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Stress transfer and seismic instabilities in the upper crust: example of the western Pyrenees

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Abstract—When the regional state of stress is close to the condition for earthquakes to occur, small stress and strain changes could trigger seismic instabilities. Stress diffusion ($\Delta\sigma$ = 1–2 MPa) has been suggested to explain, at lithospheric scale, space and time correlated earthquakes and surface deformations in subduction zones (100–200 km, 10–40 years). We test the possible regional mechanisms, at the upper crustal scale, in the western Pyrenees, where major regional events ($\mathbf{M}_{max}$ = 5.7), locally induced events and subsurface fluid pressure fluctuations are temporally and spatially correlated.

In a first phase, fluid withdrawal (decrease of pore pressure of the aquifer at the gas–water interface) affects the natural regional seismicity rate. The locked fault inhibits small events for a time, during which tectonic stresses accumulate. Major regional events then occur on the Pyrenean Fault, when the energy of deformation is large enough to allow seismic instabilities to occur. These major events ($\mathbf{M}_l > 5$) trigger seismic instabilities ($\mathbf{M}_l = 4$), in the vicinity of the gas extraction area, 30 km away. As a result of gas extraction, the Lacq field zone has local weakness characteristics similar to the neighbouring main active fault. The Pyrenean earthquakes lead the gas field events by 2 years. This correlation is consistent with a viscoelastic model, where salt rock acts as a viscoelastic channel, below an elastic bed.

The possible connections proposed here may give a new insight worldwide in understanding the role of fluids in seismic instabilities and seismic risk assessment in the neighbouring of man-made fluid pressure fluctuations, when natural seismic activity is high.

INTRODUCTION

Seismic instabilities induced by increasing pore water pressure have been reported for many years. When injection of water takes place, the effective stress value decreases and the Mohr–Coulomb failure criterion seems to be verified *in situ* in different places for small pressure fluctuations (Raleigh *et al.* 1972, Fletcher & Sykes 1977, Nicholson *et al.* 1988). It was shown that connected distances in the upper crust can reach a tenth of a kilometre scale, with a time delay for pore pressure migration of a few years (cf. Hsieh & Bredehoeft 1981). Recent modelling of the water reservoir pore pressure effect on local seismicity shows that a small stress change (<0.1 MPa), could trigger or inhibit seismic instabilities depending on the orientation of pre-existing discontinuities in the vicinity of a storage dam (cf. Simpson 1986, Roeloffs 1988, Simpson *et al.* 1988a). Moreover, co-seismic stress perturbations after moderate earthquakes are reported to affect creep on a fault 30 km away (cf. Simpson *et al.* 1988b). Associated stress changes are small (<0.01 MPa). All the previous observations and models attest the possible effect of small stress changes induced by increased pore pressure on seismic or aseismic displacements.

Elsewhere Grasso *et al.* (1992) reported numerous cases of newly recognized instabilities. They are induced by underground fluid pressure fluctuations with decreased pore pressure. These instabilities are triggered or induced by fluid extraction. The driving stress changes involved, modelled using either poroelastic stress transfer or mass effects, are of the same order as described previously; i.e. 0.1–1.0 MPa (cf. Segall 1989, McGarr 1991).

In the Lacq area of the French Pyrenees, accurate determination of the location of seismic events shows that few events, if any, occur within the gas reservoir itself, in agreement with the increase in effective normal stress (Grasso & Wittlinger 1990, Guyoton *et al.* in press). Two seismic activity phases were observed. In a first period (1974–1980) most of the seismic ruptures occur within the stiff part of the overburden, with a highly diffuse location pattern (Grasso & Wittlinger 1990). Since 1982, deep events have occurred below the reservoir and are more organized on large pre-existing faults, defining numerous clusters (Guyoton *et al.* in press). The stress change in the overburden and in the beds below the reservoir have been modelled by Segall & Grasso (1991) using poroelastic stressing. The basic result is that stress changes outside the reservoir (where seismic instabilities occur) are proportional to the decrease in pore pressure within the gas reservoir, despite the fact that no fluid connection exists between the reservoir and the overburden. The calculated change in

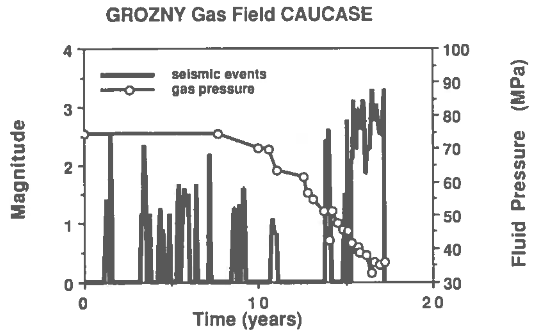

Fig. 1. Seismicity and gas pressure decrease due to gas extraction (Grosny field, Tchechen Republic. Caucasus) from 1956 to 1974 (Smirnova personal communication, 1990). Zero value on the time axis corresponds to year 1956. Pressure values are the average gas reservoir pressure. Earthquake magnitudes are local magnitude.

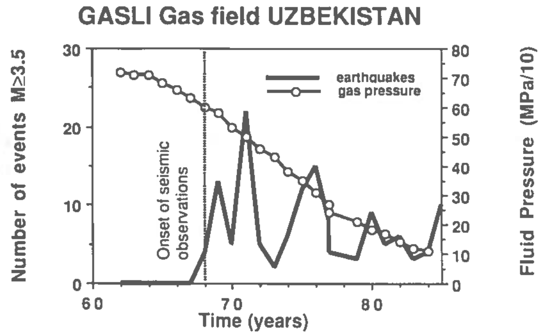

Fig. 2. Seismicity and gas pressure decrease due to gas extraction (Gasli field, Uzbekistan) from 1960 to 1984 (adapted from Plotnikova *et al.* 1989). Zero value on the time axis corresponds to year 1960. Pressure values (MPa/10 = bars) are the average gas reservoir pressure. Earthquake magnitudes are local magnitude.

shear stress for a 25 MPa pressure drop after 10 years of extraction, which corresponds to the onset of seismic activity in 1969, is less than 1 MPa. The same critical stress thresholds are necessary to explain both the rupture above and below the reservoir, except that, due to the free surface, the time delay to reach this threshold is twice as long below the reservoir as above it. Even if such stress changes are small, they are of the same magnitude as those known to drive seismic instabilities in the vicinity of artificial water reservoirs (Roeloffs 1988).

When reported, the seismic instabilities induced by fluid extraction are regional patterns in most of the observed cases: when induced seismic activity occurs within one field, it is also reported in the neighbouring ones (cf. North Sea, Holland, France, Uzbekistan; Grasso 1991). This could imply that a continental plate in mechanical equilibrium and faulted over its entire surface, corresponds to a state of the crust in which the rupture criterion is almost reached everywhere (cf. Sornette *et al.* 1990). This condition is intimately related to the observation at another scale of active faults almost everywhere within the plate. From the above setting, the possible relations between fluid pressure fluctuations, local triggered events and regional seismicity patterns can be analyzed. While the local seismic instabilities sustained in the Lacq area over more that 20 years are driven by the monotonic poroelastic stressing alone, the onset of the two seismic phases in the Lacq area, i.e. 1969—first $\mathbf{M}_l > 3$ event above the reservoir, 1982—first $\mathbf{M}_l > 3$ event below the gas reservoir, may have been triggered by the combined effects of both the local poroelastic stressing and the stress transfer due to the major regional earthquakes.

In this paper, it is proposed to investigate, at a regional scale, the possible effects of local fluid pressure fluctuation. It is interesting to note that the possible connection between regional seismicity patterns and the history of decrease in pore pressure in hydrocarbon reservoirs is observed in different regions, cf. Grozny field (Fig. 1), Tchetchen Republic, Caucasus, Gasli and Tchurtan fields (Figs. 2 and 3), Uzbekistan, Lacq field, Pyrenees, France (Fig. 4). Although processes explaining the local instabilities induced in the vicinity of fluid pressure fluctuations were recently proposed and reviewed by Grasso *et al.* (1992), the regional effect at upper crustal scale has not yet been investigated. Here it is proposed to test the possible regional mechanisms, at the upper crustal scale, in the western Pyrenees where natural seismicity ($\mathbf{M}_{max} = 5.7$) is located 20–30 km from the seismicity triggered by hydrocarbon extraction. In this area, geological and hydrological data fix the limits of the estimates of pore pressure, and stress and strain changes. At the upper crustal scale, the geometry of the layers involved is governed by décollement levels and rocksalt beds. If these processes are verified, these pressure changes must be considered in order to understand the role of fluids in fault mechanics and medium-term seismic prediction.

PRINCIPLES OF METHODS AND APPLICATION IN THE WESTERN PYRENEES

The accurate geomechanical analysis of the Lacq gas field response to fluid pressure fluctuations is performed using levelling data and induced seismicity (Grasso & Feignier 1990). The gas field induced swarm is located 25 km north of a high seismicity area (Fig. 5). This seismicity defines a narrow strip, well defined over several hundred kilometres as the main seismic fault of the Pyrenees (the North Pyrenean Fault). Only a few large earthquakes have originated in this pre-existing zone of weakness over the past centuries ($\mathbf{M}_l = 5.7$ in 1967; $\mathbf{M}_l = 5.3$ in 1980; $\mathbf{M}_l = 5.0$ in 1982). All these events occurred within 30 km of the gas field. Two possible connection mechanisms between the two seismic areas can be investigated. The corresponding processes are water pressure diffusion and viscoelastic stress transfer.

Water pressure drop

Looking at the Grosny, Gasli, Tchurtan and Lacq examples, the first part of this paper suggests that gas extraction changes the hydrological state of the area.

(i) When fluid connection is geologically possible (i.e. water flow along décollement level), the decrease in gas pressure induces a decrease in water pressure around

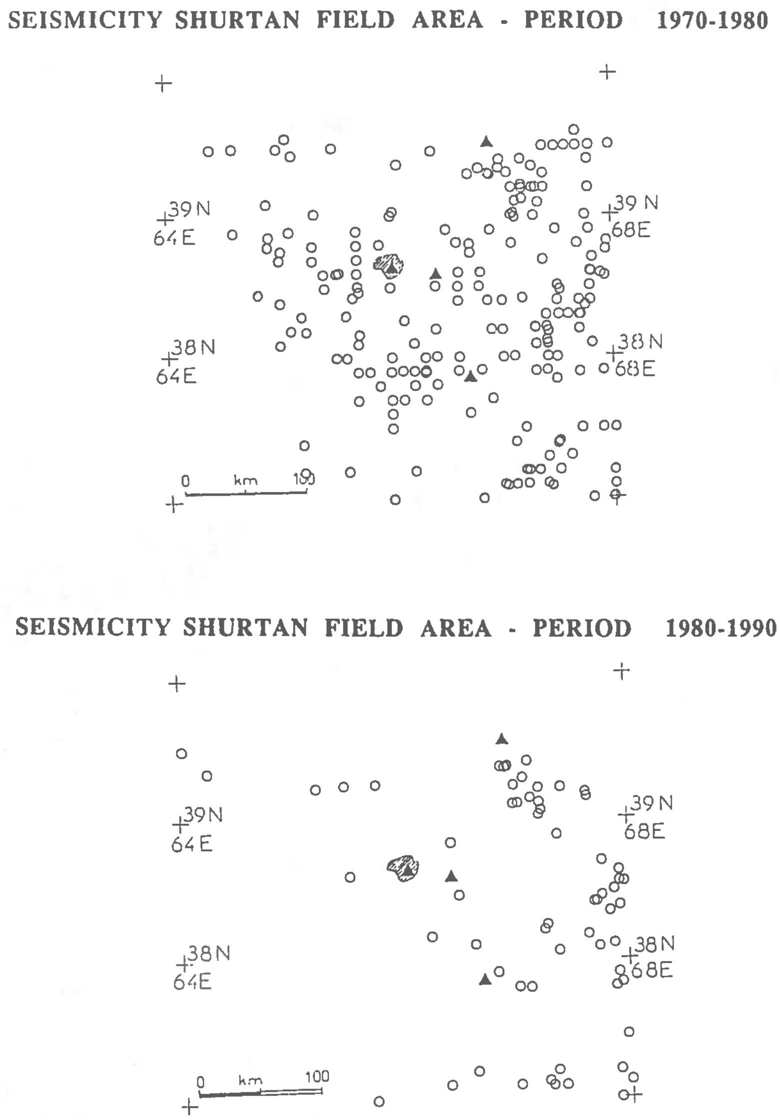

Fig. 3. Siesmicity map of Shurtan gas field area, Uzbekistan. Black triangles are local seismic stations. (a) Period 1970–1980. (b) Period 1980–1990. Note that the gas extraction began in 1979, with an initial gas pressure of 30 MPa. The gas pressure decrease since 1979 is 2 MPa for a production of 20% of the known reserves (Plotnikova personal communication, 1991).

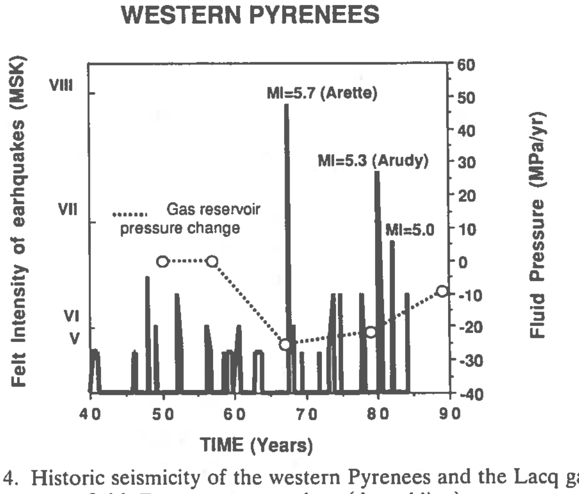

Fig. 4. Historic seismicity of the western Pyrenees and the Lacq gas field, France; pressure drop (dotted line).

the neighbouring faults. Such an increase in effective pressure inhibits seismic fractures in the area during the first years of extraction.

(ii) In a second step, these locked faults inhibit deformation and increase the seismic energy potential. This energy of deformation is released in a second phase at a greater magnitude than before. This classical asperity model is checked at greater scale by numerous seismic cycle observations. At laboratory scale, the damage theory explains the increase in seismic energy after a quiescent period (cf. Main *et al.* 1990).

Viscoelastic stress transfer

In the second part of the paper, the after-effects of these major regional events on the Lacq gas field seismicity are tested. The seismic history of the Lacq field (**M**$_l > 3$ in 1969, first felt earthquake; **M**$_l = 4.4$ in 1981, strongest induced event; deep events first observed in

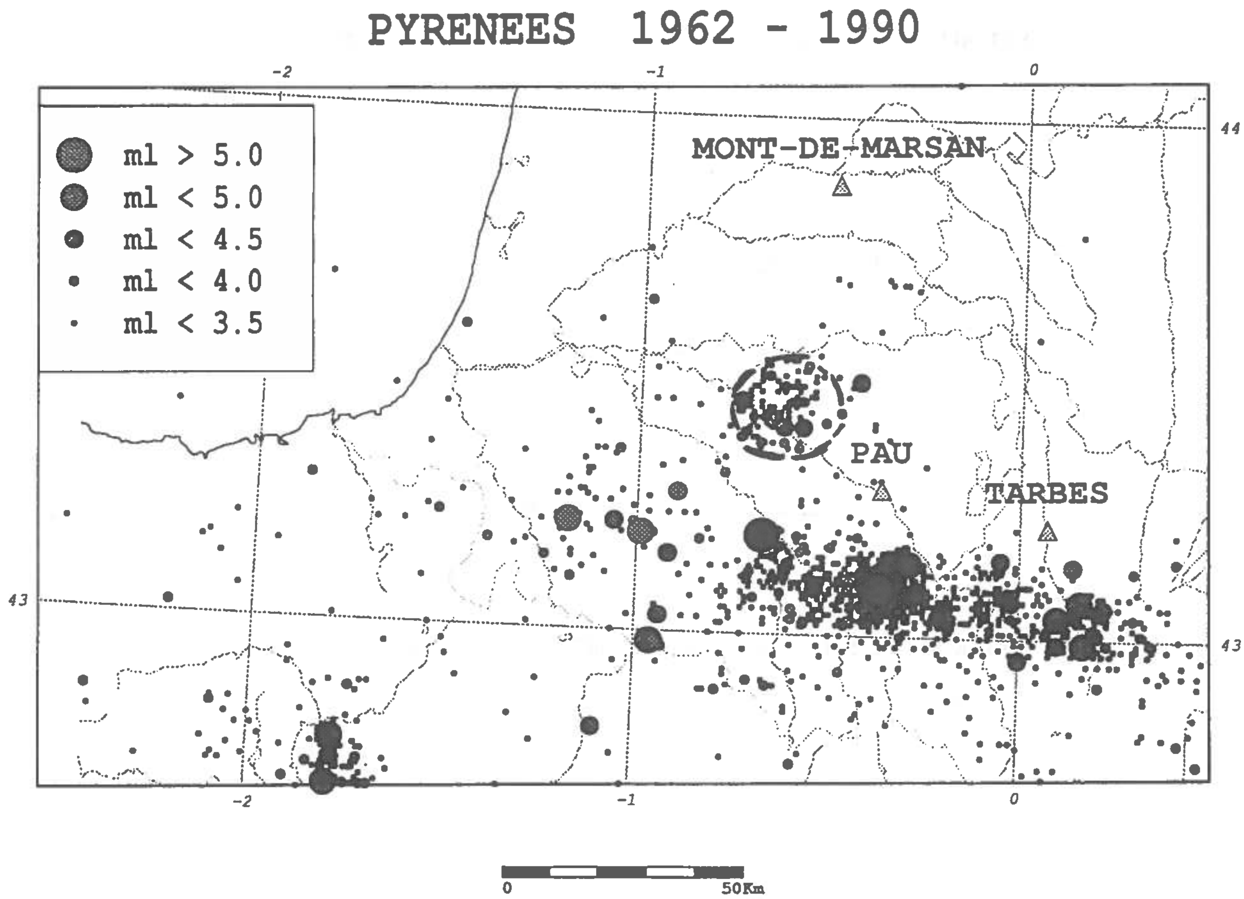

Fig. 5. Seismicity map of the western Pyrenees, France, 1962–1990 (from LDG, CEA French national network). Earthquake magnitudes are local magnitude. The seismically active zone in the Western Pyrenees defines an E–W strip no more than 15 km wide, in the lower part of the map, which is associated with the North Pyrenean Fault. The swarm located 20–25 km north of the main Pyrenean fault (circular outline) has been active since 1969 and is correlated in time and space with extraction from the major Lacq gas field in France (Grasso & Wittlinger 1990).

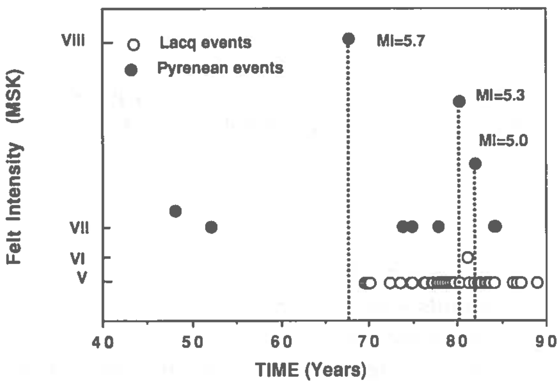

Fig. 6. Regional seismicity of the Western Pyrenees and local triggered earthquakes in the Lacq gas field area. Zero value on the time axis corresponds to year 1940. Earthquake magnitudes are local magnitude.

late 1982 with $M_l > 3$) is connected with the major seismic energy release affecting the North Pyrenean Fault (Fig. 6). As a result of increasing stress induced by gas extraction, the Lacq field zone is close to the critical failure threshold as well as the seismically active regional faults. Around this threshold, small perturbations can trigger instabilities. Earthquake after-effects and triggered seismic phenomena were reviewed by Rice & Gu (1983). One of the mechanisms tested in the Pyrenean area was proposed by these authors at a larger scale. They study major events (magnitude >7) and stress alteration over distances ranging from 100 to 500 km. The entire lithosphere and asthenosphere are used in a sense to channel stress alteration depending on the physical mechanisms of earthquake after-effects. In our model we test the same viscoelastic stress transfer model at the scale of the sedimentary cover.

The manner in which geological, mechanical and hydrological data can limit the effect of each process in this particular case study is investigated. The studies are based on simple models and anticipate that important lessons remain to be learned from first-order models. It is important to note that such possible connections are observed in different areas where fluid manipulations trigger seismic instabilities, cf. Gasli field, Uzbekistan and Monteynard dam, France. If verified, such changes must be considered for medium-term prediction.

WATER PRESSURE CHANGE RESULTING FROM WATER DISCHARGE AROUND GAS FIELD EXPLOITATION

During discharge from a well, water is removed from the aquifer, and this produces a cone of depression in the water table. Following Darcy's law, this hydraulic gradient induces a flow of water. The change in water head (Δh), is governed by various relations depending on the reservoir geometry (cf. Nicholson *et al.* 1988).

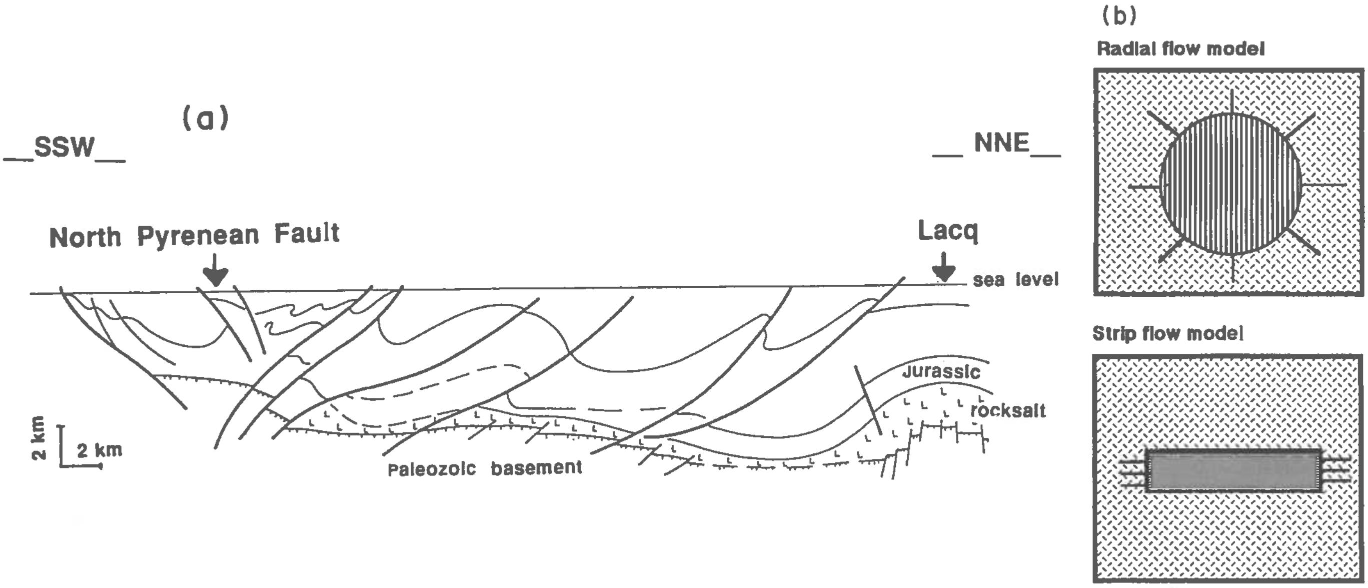

Fig. 7. (a) Geological cross-section between the Lacq field and the North Pyrenean Fault. Rock salt is indicated by small ticks below the Jurassic level and above the Paleozoic basement. The Lacq gas reservoir is located within Jurassic beds, and the aquifer is assumed to follow the décollement level above the basement. (b) Schematic models of the aquifer.

As a result of a constant flow rate (Q), from an isotropic reservoir of infinite area (with radial flow into the well), the change of head at distance r and time t is given by the Theis equation:

$$\Delta h(r,t) = \left(\frac{Q}{4\pi T}\right) \int_u^\infty \frac{e^{-\xi}}{\xi} d\xi$$

with T equal to transmissivity (hydraulic conductivity multiplied by aquifer thickness, $T \sim L^2 T^{-1}$), S equal to storativity (amount of fluid released per unit column of aquifer, for a unit decline of head, S is dimensionless), Q equal to out-flow rate and

$$u = r^2 S/4Tt \quad \text{(Freeze \& Cherry 1979).}$$

If fluid flow is confined to a narrow reservoir (strip aquifer model), the head change at a given distance is higher than for the radial flow model (cf. Hsieh & Bredehoeft 1981). For discharge from a well situated at the centre of a strip of width w and infinite extent in the x direction, a constant discharge rate induces a head change given by the following relation (cf. Hsieh & Bredehoeft 1981):

$$\Delta h(x,y,t) = \left(\frac{Q}{4\pi T}\right) \sum_{m=-\infty}^{m=\infty} \int_{u_m}^\infty \frac{e^{-\xi}}{\xi} d\xi$$

with

$$u_m = (x^2 + (y + mw)^2)\, S/4Tt,$$

y being the distance from the centre of the strip in the direction perpendicular to x, w is the width of the strip. The length of the strip is infinite in the x direction.

To introduce more clearly this head change in the mechanical behaviour of the deep rocks, most authors use a relation between head change and fluid pressure change (p) as follows:

$$p = \Delta h \rho g,$$

with ρ = density of the fluid.

Such relations were applied to the largest gas field in France (Lacq gas field, Nicolai *et al.* in press), in order to test a possible correlation between the gas production and the regional seismic activity in the nearby mountain chain of the Pyrenees (25–35 km). Since the distance between the gas field wells is rather small compared to the distance from the gas field to the Pyrenees, the cluster of wells were modelled as a single point source of 100 m in diameter.

Using more than 40 deep boreholes in the area, as well as numerous deep seismic soundings from Elf (SNEA(P)) and the preliminary results of Ecors Arzaq seismic profile (Fig. 7), the limit values for the geometry and hydromechanical characteristics of the medium between the Lacq gas field and the North Pyrenean Fault can be determined.

The gas producing reservoir includes limestones and dolomitic formations (Cretaceous and Jurassic), with average thickness of 500 m, but with heterogeneous permeability values (Nicolai *et al.* in press). Using the simplified geological cross-section in Fig. 7, a continuous aquifer through these formations was assumed from the root of the Lacq field to the Pyrenees. Initial gas pressure was 66 MPa at 3.2 km depth. A gas–water contact was never found but was estimated from the abnormally high gas pressure to be located at least at 7 km depth. As a result, this gas reservoir is known as a dry reservoir. The estimated horizontal aquifer is presumed to be down-bounded by the well-known décollement level of the area. A mean thickness of 200 m, a mean porosity of 3%, and a mean hydraulic conductivity of 10^{-8} ms^{-1} were assumed within the aquifer according to the geomechanical data from deep boreholes (Grasso & Feignier 1990).

For given values of the storativity and transmissivity,

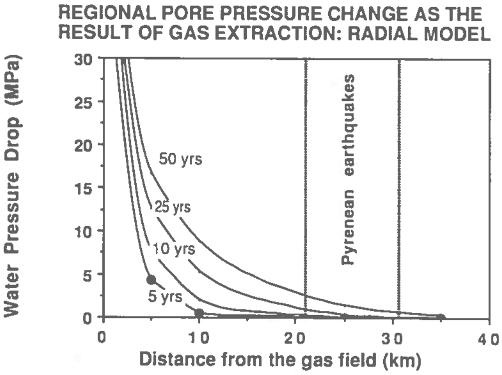

Fig. 8. Radial flow model. Pressure drop of the aquifer as a function of time delay after gas extraction began. Each curve corresponds to a time delay after gas extraction began.

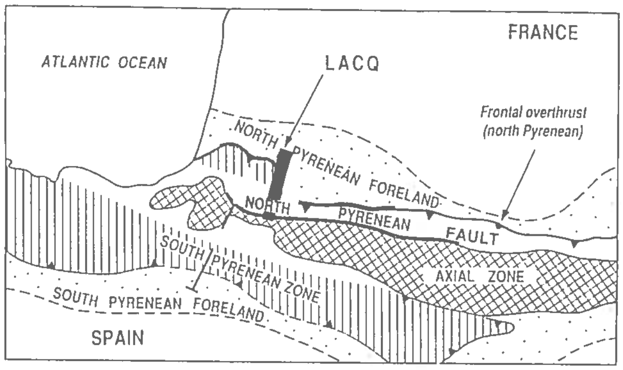

Fig. 9. Geological sketch map of the western Pyrenees. The rectangular box shows the geometry of the strip model used to compute pore pressure changes in Fig. 10.

the value of the water discharge rate must be estimated. The gas pressure drop associated with gas extraction is not isotropic but from an initial pressure of 66 MPa, the gas pressure drop is about 60 MPa at the single point source and 50 MPa at distances ranging from 5 to 10 km, and 35 MPa from 7 to 15 km (Elf (SNEA(P)) company personal communication 1991). According to the production data, with a 60 MPa gas pressure drop, the decrease in water pressure is fixed at 60 MPa at the centre of the gas field, after 25 years of extraction (Grasso & Wittlinger 1990). This means that at this point the water head is presumed to be equal to the gas head. The other boundary condition assumes that the maximum value of the water pressure decrease remains below the initial gas pressure, 66 MPa, after a long time interval (50–100 years). With these constraints, water flow rate (Q) must be close to 16×10^{-3} $m^3 s^{-1}$. This corresponds to a total water inflow volume of 12.6×10^6 m^3 over the last 25 years, a relatively small value vs the 9×10^9 m^3 reservoir volume. Note that the depth of major events located on the North Pyrenean Fault—Arette, Arudy Gagnepain (1987)—is of the same order as the depth of the estimated aquifer which is bounded by the knowledge of décollement level. Consequently, the question of downward migration of pressure effects does not need to be tackled.

Assuming a constant water flow rate, the change in head, and thus the change in pressure in the confined aquifer, were calculated using the above relations. The water pressure change is calculated using the radial flow model, and the values are plotted in Fig. 8, on a distance vs time diagram. After 25–50 years a decrease of less than 1 MPa (a few bars) may be obtained between distances ranging from 25 to 35 km.

The water pressure change is also calculated for the strip aquifer model. A strip zone of 5 km in width is used in order to represent an area of fractured rocks from the Lacq anticline to the North Pyrenean Fault. The location of this strip is shown on Fig. 9 with the main tectonic structure of the area. Note that according to a N–S geological discordance, the strip is located in an area where the North Pyrenean Thrust disappears. In this case the transmissivity value is assumed to be 2.5×10^5 $m^2 s^{-1}$ and the porosity is assumed to be 3.75% to account for the fault zone effect. The same boundary conditions are used as for the radial flow model, i.e. after 25 years delay, a 60 MPa water pressure decrease is reached at the single discharge source (100 m diameter) and, after a long period (50–100 years), the maximum decrease value remains below the initial pressure of the reservoir (66 MPa). To fit these limit values a 43×10^{-3} m^3 s^{-1} discharge rate has to be used. This corresponds to a total water flow volume of 3.4×10^7 m^3 since the beginning of gas extraction. In this case the water pressure decrease value is significantly higher, especially over a period ranging from a few years to 10 years (Fig. 10). Water pressure changes greater than 1 MPa can occur at 20–30 km distance from the reservoir within a period of 2 years. The seismicity patterns of the North Pyrenean Fault is compared to the water pressure change in Fig. 11(a). Such an estimation is checked using near-field data close to the gas reservoir. During the past 14 years, deep injection of waste water has occurred below the gas reservoir, with a total estimated injection volume of 3.8×10^6 m^3. Producers estimate that, after 20 years of production, gas volumes are trapped on the southern flank of the field due to the progressive effect of water injection. In Fig. 11, the estimated water pressure value in this area is locally greater than the gas pressure value. The progressive trapping of gas reserves, initially thought to be induced by artificial water injec-

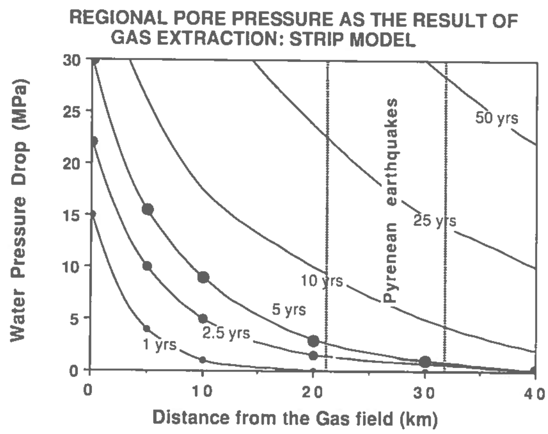

Fig. 10. Strip flow model. Pressure drop of the aquifer as a function of the distance from the gas field. Each curve corresponds to a time delay after gas extraction began.

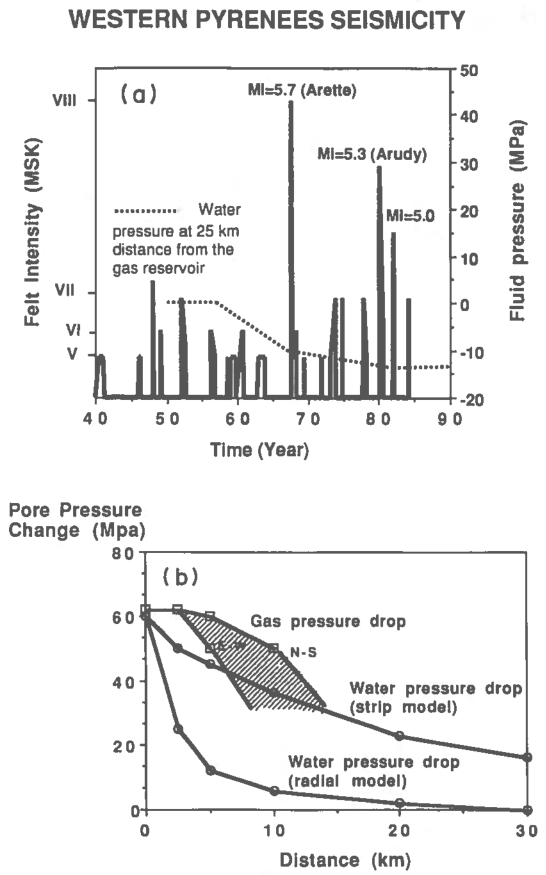

Fig. 11. (a) Historic seismicity of the western Pyrenees and the estimated pressure change on the North Pyrenean Fault (dotted line), estimated with the value calculated with the strip flow model. Zero value on the time axis corresponds to year 1940. Earthquake magnitudes are local magnitude. (b) Gas pressure drop and water pressure drop vs distance in the neighbouring region of the gas reservoir, after 25 years of extraction. Curves labelled with open squares are the gas pressure change within the gas reservoir deduced from well data. The two curves correspond, respectively, to E–W and N–S pressure profile. Curves labelled with open circles are the water pressure changes modelled by strip model and isotropic radial model. Intersection between water pressure drop (strip model) and depletion of gas reservoir can be interpreted as water inflow into the gas reservoir.

tion, could be indicative of natural water inflow. This possibility is added to show that the parameters used to compute water pressure change at great distance from the field are realistic even for the local reservoir behaviour.

ROCKSALT AS A VISCOELASTIC STRESS DIFFUSION CHANNEL IN THE UPPER CRUST

The mechanical connection between earthquakes by a modified Elsasser (1969) model was first proposed by Rice (1980) and further analysed by Lehner *et al.* (1981). Investigations at lithospheric scale were performed (cf. Thatcher *et al.* 1980, Rice & Gu 1983, Li & Kisslinger 1985, Melosh 1988, Linker & Rice 1991). Rydelek & Sacks (1990) discovered an apparent correlation between earthquakes in the subduction zone of northeast Japan and earthquakes in the overthrust block. The land earthquakes precede the events at sea by 36 years and their epicentres are located, on average, 200 km apart. Plate convergence causes the overthrust block to buckle, resulting in the occurrence of thrust events on land. The stresses released by the land earthquake slowly diffuse toward the subduction zone due to viscoelastic coupling between the brittle upper crust and the ductile asthenosphere. The diffusing extensional deformation pulls the overthrust block away from the subducting slab, thus reducing friction and triggering an earthquake. The extensional deformation also reduces the shear stress between the slab and the overthrust block.

The 'Elsasser'-like model which Rydelek & Sacks (1990) employ to investigate stress diffusion has one dimension of freedom with all displacements being horizontal. An elastic layer (30 km thick lithosphere) lies over a viscoelastic layer (30 km thick asthenosphere). The upper layer is displaced by an edge dislocation while the bottom of the lower layer is rigid. This basic first-order model is much simpler than the three-dimensional viscoelastic finite-element technique used to model stress transfer and deformation associated with large earthquakes, when geodetic measurements and fault plane geometry fix the limits of the model (cf. Linker & Rice 1991).

Use of an 'Elsasser'-like model is proposed here to test the possibility of stress transfer at upper crustal scale when moderate shallow earthquakes are used as an initial stress and strain pulse. If the elastic medium (sedimentary level within which shallow earthquakes occurred) lay on a viscous medium (rocksalt beds and décollement level below frontal overthrust in the vicinity of a mountain chain), wave diffusion is delayed, due to the coupling effect with the viscous material. If the diffusing strain wave encounters a locked fault, it can reduce the frictional force, when well oriented, thus unlocking the fault and triggering an earthquake.

As the strain pulse is propagating, its maximum value and velocity decrease with increasing distance from the source. The velocity of the strain pulse is equal to the distance x of a point from the source divided by the rise-time of the pulse at this point. The standard definition of the rise-time (T) is the time taken by the strain, at the distance x from the source, to reach 63% of its maximum value. The expression of strain as a function of x and t can be derived from the equations of Rydelek & Sacks (1990):

$$\varepsilon(x,t) = \frac{U_0}{\sqrt{\pi}\, H_e} \sqrt{\frac{\tau}{t}}\, e^{-x^2\tau/4H_e^2 t}$$

with u_0 the horizontal component of displacement at the source, and

$$H_e = \sqrt{\frac{Y_e}{\mu_v}(H_1 H_2)},$$

where H_e is the effective elastic thickness, Y_e is the Young's modulus of the elastic sedimentary cover, H_1 and H_2 are the thickness of the layers and $\tau = \eta/\mu_v$, ratio of viscosity to shear modulus of the viscous layer. It yields the rise-time T as a function of the distance x from the source:

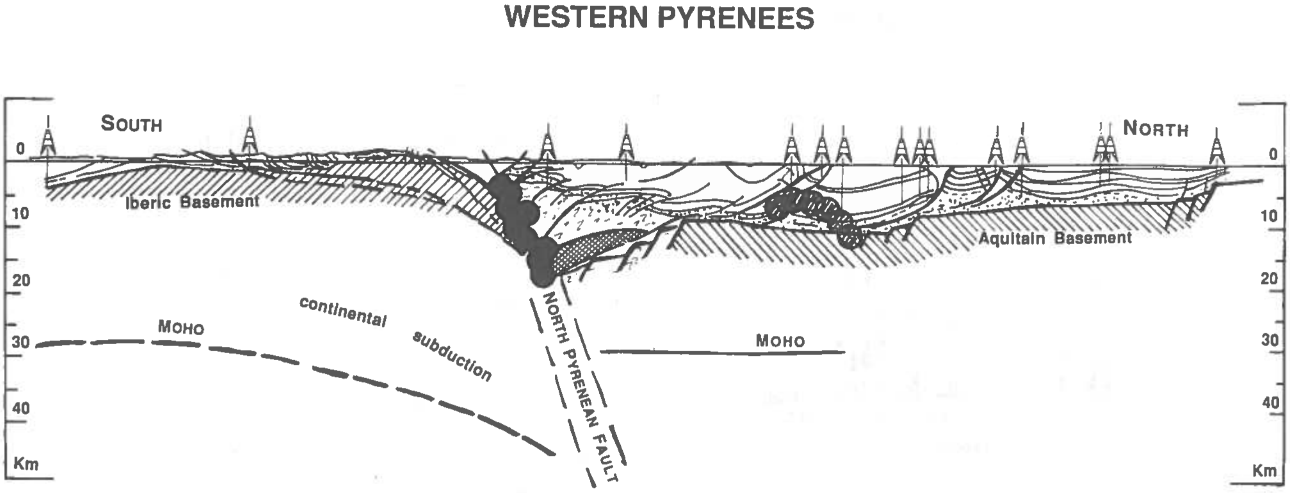

Fig. 12. Tectonic cross-section of the western Pyrenees area (adapted from Ecors Arzacq Program). Black dots are Pyrenees seismicity and hatched dots are induced events in the Lacq gas reservoir area.

$$T(x) = \frac{1}{6}\tau\frac{x^2}{H_e^2}.$$

Since

$$\sigma_{xx} = Y_e\varepsilon_{xx}$$

the corresponding stress drop for 1 m horizontal displacement component of the fault is:

$$\Delta\sigma(x) = \frac{1}{x}\frac{63}{100}Y_e\sqrt{\frac{2}{\pi\rho_E}}.$$

Using numerous borehole and seismic profiles a geomechanical cross-section of the western Pyrenees area was built (Figs. 7 and 12). The corresponding mechanical model is an elastic layer (8 km thick sediments) over a viscoelastic channel (0.2–2 km thick salt bed). This model raises several questions when applied to the upper crustal scale. Can a diffusive deformation in the upper crust stimulate the occurrence of a shallow seismic instability? Are faults between the site of the natural earthquakes and the gas field likely to be excited to rupture?

To check these questions, the $T(x)$ value is calculated at distance ranging from 10 to 60 km. The values of $T(x)$ and $\Delta\sigma(x)$ obtained (Fig. 13) must be considered only as orders of magnitude, since the geological structure is not perfectly horizontal, the evaporites probably have a variable thickness and also because the behaviour of the upper crust is not perfectly brittle (Gratier & Gamond 1990).

As the elastic sedimentary cover is mainly limestones, the Young's modulus Y_e is taken to be 6×10^{10} Pa.

The shear modulus of the viscous Triassic layer is calculated with the relation $\mu_v = 2Y_v/5$. Using $Y_v = 5 \times 10^9$ Pa for evaporites (Carmichael 1982), this gives $\mu_v = 2 \times 10^9$ Pa. The viscosity η of evaporites is considered as ranging from 10^{17} to 10^{19} Pa.s according to both experimental laboratory studies and numerical values used in geomechanics (cf. Last 1988, Spiers *et al.* 1990).

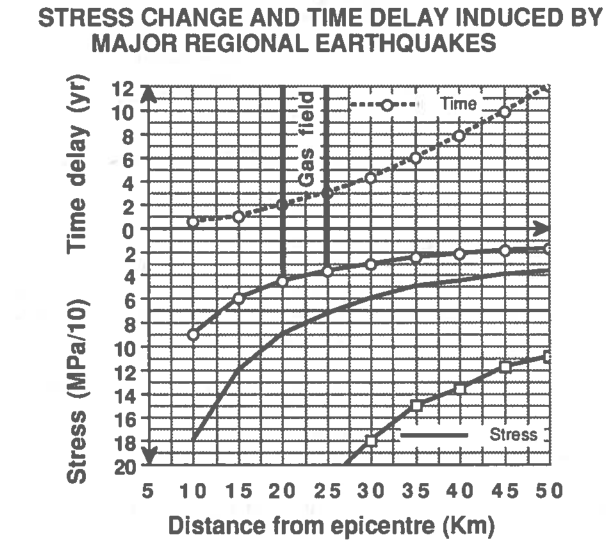

Fig. 13. Time delay (dotted line with open circles) and stress changes (continuous line with open circles) induced by stress transfer from the Arette (M_l = 5.7) Pyrenean events at different distances from the epicentre. A 0.5 m horizontal source displacement is used to model the M_l = 5.7 earthquake after-effects (open circle). The two other stress curves represent stress changes induced by a 1 m source displacement (thick line), and a 3 m source displacement (M ~ 7 event), (open square).

Using a salt viscosity of 10^{17} Pa.s (Last 1988, Spiers *et al.* 1990) and a 0.2 km thickness, the theoretical 2 years $T(x)$ diffusion time of the major Pyrenean earthquake at a distance ranging from 20 to 25 km is of the same order of magnitude as the 25-month time interval between the Arette Pyrenean earthquake and the first event in the Lacq area. Using a salt viscosity of 10^{19} Pa.s, the diffusion time is about 200 years for the same thickness of rocksalt bed. Using a viscosity of 10^{18} Pa.s and a 2 km thickness for the viscous bed, the same theoretical 2 years $T(x)$ diffusion time is obtained for the major Pyrenean earthquake after-effect to reach the Lacq gas field. Thus, within the range of plausible viscosity, the Lacq earthquake could be considered to have been triggered by the stress diffusion from the Pyrenean event. The estimated implied stress change is less than 1 MPa (Fig. 13) and the same order of magnitude as that deduced from the poroelastic stressing, locally induced by gas reservoir depletion (Segall & Grasso 1991). The onset of seismic instabilities in the Lacq area could thus be triggered by two effects. The first effect is a monotonic increase in poroelastic stressing proportional to the gas pressure drop (25 MPa depletion between 1959 and 1969), without any seismicity, resulting in an increasing shear stress around the reservoir of 0.2 MPa. The second stress change could be the Arette shock after-effect given on Fig. 13 which, in a shorter time period, gives an equivalent stress increase. Note that the same time delay is observed between the Arudy shock and the first deep seismic swarm (events below the gas reservoir) in the Lacq area. Segall & Grasso (1991) show that the poroelastic stress changes below the reservoir where the deep events have occurred since late 1982, are half those above the reservoir. The earthquake after-effect of the 1980 Pyrenean event, could therefore have the same role in triggering the deep Lacq seismicity as the 1967 Arette event for the shallow Lacq seismicity.

Using a salt viscosity of 10^{17}–10^{18} Pa.s and a 0.2–2 km viscous bed, a Pyrenean shock will generate a strain wave that affects the weakened fault at the gas field area 2 years later. Despite the absence of measurements of surface deformations resulting from strain wave migration, which would give an independent estimate of the viscosity, the estimated time delays show that major Pyrenean events and major Lacq field earthquakes can be correlated by a viscoelastic model, In this way, a strain wave from a Pyrenean earthquake should trigger the release of locally accumulated induced strain. The triggering model is not investigated in this preliminary study. If the mechanism proposed here has any veracity, further studies on other areas, where the upper crust is locally weakened either by man-made manipulations or by natural tectonic processes, will check the proposed connections and investigate in more detail the mechanism of triggering events using fault geometry analysis.

CONCLUSIONS

It has been shown that major regional events, locally induced events and subsurface fluid pressure fluctuations in the western Pyrenean area are temporally and spatially correlated. In a first phase, gas withdrawal induces a decrease of water pressure in the aquifer at the gas–water interface and then affects the natural regional seismicity rate. The locked fault inhibits small events for a time, during which tectonic stresses accumulate. Large events then occur on the Pyrenean Fault, when the energy of deformation is large enough to allow seismic instabilities to occur. These large events ($\mathbf{M}_l > 5$) trigger seismic instabilities ($\mathbf{M}_l = 4$), in the vicinity of the gas extraction area, 30 km away. As a result of gas extraction, the Lacq field zone has weakness characteristics similar to the neighbouring main active fault. The Pyrenean earthquakes lead the gas field events by 2 years. this correlation is consistent with a viscoelastic model, where salt rock acts as a viscoelastic channel, below an elastic bed.

If connections of tens of kilometres are confirmed, major tectonic features located in the vicinity of hydrocarbon fields can be affected by development-induced fluid flows and deformations. Major induced earthquakes should be found on such tectonic features located in the vicinity of fluid pressure fluctuations such as hydrocarbon fields or artificial water reservoirs. Since the role of salt in fold and thrust belts (cf. Davis & Engelder 1985) and the effect of fluids on deformation of rocks (cf. McCaig 1988, Carter *et al.* 1990) are accepted, the possible connections proposed here may give a new insight into understanding the role of fluids in seismic instabilities and seismic risk assessment in the vicinity of fluid pressure fluctuations, when natural seismic activity is high.

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