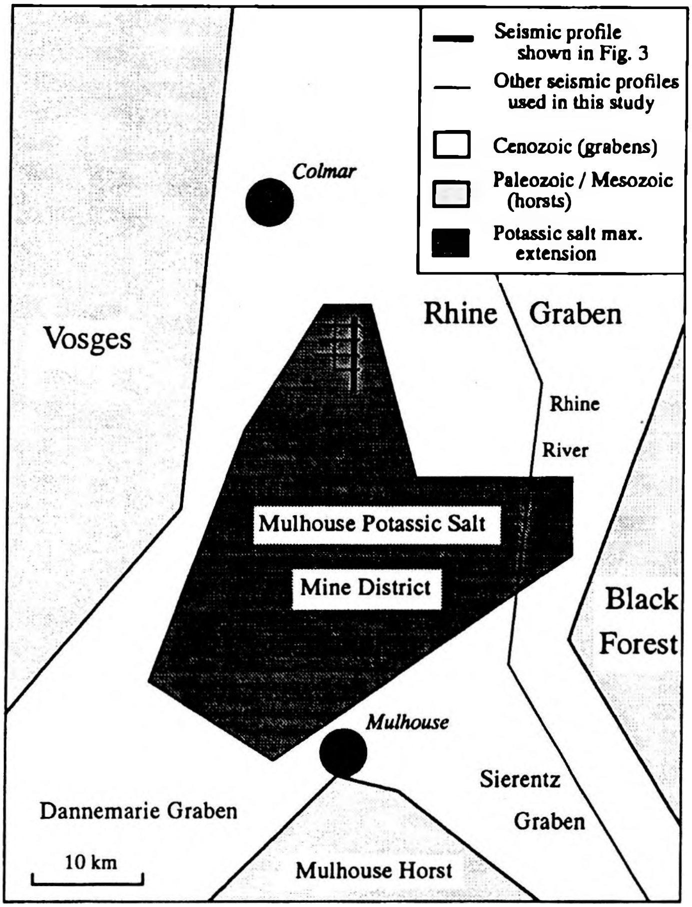



*Fig. 2.* Location of the area studied on the northern border of the Mulhouse potassic district, southern Rhine Graben, France. The profile displayed in Fig. 3 is underlined

*2. ábra.* A vizsgált terület elhelyezkedése a franciaországi Rajna-árok déli részén található Mulhouse kálisóbánya északi határán. A 3. ábrán látható szelvényt vastag vonal jelzi

*Рис. 2.* Исследуемый участок расположен на северной границе месторождения калиевой соли Мюлхоус в южной части Рейнского грабена во Франции. Жирная линия показывает расположение профиля, изображенного на Рис. 3

analysis in diapiric areas, due to strong velocity contrasts, large dips and numerous scatterers in the salt [CORDIER 1983].

Thus, we have tried to use the 'pull up' effect observed on reflection *O* to determine the interval velocities of the diapiric sequence. We assumed that the previously found interval velocities of the near surface sequence are good and used them to calculate the depth of reflector *T*. Next, we assumed that reflector *O* is roughly flat (*Fig. 5*) and we used this hypothetical depth section and the observed time section to compute the interval velocities in the diapir.

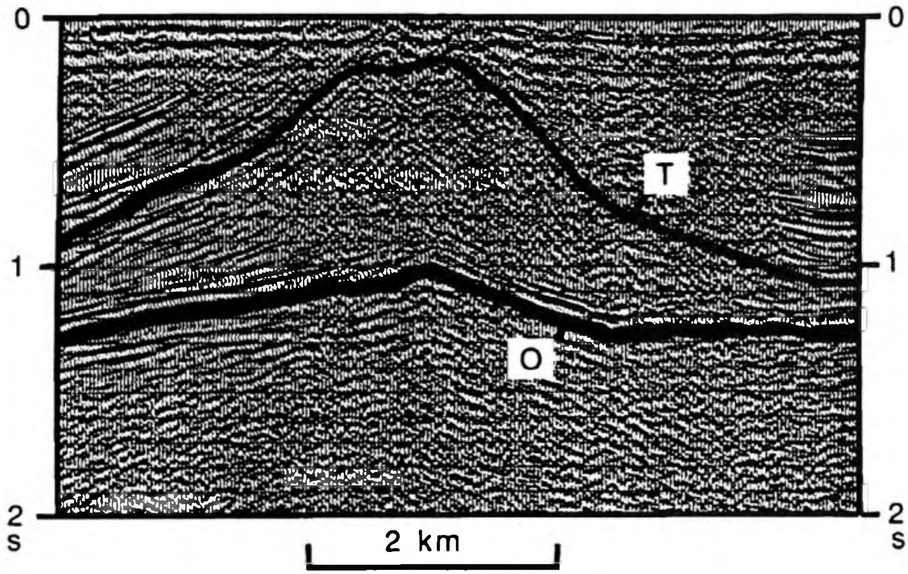



*Fig. 3.* Typical example of 'pull up' beneath a salt diapir observed in the southern Rhine Graben, France; *T* is the top of the diapir, *O* is the reflection on the 'Grande Oolithe' limestone, which marks the base of the diapir

*3. ábra.* Tipikus példa só diapír alatti látszólagos felboltozódásra („pull up") a Rajna-árok déli részéről, Franciaországból; *T* a diapír teteje, *O* a „Grande Oolithe" mészkőről érkező reflexió, amely a diapír alját jelöli

*Рис. 3.* Типичный пример кажущегося свода под соляным диапиром в южной части Рейнского грабена (Франция). *T* кровля диапира, *O* отражения с известняковой формации „Гран Оолит", указывающие подошву диапира

Several observations support the hypothesis of a roughly flat reflector *O*. First, the 'Grande Oolithe' limestone is a very compact layer and cannot be deformed easily. Thus, we expect that it will undergo brittle deformation and faulting rather than form a fold as indicated by the seismic section if the 'pull up' effect is not assumed. In addition, such a fold is not likely in the Rhine Graben considering the overall extension in this area. The flatness of reflector *O* seems consistent with the available geological data. In the areas around the diapir, reflector *O* appears roughly flat. Based on a well and some velocity analysis from non diapiric zones, we determined its depth to be 2075 m.

Using $D_O$ as the depth of reflector *O* (2075 m), $D_T$ as the depth of reflector *T* determined from the velocity analyses, $t_O$ as the two way travel time from surface to reflection *O*, and $t_T$ as the two way travel time from surface to reflection *T*, we obtain the relation

$$V_{T\text{-}O} = 2\,(D_O - D_T)\,/(t_O - t_T)$$

where $V_{T-O}$ is the interval velocity in the diapir. This relation was used to calculate an alternative interval velocity at each location.

Such calculation provides interval velocities in the diapir which range between 2.5 and 5.0 km/s. These values are, most probably, more reasonable than the previous values using conventional analysis. Salt velocities usually range between 4.0 and 5.0 km/s [CORDIER 1983]. The diapir consists of salt, but probably also includes marl and clay which have lower seismic velocities. Moreover, when the interval velocities versus the double travel time between reflections *T* and *O* are plotted, a linear relation is obtained (*Fig. 4b*). This velocity function indicates higher seismic velocities in the thick part of the diapir than in the flanks. The averaged velocity in the thickest parts of the diapir is 4.0 km/s, which is almost the pure salt velocity, 3.5 km/s in the flanks, and only 3.0 km/s in the thinnest parts of the salt layers. This result is consistent with the classical models of salt migration [e.g. JENYON 1986]. Layers mainly

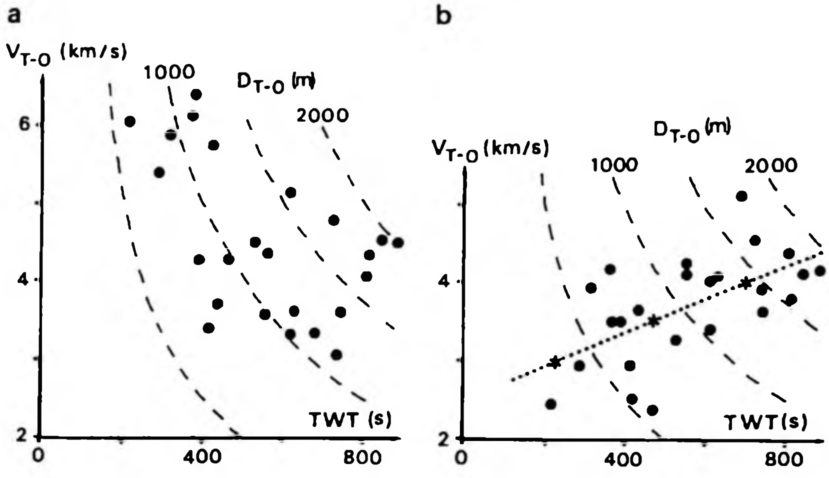



*Fig. 4.* Velocity versus two way travel time in the diapiric sequence: a) from velocity analysis; b) using the 'pull up' effect. TWT is the two way travel time between *T* and *O*, $V_{T-O}$ the interval velocity of the sequence *T–O*, and $D_{T-O}$ the thickness of this sequence. A lateral velocity function can only be derived from b), due to the important dispersion in a)

*4. ábra.* A kétszeres menetidő-sebesség függvény a diapír rétegsorban: a) sebességanalízisből, b) a „pull up" jelenség felhasználásával. TWT a *T* és az *O* közötti kétszeres terjedési idő, $V_{T-O}$ a *T–O* rétegsor intervallumsebessége, a $D_{T-O}$ pedig a rétegsor vastagsága. Laterális sebességfüggvényt csak a b) alapján tudunk számítani, mivel az a)-n igen nagy a szórás

*Рис. 4.* Функция двойное время-скорость внутри диапира а) по анализу скоростей, b) при использовании явления „pull up". TWT двойное время хода между поверхностями *T* и *O*, $V_{T-O}$ скорость интервала *T-O*, $D_{T-O}$ мощность интервала *T-O*. Латеральная скоростная функция была определена лишь в случае b), в связи с большой дисперсией данных в случае а)

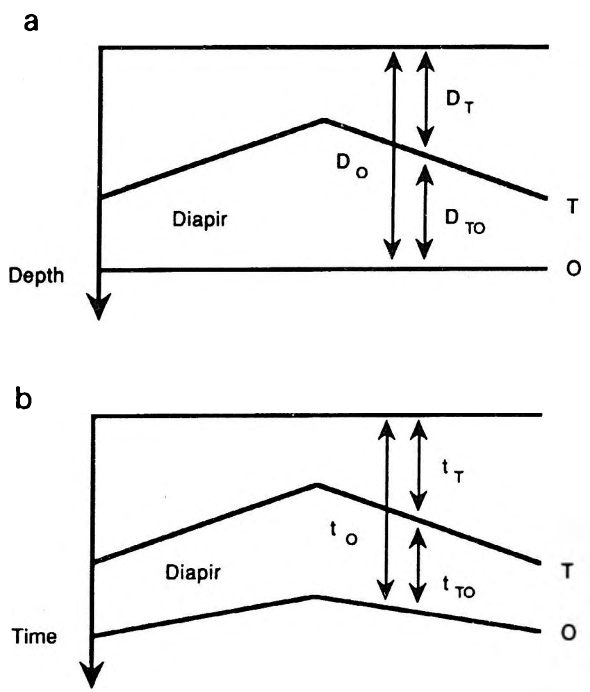



*Fig. 5.* a) A priori assumption about the reflector geometry. Reflector *T* is depth converted using velocities given by velocity analysis. Reflector *O* is assumed flat from geological evidence; b) corresponding seismic section

*5. ábra.* a) A reflektor geometriájára vonatkozó a priori feltételezés. A *T* reflektor mélység-konvertálását a sebességanalizisből számolt sebeségekkel végeztük. Az *O* reflektort földtani megfontolások alapján síknak vettük; b) a megfelelő szeizmikus szelvény

*Рис. 5.* a) A priori предположение о геометрии отражающего горизонта. Определение горизонта *T* выполнено по скоростям, определенным анализом скорости. Горизонт *O* считается горизонтальным по геологическим соображениям; b) соответствующий сейсмический разрез

composed of massive and pure salt flow laterally and accumulate in dome structures. On the contrary, layers composed of melted salt, clay and/or marl remain in the flanks. Consequently, the seismic velocities in the diapir reach the velocity of the pure salt while the flanks show lower velocities.

## 4. Estimation of the seismic velocities in the crust

A second example demonstrating the use of 'pull up' and 'push down' effects is drawn from deep seismic reflection data. The SWAT (South Western Approaches Traverse) profiles were acquired in the Celtic Sea-Western Channel area by the BIRPS and ECORS groups in 1983 [BIRPS-ECORS 1986].

The 15 seconds (TWT) records allow the examination of the entire crust in this area [e.g. DYMENT 1989]. Two important Mesozoic and Cenozoic sedimentary basins, the deep North Celtic Sea Basin and the shallower South Celtic Sea Basin, can be observed on the sections of SWAT 2 to 5 profiles (*Fig. 6*). Strong dipping reflectors interpreted as Variscan thrusts [BIRPS-ECORS 1986] cross the nearly transparent upper crust. As generally observed in Western Europe, this upper crust is underlain by a very reflective lower crust [MEISSNER, WEVER 1986; MATTHEWS, CHEADLE 1986; BOIS et al. 1987]. Wide-angle seismic experiments in neighbouring areas [MOONEY, BROCHER 1987] determine the Moho to be at the base of the reflective unit. In contrast, the upper mantle is nearly transparent.

A problem encountered in deep seismic reflection studies is that of determining the crustal velocities. Velocity analysis is usually effective in the sedimentary section only, but cannot be used to study the velocity distribution of the deeper parts of the crust because of the flatness of the CDP hyperbolae. Wide-angle experiments (for example ESP) are most probably the best way to calculate accurate velocities in the deep crust, despite possible biases related to seismic anisotropy and inhomogeneous plane wave nature. Nevertheless, such data are not available in the Celtic Sea area. To overcome this problem, BANO

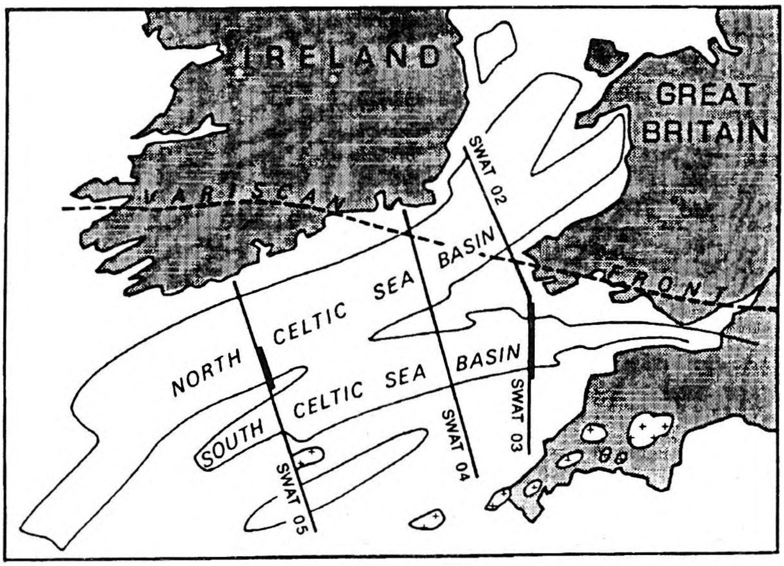



*Fig. 6.* Location of the SWAT profiles in the Celtic Sea area. The Variscan Front, North and South Celtic Sea Basins and Cornwall batholiths are also shown. Parts of the profiles displayed in Figs. 8 and 9 are underlined

*6. ábra.* A SWAT szelvények elhelyezkedése a Kelta-tenger területén. A variszkuszi frontot, az Észak- és Dél-Kelta-tenger medencéit és a Cornwall batolitokat szintén feltüntettük. A 8. és 9. ábrákon bemutatott szelvényrészletek helyét vastag vonallal jelöltük

*Рис. 6.* Расположение профилей SWAT на участке Келтийского моря. На карте также приведены расположения варисского фронта, бассейны Северо- и Южно-Келтийских морей и батолитов Корнуалла. Жирными линиями отмечены интервалы профилей,

[1989] and DYMENT, BANO [1991] modelled diffraction hyperbolae of short flat reflectors in complex parts of the lower crust. Having applied this method to a part of the SWAT 5 profile, they explain the complex patterns of reflections observed on this section. They also obtain, adjusting the RMS velocity to the observed hyperbolae, a crustal RMS velocity estimation of 6 km/s. We propose here another approach to obtain upper crustal velocities using the 'push down' effect of sedimentary basins.

A characteristic feature observed on the SWAT data is the rough flatness and relatively constant thickness of the reflective lower crust [DYMENT 1989; DYMENT et al. 1990; SIBUET et al. 1990]. The constant thickness appears to characterize the whole area. The flatness is particularly evidenced in areas where the sedimentary basins are not present. WARNER [1987] suggested that no velocity 'push down' is observed beneath the deep sedimentary basins. He explained this observation by the opposing effects of local isostasy and velocity: the velocity 'push down' would be compensated by an isostatic uplift of the Moho. However, careful analysis of the Moho and lower crustal reflections beneath the sedimentary basins reveals more complex patterns.

Bano's automatic extraction of reflections [BANO et al. 1988; BANO 1989] was applied to the SWAT 2 to 5 profiles (Figs. 8 and 9), to increase the quality of the analysed sections. Undulations of the Moho and lower crust reflections were evidenced on the resulting time sections, with a marked correlation with the younger, mainly Cretaceous and Cenozoic, sedimentary basins. The similar shape of these seismic reflections suggests a 'push down' effect of the low velocity sedimentary basins on the presumably flat underlying lower crustal reflections. Velocity analyses confirm that the velocities of the sedimentary basins are lower than those of the surrounding areas.

In order to verify the hypothesis of a 'push down' effect of the sedimentary basins on a physically flat lower crust (*Fig. 7)* and to estimate upper crustal velocities, we analysed two areas where the lower crust reflections are quite continuous and their undulations well marked. The first (*Fig. 8*), from the SWAT 5 profile, shows a narrow sub-basin of the North Celtic Sea Basin (Fig. 6), mainly filled by Cretaceous sediments [DYMENT 1989]. The other (*Fig. 9*), from the SWAT 3 profile, shows the South Celtic Sea Basin-Bristol Channel Basin junction (Fig. 6).

In both examples, we marked some strong and continuous crustal reflections (Figs. 8 and 9). The sedimentary velocities are determined from the velocity analysis. Using, at location *i*, $D_S{}^i$ as the depth of the sediments, $t_S{}^i$ as the two way travel time to the basement and $t_C{}^i$ as the two way travel time to an assumed flat crustal reflection, we obtain the relation

$$V_C{}^{ij} = 2\,(D_S{}^i - D_S{}^j)\,/[(t_S{}^i - t_S{}^j) - (t_C{}^i - t_C{}^j)]$$

where $V_C{}^{ij}$ is the velocity of the upper crust between the top of the basement and the maximum depth of the sedimentary basin $D_S^{max}$, considered to be uniform between locations *i* and *j*.

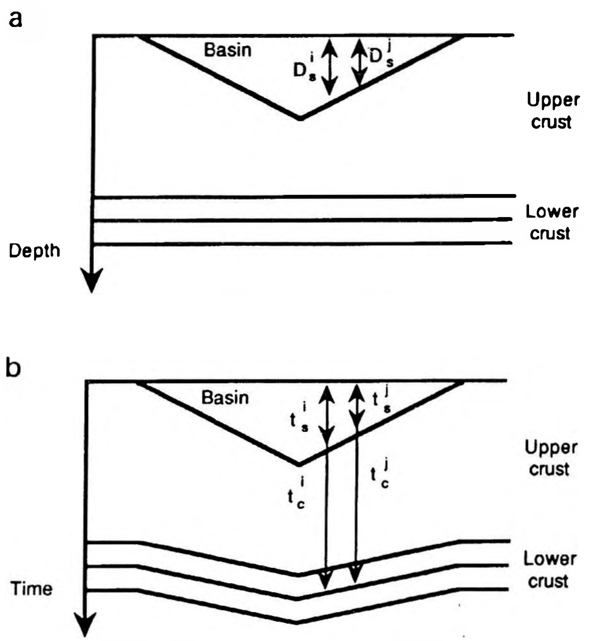


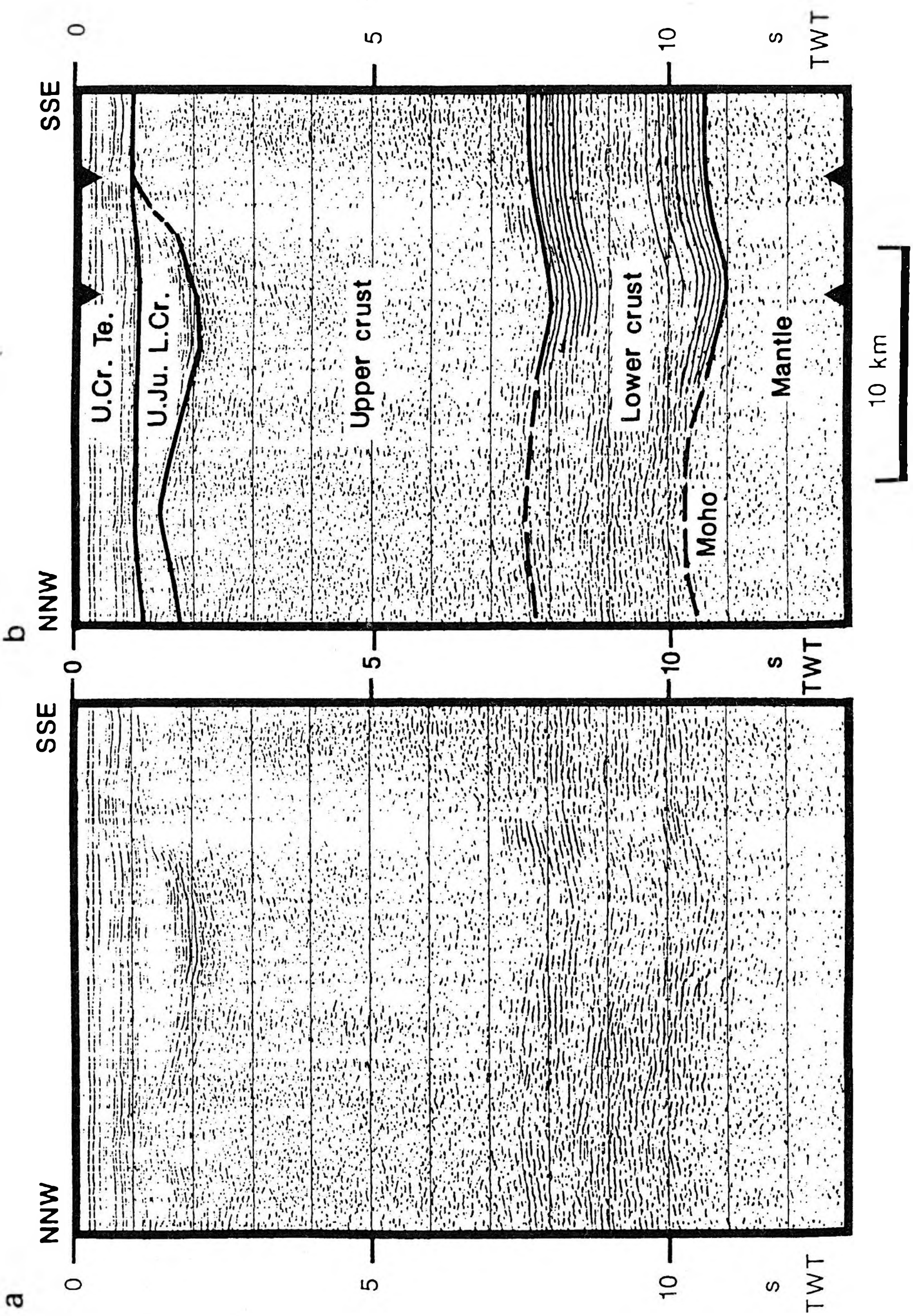

*Fig. 7.* a) A priori assumption about the reflector geometry. The basin is depth converted using velocity analysis. The lower crustal reflectors are assumed flat (see text); b) corresponding seismic section

*7. ábra.* a) A reflektor geometriájára vonatkozó a priori feltételezés. A medence mélység-konvertálását a sebességanalízis alapján végeztük. Az alsókéregbeli reflektáló felületeket vízszintesnek vettük (lásd a szöveges részt); b) a megfelelő szeizmikus szelvény

*Рис. 7.* a) A priori предположение о геометрии отражающего горизонта. Определение глубины бассейна выполнено по данным анализа скоростей. Отражающие горизонты, расположенные в нижней коре, считаются горизонтальными (см. текстовую часть); b) соответствующий сейсмический разрез

The main limitation of this method is its relatively high sensitivity to the measured times. The differences ($t_s^i - t_s^j$) and ($t_c^i - t_c^j$) are usually quite small and represent only a few times the uncertainty in their measurement. The uncertainty in the velocity can reach 23 per cent of its value. In order to reduce the inherent uncertainty, we applied the method to several different locations and reflections. We obtained an average upper crustal velocity of 6.16 km/s on SWAT 5, and 6.20 km/s on SWAT 3. The standard deviation of the computed values is only 0.2 km/s.

The estimates of the upper crustal velocities that we derived are in good agreement with the one found by BANO [1989] and DYMENT and BANO [1991] from diffraction modelling on the southern part of the SWAT 5 profile. In neighbouring areas, refraction and ESP experiments provided crustal velocities of 6.2 km/s in the North Sea [KLEMPERER 1988] and 5.9 to 6.2 km/s for the upper crust in the Bay of Biscay [PINET et al. 1987]. The values determined using the 'push down' effect are consistent with these results and confirm in some way the assumption of flat geometry for the lower crustal reflections and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Moho
Mantle
10 km

the Moho under the Celtic Sea basins. This result may have important consequences on the still unknown nature and origin of the lower crustal reflectivity, and may help to discriminate among the numerous existing models [e.g. DYMENT, BANO 1991].

## 5. Conclusion

In some cases and under simple assumptions, 'pull up' and 'push down' effects can provide a useful constraint to the knowledge of seismic velocities and the interpretation of seismic sections in complex areas, as illustrated by the previous examples. The poor suitability of velocity analysis in complex areas such as diapirs, reefs or volcanic intrusions makes difficult the determination of accurate velocities and thus the depth conversion of time sections in these areas. A simple a priori assumption about the geometry of underlying reflectors is used as a basis for calculating interval velocities in the anomalous body. The geological plausibility of the computed velocities a posteriori confirms or negates the assumption about the geometry.

Application of this method to the first example gives additional information on the velocities in a salt diapir, and thus on the salt migration processes. The second example provides an estimate of the upper crustal velocity in the Celtic Sea area, and confirms in some way the flat character of the lower crustal reflectors in this area, which is related to the still controversial nature and origin of the lower crustal reflectivity. Such results demonstrate the applicability of the method despite (and perhaps because of) its quite simplistic background.

*Fig. 8.* Part of the deep seismic reflection profile SWAT 5, showing the reflective lower crust affected by 'push down' effect of the southernmost part of the North Celtic Sea basin. a) the data after automatic extraction of reflections; b) the same, interpreted. Crustal velocities are calculated for the underlined lower crustal reflections at the locations marked by black triangles. U and L for upper and lower, Tr, Ju, Cr and Te for Triassic, Jurassic, Cretaceous and Tertiary

8. ábra. A SWAT 5 mély szeizmikus reflexiós szelvény egy részlete, amely az Észak-Kelta-tenger medencéjének legdélibb része okozta „push down" jelenség által érintett alsó kéregbeli reflektáló felületet mutatja. a) az adatok az automatikus reflexiókiemelési művelet után; b) ugyanez értelmezve. A kiemelt kéregbeli reflexiókra vonatkozó sebességeket a fekete háromszögekkel jelölt helyeken számították. Az U felsőt, az L alsót, a Tr, Ju, Cr, és Te pedig triászt, jurát, krétát és negyedidőszakot jelöl

*Рис. 8.* Интервал глубинного сейсмического разреза SWAT 5, указывающий отражающий горизонт в нижней коре, затронутый влиянием „push down" от южной части бассейна Северо-Келтийского моря. а) данные после автоматического выделения отражений, b) то же самое с интерпретацией. Расчет скоростей для выделенных горизонтов выполнен по точкам, обазначенным черными треугольниками. U означает верхний, L нижний, а Tr, Ju, Cr, и Te триасовый, юрский, меловой и третичный возраст соответственно

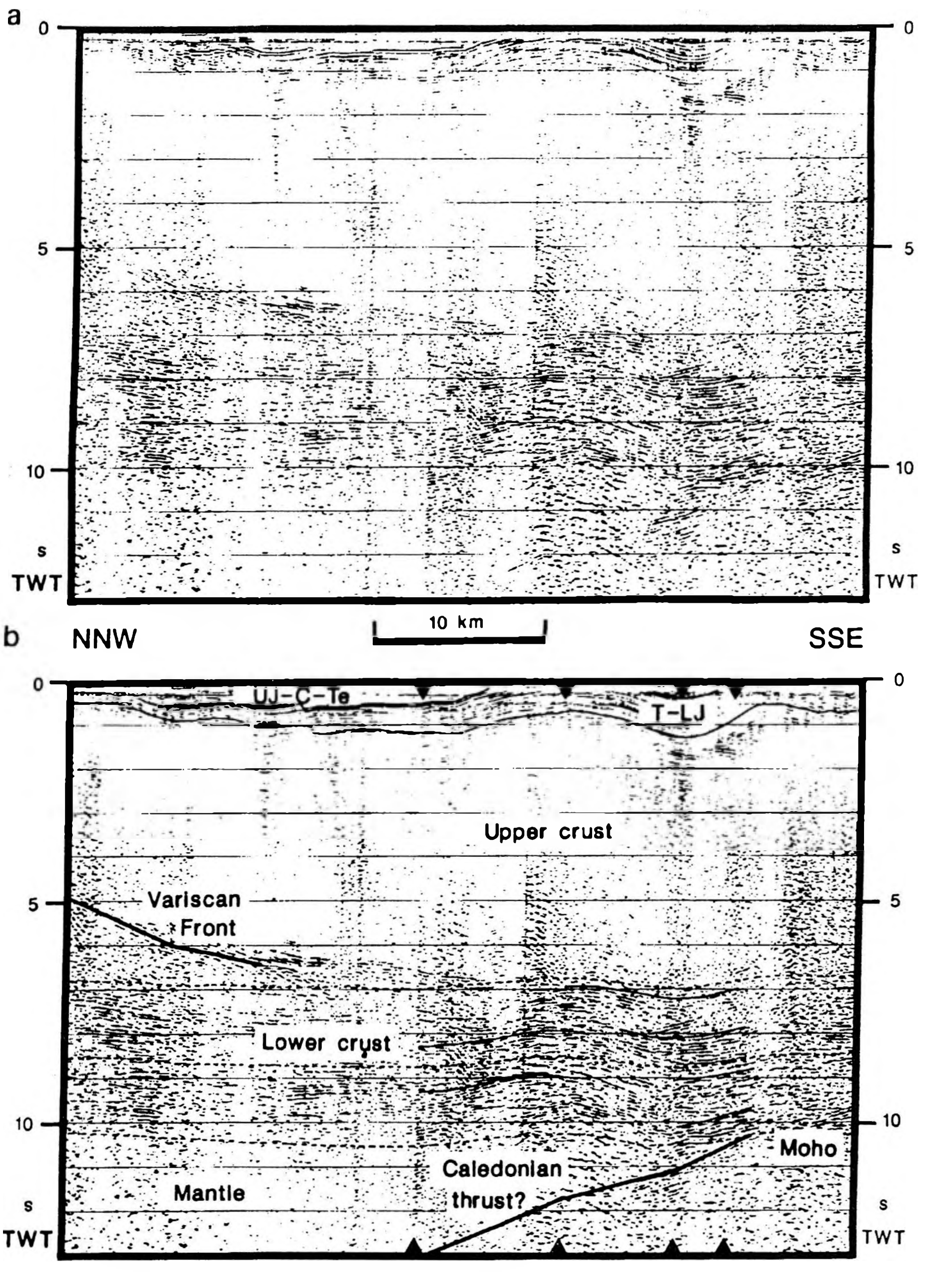
a
0
5
10
s
TWT
b
NNW
10 km
SSE
UJ-C-Te
T-LJ
Upper crust
Variscan
Front
Lower crust
Mantle
Caledonian
thrust?
Moho

*Fig. 9.* Part of the SWAT 3 profile, showing the reflective lower crust affected by 'push down' effect of the South Celtic Sea and Bristol Channel basins. a), b) see Fig. 8. Dotted lines are time converted reflections of flat reflectors, assuming a crustal velocity of 6.2 km/s. Note the 'push down' effect on the Variscan Front and on a presumed Caledonian thrust

*9. ábra.* A SWAT 3 szelvény egy része; az alsó kéregbeli reflektáló felületen a Dél-Kelta-tenger és a Bristol csatorna medencéi által keltett „push down" jelenség látszik. a) és b) mint a 8. ábrán. A pontozott vonalak sík reflektáló felületek idõkonvertált reflexiói, 6,2 km/s-os kéregbeli sebességet feltételezve. Figyeljük meg a „push down" hatást a variszkuszi fronton és a feltételezett kaledoniai törésen

*Рис. 9.* Интервал профиля SWAT 3; по отражающему горизонту в нижней коре отмечается влияние „push down", вызванное бассейнами Южно-Келтийского моря и Бристольского канала. a) и b) как на Рис. 8. Пунктиром обозначены пересчитанные на время отражения от горизонтальных отражающих горизонтов, при предположении скорости 6.2 км/с внутри коры.Обращается внимание на влияние „push down" на варисском фронте и на предпологаемом каледонском разломе

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## LÁTSZÓLAGOS FELBOLTOZÓDÁSOK („PULL UP") ÉS BEMÉLYEDÉSEK („PUSH DOWN") A SZEIZMIKUS REFLEXIÓKBAN— SEGÍTSÉG AZ ÉRTELMEZÉSBEN

Jérôme DYMENT és Maksim BANO

A szeizmikus időszelvényeken gyakran megfigyelhetünk felboltozódásokat és bemélyedéseket. A szeizmikus reflexióknak ezek a helyi sebességanomáliák alatt létrejövő látszólagos deformá-

ciói, ha nem azonosítjuk őket, akkor a szeizmikus szelvény félreértelmezéséhez vezethetnek. Mindamellett bizonyos esetekben, egyszerűsítő feltevéseket használva ezek a hatások hasznos megszorítást jelenthetnek a szeizmikus sebességekre vonatkozóan és így segíthetnek komplex területek értelmezésénél. Két példát mutatunk be: az első egy nagy szeizmikus sebességű só diapírhez kapcsolódó határozott látszólagos felboltozódási hatást mutat be, a második egy kissebességű üledékes medencéhez kötődő látszólagos bemélyedést ábrázol.

## КАЖУЩИЕСЯ СВОДЫ („PULL UP") И ВПАДИНЫ („PUSH DOWN") СЕЙСМИЧЕСКИХ ОТРАЖЕНИЙ: ПОЛЕЗНЫЕ НЕДОСТАТКИ

Жером ДИМЕН и Максим БАНО

На сейсмических временных разрезах часто наблюдаются своды и впадины. Кажущиеся деформации сейсмических отражений связаны с локальными скоростными аномалиями и могут привести к ложной интерпретации разреза, если их характер не выявлен. Тем не менее, подобные явления могут служить полезной информацией о скорости сейсмических волн и способствуют комплексной интерпретации участков. Показаны два примера: на первом наблюдается кажущийся свод, связанный с высокоскоростной соляной диапировой структурой, а на — втором кажущаяся впадина, связанная с низкоскоростной аномалией осадочного бассейна.

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# NEOGENE STRUCTURAL EVOLUTION AND BASIN OPENING IN THE WESTERN CARPATHIANS*

M. KOVÁČ**, F. MARKO** and M. NEMČOK***

The Oligocene-Lower Miocene transpressive stage of the Neo-Alpine development governed the tectonic escape of the Carpathians from the Eastern Alpine domain and opening of the W-E trending sedimentary area in the western part of the Central Zone of the Western Carpathians. Clockwise rotation of the maximum principal stress from N to NE caused changes in the structural pattern well observable in the Little Carpathians and formed the present-day Vienna and Danube Basins during the Middle Miocene. Upper Miocene and Pliocene basin evolution points to extensional conditions.

**Keywords: Western Carpathians, Little Carpathians, Vienna Basin, Danube Basin, Neogene, paleogeography, structural evolution, basin evolution**

## 1. Introduction

Neo-Alpine structural evolution of the western area of the Carpathians was governed by a collision of continent-continent type, between platform and the overriding Alpine-Carpathian system. The present configuration of the suture zone in the Alpine-Carpathian junction (*Fig. 1*) suggests that relative movement of colliding plates was oblique and had a notable effect on the bending of the Carpathians [VASS et al. 1988].

The ending of overthrust movements in the frontal part of the orogene from W to E during the Neogene [BUDAY et al. 1965, JIŘÍČEK 1979] corresponds well to paleostress field changes in the western part of the Central Carpathians. The maximum principal stress rotated here from the Oligocene-

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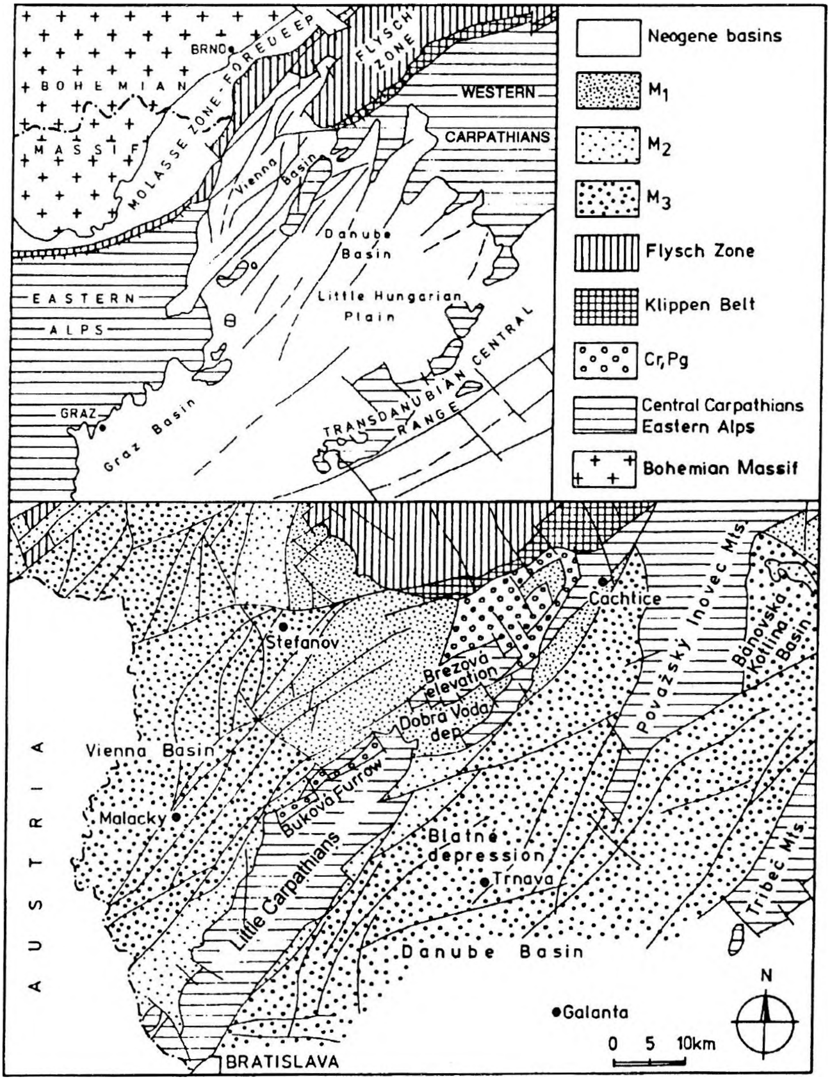


*Fig. 1.* Geological structure of the Alpine-Carpathian-Pannonian junction
$M_1$—Lower Miocene sediments; $M_2$—Middle Miocene sediments; $M_3$—Upper Miocene sediments; Cr, Pg—Cretaceous and Paleogene Units of the Central Zone of the Western Carpathians

*1. ábra.* Az alpi-kárpáti-pannon csatlakozás geológiai szerkezete
$M_1$—alsó miocén üledékek; $M_2$—középső miocén üledékek; $M_3$—felső miocén üledékek; Cr, Pg—a Nyugati-Kárpátok középső zónájának kréta és paleogén egységei

*Рис. 1.* Геологическая структура стыковки Альпийского, Карпатского и Паннонского региона
$M_1$ —осадочные образования нижнемиоценового возраста, $M_2$—осадочные образования среднемиоценового возраста, $M_3$—осадочные образования верхнемиоценового возраста, Cr, Pg—меловые и палеогеновые образования средней зоны Западных Карпат

Lower Miocene to the Upper Miocene from NW-SE to NE-SW. The Oligocene-Lower Miocene orientation of the paleostress field may be associated with paleomagnetically determined anti-clockwise rotation of about 40° in the Flysch Zone and partly also with internal fault block rotation in the Central Zone of the Western Carpathians, the maximum of which was dated at the Lower-Middle Miocene boundary [KRS et al. 1977, TÚNYI, KOVÁČ 1991].

By means of microfault slip data, macrostructural analysis, sedimentological and ecostratigraphical methods, several stages of Neo-Alpine tectonogenesis and basin evolution have been recognized.

## 2. Tectonic escape of the Western Carpathians from the Eastern Alpine domain (Oligocene-Early Miocene)

The first transpressive stage of development, with strong compression trending approximately to NW-SE (WNW-ESE) in the western area of the Carpathians was active during the Upper Oligocene and Lower Miocene. In this time, the Paleo-Alpine structures were reworked, and even remnants of the Paleogene basin were incorporated into the Neo-Alpine fabric (Fig. 1 and *Fig. 2*).

The tectonic escape of the Central Zone of the Western Carpathians made possible the present-day N-S trending sinistral strike slips with marked amplitude of motion. They can be compared with NE-SW sinistral strike slips in the Alpine domain, which played a dominant role during the whole Neo-Alpine tectonogenesis [NEUBAUER 1988], and they are suggested by Alpine geologists to be logical products of the collision advance to N-NE. In the western part of the Carpathians, synchronal ENE-WSW trending dextral strike slip faults were active, represented by the Myjava shear zone [POSPÍŠIL et al. 1986]. They were accompanied by NE-SW trending reverse faults and back thrusts.

This stage of structural evolution is documented in the northern part of the Little Carpathians by the deformation of the turbidities of the Hrabnik Formation in the Buková Furrow by en-echelon type folds [MARKO et al. 1990]. This Kiscellian part of the sequence consisting of dark claystones and clayey siltstones with several intercalations of graded sandstones contained poor associations of calcareous nannoplankton similar to associations in the Ždánice Unit and differing from those occurring in the Central Zone of the Western Carpathians [BYSTRICKÁ 1979]. Similarly, findings of redeposited calcareous nannoplankton of the NP 24-25 Zone in the Lower Miocene sediments of Dobrá Voda [KOVÁČ et al. 1991] indicate that sedimentation in this area persisted at some places till the end of the Oligocene, in spite of the general uplift of the region, connected with erosion of Cretaceous and Paleogene sequences.

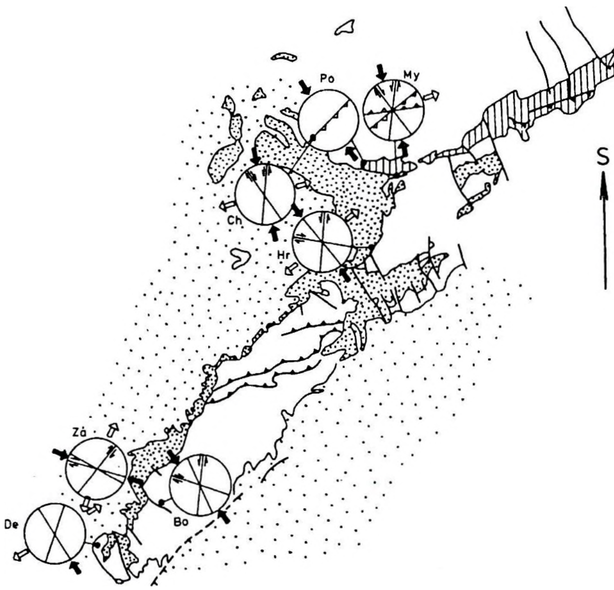


*Fig. 2.* Scheme of maximum compressional stress during the Oligocene-Lower Miocene
Localization of measurements: De—Devín; Zá—Záhorská Bystrica; Bo—Borinka; Ro—Rohožník; So—Sološnica; Ra—Raková; Hr—Hradište; Dc—Dechtice; Me—Melčice; Fa—Fajnorov lom; Ja—Jablonica; My—Myjava; Po—Podbranč; Ch—Chropov; Br—Brezová

*2. ábra.* A kompressziós feszültség maximumának vázlata az oligocén-alsó miocén ideje alatt
Mérési helyek: De—Devín; Zá—Záhorská Bystrica; Bo—Borinka; Ro—Rohožník; So—Sološnica; Ra—Raková; Hr—Hradište; Dc—Dechtice; Me—Melčice; Fa—Fajnorov lom; Ja—Jablonica; My—Myjava; Po—Podbranč; Ch—Chropov; Br—Brezová

*Рис. 2.* Схема ротации максимума компрессионного напряжения с олигоцена до нижнего миоцена
Точки измерений: De—Девин; Za—Загорска Быстрица; Bo—Боринка; Ro—Рохожник; So—Солошница; Ra—Ракова; Hr—Градиште; Dc—Дехтице; Me—Мельчице; Fa—Файноров лом; Ja—Яблоница; My—Мыява; Po—Подбранч; Ch—Хропов; Br—Брезова

## 3. Early Miocene basins

The opening of early Miocene basins in the Alpine-Carpathian junction was caused by NW thrusts, reverse fault and strike-slips to a lesser extent in the Flysch Zone [NEMČOK 1991]. In the rheologically different Central Zone of the Western Carpathians, a combination of dextral WSW-ENE trending strike slips and back thrusts as compensation structures were predominant. With advancing collision and paleostress field rotation the dextral strike-slips also acquired an important component of reverse motion (*Figs. 3* and *4*). Active faults rimming the borders of the sedimentary basin gave rise to source areas of coarse material and induced slumping on basin slopes. Their activity is documented also by folding of Ottnangian sediments along the WSW-ENE trending Dobrá Voda fault zone.

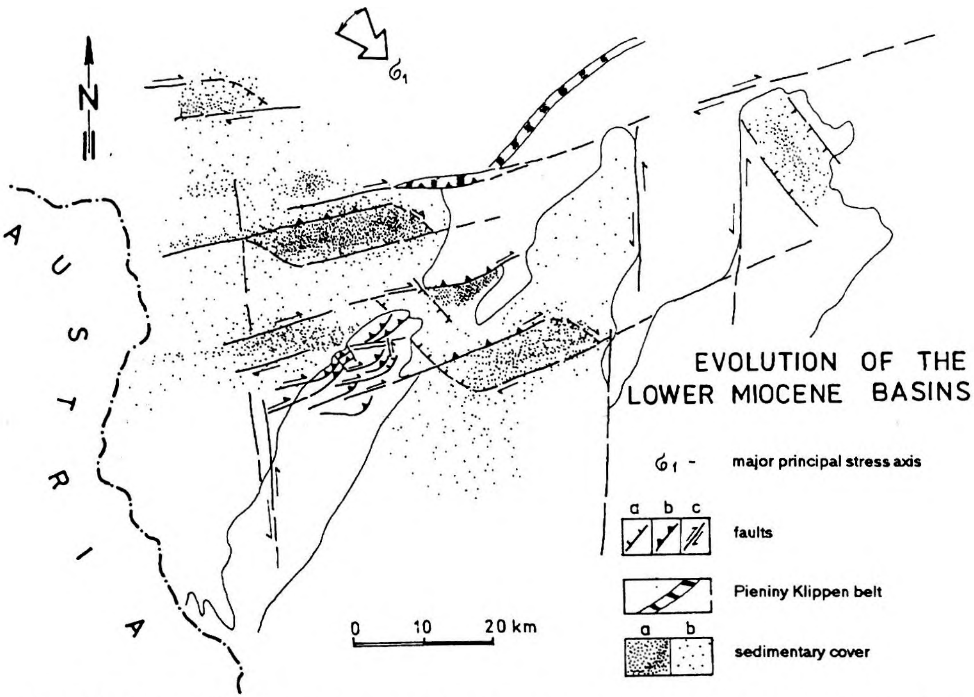


*Fig. 3.* Evolution of the Lower Miocene sedimentary area in the Central Zone of the Western Carpathians
Faults: a—normal; b—reverse; c—srtike-slip; Sedimentary cover [by JIŘÍČEK 1988, ADAM, DLABAČ 1969]: a— maximum thickness of sediments; b—occurrence of sediments

*3. ábra.* Az alsó miocén üledékes terület fejlődése a Nyugati-Kárpátok középső zónájában
Vetők: a— normal; b—feltolódásos; c—oldaleltolódásos; Üledékes takaró [JIŘÍČEK 1988, ADAM, DLABAČ 1969 után]: a—maximális üledékvastagság; b—üledékek előfordulása

*Рис.3.* Развитие осадочного бассейна в средней зоне Западных Карпат в нижнем миоцене
Разломы: а— нормадьные; –b—надвиги; с—сдвиги; Осадочные покровы [JIŘÍČEK 1988, ADAM, DLABAČ 1969]: а—максимальная мощность осадков; b—проявление осадков

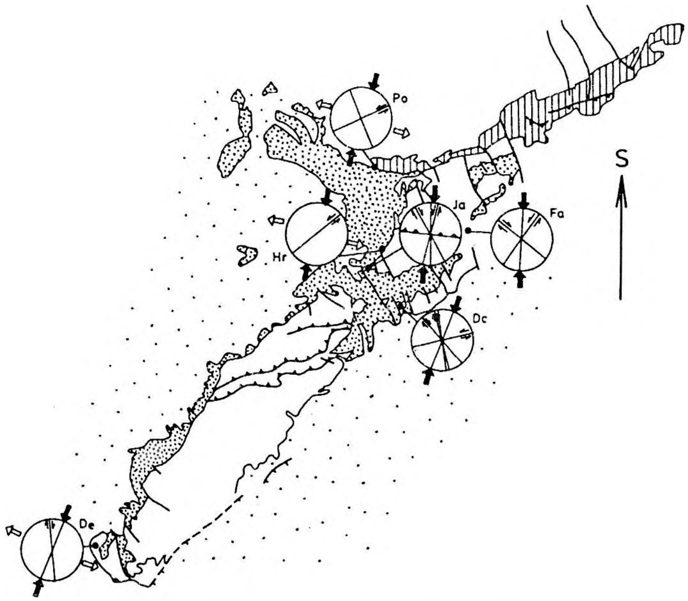


*Fig. 4.* Scheme of maximum compressional stress orientation during the Lower-Middle Miocene transition (Explanations, see Fig. 2)

*4. ábra.* A kompressziós feszültség irányítottságának vázlata az alsó miocén-középső miocén átmenet ideje alatt (jelmagyarázat lásd 2. ábra)

*Рис.4.* Схема направленности компрессионного напряжения на границе нижнего и среднего миоцена (условные обозначения как на Рис. 2)

The Lower Miocene basins in the western part of the Carpathians opened during the Eggenburgian. Depocentres of EW trending sedimentary area were situated on the Flysch Zone basement in the present-day Vienna Basin, where the Bathysiphon-Cyclammina schlieren reached its maximum thickness of 700 m in the Lužice depression [JIŘÍČEK, TOMEK 1981]. Eastwards, on the Central Carpathian basement, in contrast to the open, deep-marine regime in the western part, a shallow archipelago sea environment prevailed. Eggenburgian transgressive clastic sediments are composed of Triassic limestones and dolomites from local sources. However, the heavy mineral associations indicate a source lying now north and south of the basin. Conglomerates and breccias forming talus cones bound to reverse faults can be found in a belt extending from Dobrá Voda to Čachtice [KOVÁČ et al. 1989, KOVÁČ et al. 1991].

Decrease of salinity during the Ottnangian was caused by the isolation from the Mediterranean. In the Vienna Basin, Cibicides-Elphidium schlieren was deposited with maximum thickness of 600 m in the Lužice and 700 m in the Štefanov depression [JIŘÍČEK 1988]. Eastwards, on the Central Carpathian basement, dark aleuropelitic deposits with frequent fish scales and bioturbation attain a thickness of 100-150 m in the Dobrá Voda depression. Intercalations of fine-grained conglomerates contain quartz pebbles in addition to the carbonates; the quantity of quartz pebbles increases upwards. The shore of the 'wadden' tidal flat dominated sea provided clastic material mainly from sources lying south of the basin, currently represented by the basement of the Danube Basin [KOVÁČ et al. 1991].

The Lower Karpathian marine sedimentation culminated in the deposition of 'Karpathian schlieren', due to a paleogeographic change which opened the connection again to the Mediterranean. The dynamics of intramontane basins in the western areas of the Carpathians is indicated by frequent occurrence of turbidities and slump bodies in pelitic facies and cyclic sedimentation on the basin margin. The marginal facies is represented by conglomerates, sandstones and claystones known from the Dobrá Voda depression and the NW part of the Danube Basin. The composition of clastic material is similar to that in the Ottnangian, its supply continued to be derived from local sources and from the south. Quick subsidence and high accumulation rate resulted in good preservation of organic matter. The thickness of the basin Lakšár Formation in the Vienna Basin reached 600–900 m [ŠPIČKA 1969], its equivalent in the Bánovská kotlina Basin is 500 m thick [BRESTENSKÁ 1980].

In the Upper Karpathian, paleobathymetric differentiation of the sedimentary area took place. The depocentres of the intramontane basins migrated continuously southward. In the northern part of the Vienna Basin, brackish sedimentation of the Závod Formation started with Šaštin sands. This formation is sometimes up to 400 m thick, passing upwards into calcareous clays alternating with sandy beds [ŠPIČKA 1969]. In the southern part of the Vienna Basin a fresh-water sequence of the 1000–1500 m thick Láb ostracode beds was deposited corresponding to the Aderklaa Formation in Austria [JIŘÍČEK, TOMEK 1981].

The Upper Karpathian Jablonica conglomerates and sandstones corresponding to the Aderklaa conglomerates in Austria attain a thickness of 300–500 m in the wider area of the Dobrá Voda depression. Their polymict pebble material is composed of Mesozoic carbonates, Cretaceous and Paleogene sandstones, quartz, quartzites, granitoids, crystalline schists, acidic and basic volcanics [KOVÁČ 1986, MIŠÍK 1986]. The diversity of the pebble material and the heavy mineral associations transported in humid climatic conditions into this alluvial delta fan point to sources situated at the presenst time south and north of the Dobrá Voda depression [KOVÁČ et al. 1991].

## 4. Middle Miocene changes of paleogeography and the structural pattern

During the Middle Miocene the western area of the Central Zone of the Carpathians underwent important paleogeographic changes. The Lower Miocene deposits of the Jablonica and Dobrá Voda depressions were incorporated into the arising Little Carpathians horst structure, separating the Vienna and Danube Basins which started to open.

This stage of development is characterized by the Badenian clockwise rotation of maximum compression axis to NE-SW (NNE-SSW) direction, which caused left lateral slips along the above-mentioned WSW-ENE trending reverse fault system (*Figs.* 5 and 6). Good evidence is provided by the Dobrá Voda fault zone which formed a transpressional push-up dome in the Brezovské pohorie Mts. bordering a wrench fault furrow filled by Lower Miocene sediments (Fig. 1). The fault zone activity caused also block rotation between major strike-slip faults. NW-SE dextral strike-slips or dextral-normal faults and NE-SW normal faults compensated the deformation [MARKO et al. 1991]. Counter-clockwise rotation of blocks in the northern part of the Little Carpathians is confirmed by paleomagnetic observations [TÚNYI, KOVÁČ 1991]. This resembles a similar phenomenon in the northern Pannonian domain [MÁRTON 1987].

## 5. Middle and Late Miocene basins

The Badenian origin of the Vienna and Danube Basins was explained by wrench faulting producing local extensional areas (pull-apart basins), where intensive accumulation of sediments took place [ROYDEN et al. 1982, FODOR et al. 1990 and others]. This splendid idea was also confirmed by basin analysis carried out later in the Pannonian domain [ROYDEN, HORVÁTH 1988]. Rapid Badenian subsidence points to tectonic activization of NNE-SSW and NE-SW trending normal and oblique slip faults, with high amplitude of throw accompanied by NE-SW and ENE-WSW trending sinistral strike slips. They were related to stress field whose maximum principal stress axis rotated from N to NE, during the Lower-Middle Badenian [NEMČOK et al. 1989]. The faults rim several thousand metres deep rhombohedral depocentres both in the Vienna and Danube Basin. Only a tectonic explanation of this phenomenon is acceptable for such local thinning of the crust. Also conspicuous is the time shift of the culmination of subsidence from the Vienna to the Danube Basin [VASS, ČECH 1983], that may be related to an eastward migration of the active Carpathian collision [ROYDEN et al. 1982, JIŘÍČEK 1979].

The Lower Badenian represents a separate sedimentary cycle displaying in marginal parts an angular discordance with the Karpathian strata. From the paleogeographic point of view, the sedimentary area of the 'Lanzendorfer Serie' shows greater affinity to the Karpathian one than to the sedimentary one of the Middle and Upper Badenian. In the western part of the Central Carpathians the Lower Badenian transgression intensively affected only the Vienna Basin. The marginal facies along the SW slopes of the Little Carpathians is represented by conglomerates (with pelitic intercalations), the pebble material of which consists of Mesozoic carbonates. Granites occur sporadically. Overlying granitoid conglomerates and breccias of the Devínska Nová Ves Beds [VASS et al. 1988] document steep morphology. The Middle Badenian talus cones and debris aprons with some boulders bigger than 2 m attain a thickness

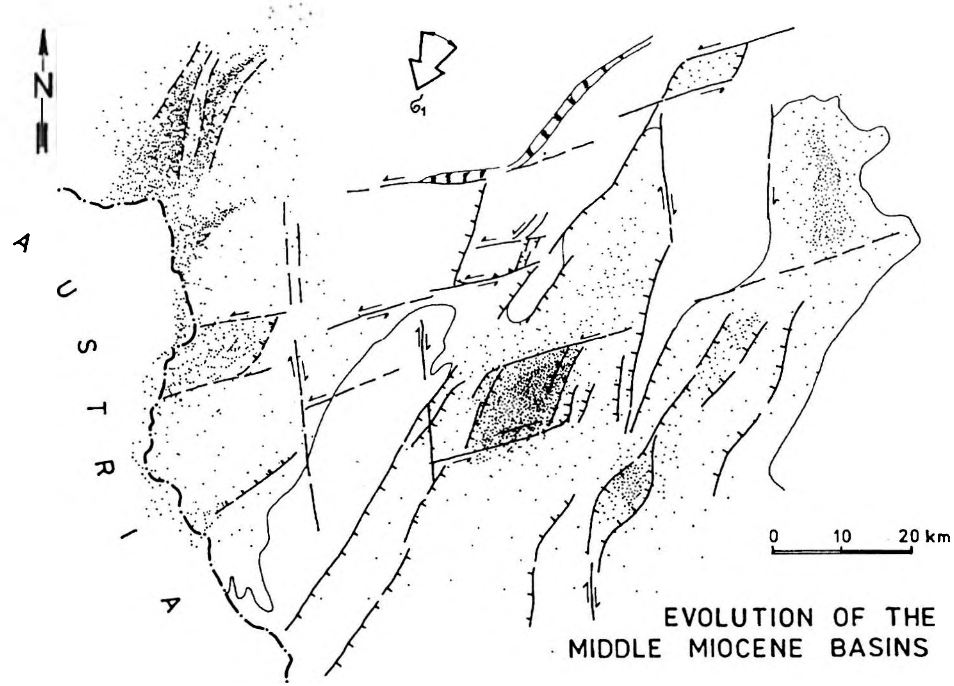


*Fig. 5.* Evolution of the Middle Miocene basins in the western area of the Central Zone of the Western Carpathians
5. *ábra.* Középső miocén medencefejlődés a Nyugati-Kárpátok középső zónájában
*Рис.5.* Развитие бассейна в средней зоне Западных Карпат в среднем миоцене

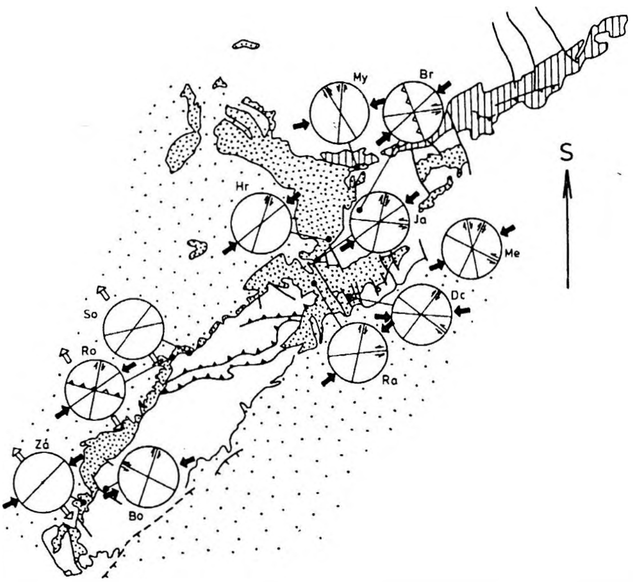


*Fig. 6.* Scheme of maximum compressional stress orientation during the Middle Miocene (Explanations, see Fig. 2)

*6. ábra.* A kompressziós feszültség irányítottságának vázlata a középső miocén ideje alatt (jelmagyarázat lásd 2. ábra)

*Рис.6.* Схема направленности компрессионного напряжения в среднем миоцене (условные обозначения как на Рис. 2)

of 300-350 m. In some places the Upper Badenian marine clastic shore sediments algal bioherms and patch reefs were formed. Abrupt transition of the marginal facies into basinal ones indicates the activity of NE-SW trending normal faults on the eastern margin of the Vienna Basin. The Lower Badenian pelitic facies of the 'marine Tegel' (Lanžhot Formation) reaches a thickness of 500-800 m. In contrast to this, the Middle and Upper Badenian basinal deposits often contain shallow and freshwater intercalations. The sequence is represented mainly by clays with increasing sandy intercalations upwards. The Middle Badenian Jakubov Formation attains a thickness of 400-700 m and the Upper Badenian Studienka Formation ranges from 400 m to 600 m [JIŘÍČEK, TOMEK 1981].

In the NW part of the Danube Basin (Blatné depression) the Lower Badenian deposits are known only from one borehole in the vicinity of Trakovice. Conglomerates, sands and clays contain tuffitic intercalations in the central part of the 560 m thick sequence that may be associated with volcanites buried beneath sediments of the Danube Basin. The subsidence during the Middle and Upper Badenian was very fast. The basal part of the 2700-3000 m thick sequence represents conglomerates and sandstones passing upwards into clays with sandy intercalations [GAŽA 1973]. Their foraminifera associations, similarly to those in the Vienna Basin, indicate stratification of water masses, a poorly-aerated normal salinity lower layer and an upper layer with decreased salinity [MOLČÍKOVÁ 1962]. The marginal facies on the NE slopes of the Little Carpathians is represented by the Upper Badenian Dolany polymict conglomerates with sand lenses. The pebbles consist predominantly of limestone, quartz, less amount of granitoid, phyllites, amphibolites and other crystalline pebbles from local sources [BUDAY et al. 1962].

The Sarmatian depocentres in the Vienna and Danube Basins coincide with the Badenian ones. The subsidence along NNE-SSW to NE-SE faults becomes moderate. In brackish environment, due to isolation from the Mediterranean, basal clays were deposited with increasing sandy intercalations and beds, upwards. Their thickness varies from 300 m to 600 m [GAŽA 1983, JIŘÍČEK 1988]. Marginal sandy facies rich in $CaCO_3$, containing in some places lumachelle beds, Bryozoa-Serpulides bioherms, oolitic and Nubecularia limestones are preserved as denudation remnants on both sides of the Little Carpathians horst structure.

During the Upper Miocene and Pliocene the transpressional conditions changed to transtensional ones in the western part of the Carpathians. During the regional extension NE-SW normal faults governed the sedimentation. The intramontane basins in the western part of the Central Zone of the Western Carpathians became bays of the brackish Pannonian lake; later, in the Pontian stage, of a fresh-water lake. In the Vienna and Danube Basins a complex of clays, sands and conglomerates were deposited and in the marginal part also coal seams. Centres of sedimentation migrated southwards. In the Austrian part of the Vienna Basin and in the Danube Basin (Gabčíkovo depression) the thickness of sediments reaches 1500 m. Deformation of the Pannonian sediments near the village of Solośnica on the NW margin of the Little Carpathians indicates emplacement along NE-SW normal faults, but also along ENE-WSW trending sinistral strike-slips, which could be considered as a result of the NW-SE trending extension [MARKO et al. 1990].

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## NEOGÉN SZERKEZETI FEJLŐDÉS ÉS MEDENCE KIALAKULÁS A NYUGATI-KÁRPÁTOKBAN

M. KOVÁČ, F. MARKO és M. NEMČOK

A neoalpi fejlődési szakasz oligocén-alsó miocén transzpresszív fázisa irányította a Kárpátok tektonikai leválását a kelet-alpi tartományról és a Ny-K irányú üledékes terület kinyílását a Nyugati-Kárpátok középső zónájának nyugati területén. A főfeszültségi maximum órajárással egyező rotációja északról északkelet felé a szerkezeti felépítésben változásokat okozott, melyek a Kis-Kárpátokban jól megfigyelhetők, és kialakította a jelenlegi Bécsi- és Duna-medencéket a középső miocén ideje alatt. A felső miocén és pliocén medencefejlődés extenziós körülményekre utal.

## СТРУКТУРНОЕ РАЗВИТИЕ И ОБРАЗОВАНИЕ БАССЕЙНА В НЕОГЕНЕ В ЗАПАДНЫХ КАРПАТАХ

М. КОВАЧ, Ф. МАРКО, и М. НЕМЧОК

Тектоническое отделение Карпат от восточно-альпийского пояса и образование осадочного бассейна восточного простирания на западной части средней зоны Западных Карпат были связаны с транспрессивной фазой новоальпийского развития в олигоцене-нижнем миоцене. Ротация направления максимального напряжения с севера до северо-востока по часовой стрелке вызвала изменения структурных особенностей, хорошо наблюдаемых в настоящем в Малых Карпатах, и привела к образованию в среднем миоцене Венского и Дунайского бассейнов. Развитие этих бассейнов в верхнем миоцене и плиоцене показывает на экстенсионные условия.

## INVITATION

The Association of Hungarian Geophysicists decided at its annual meeting to establish the "Foundation for Hungarian Geophysicists" and elected its first Advisory Board for 3 years. The foundation has been started with a moderate initial capital of 300 000 HUF, which has by now increased to more than 3 million and it is open for everybody.

The aim of the foundation is to help Hungarian geophysicists. There are two main target groups whose application for grants will be accepted with preference: young geophysicists needing assistance (travels, participation at conferences, publications,post-graduate education etc.) at the beginning of their professional life as well as retired and unemployed colleagues whose economic and social position became especially unfavourable.

The nine members of the Advisory Board invite everybody to join this foundation; donations should be communicated with the Board. Organisations and persons donating sums exceeding the initial capital will have the opportunity to delegate representatives into the Board. Detailed information is available at the following address:

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"Foundation for Hungarian Geophysicists"
H–1371 Budapest, P.O.B. 431
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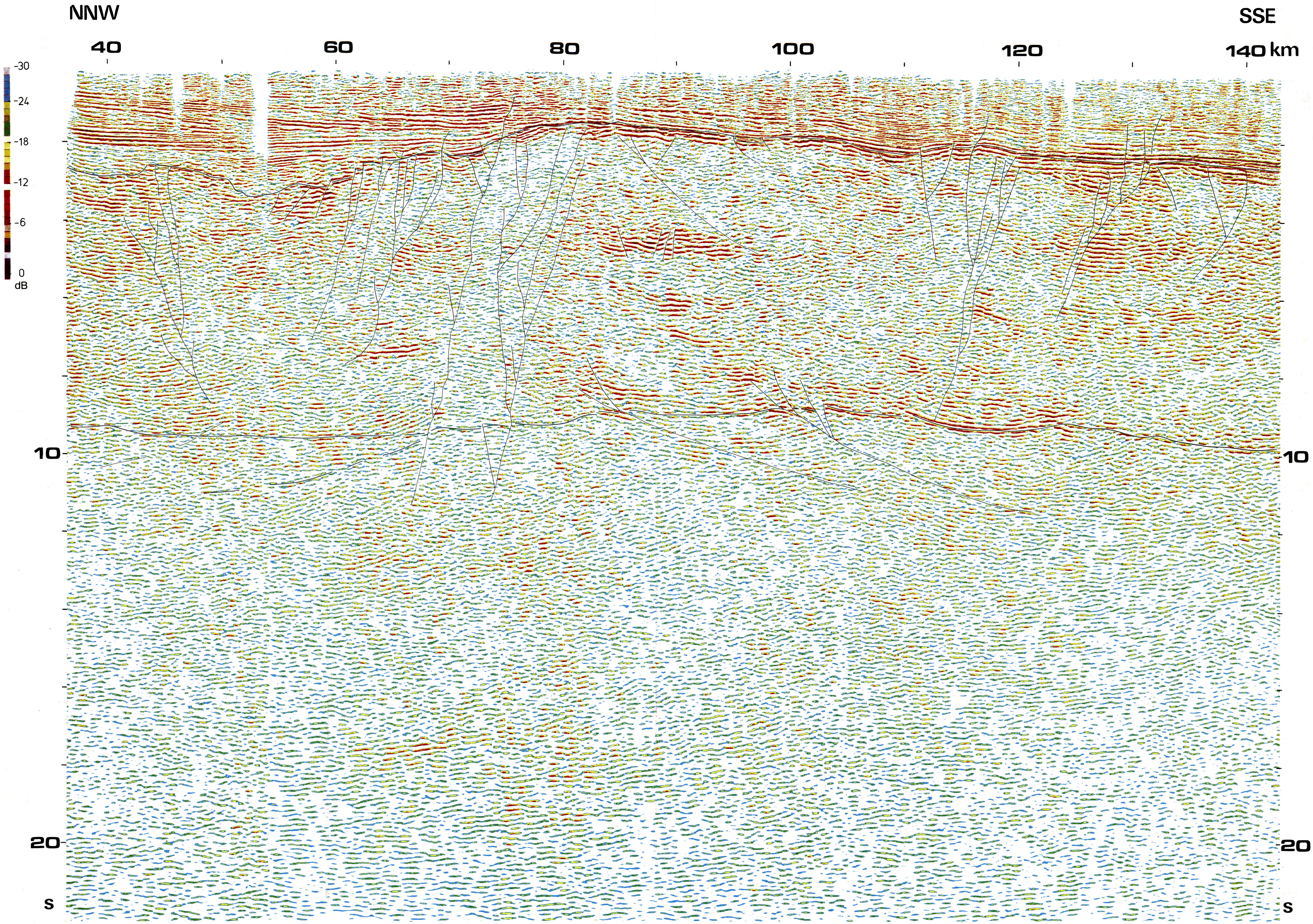
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