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# b-VALUE, ASEISMIC DEFORMATION AND BRITTLE FAILURE WITHIN AN ISOLATED GEOLOGICAL OBJECT : EVIDENCES FROM A DOME STRUCTURE LOADED BY FLUID EXTRACTION

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Abstract. Studies of b-values are usually performed either at rock sample scale (laboratory experiments) or at crustal scale (earthquakes). But interpretations at crustal scale are extrapolated from small scale experimental laws with well defined boundary conditions, to a larger object with no clear boundary conditions. We examine variations of the b-value in time and space at spatial (10x10x10 km$^3$) and temporal (20 years) scales intermediate between laboratory analyses and tectonic processes, in a dome structure that is an isolated geological object with defined boundaries and known geomechanical properties. Seismic activity (about 1000 events with magnitude $\leq$ 4.5) and aseismic displacements (6 -7 cm of cumulative subsidence) have been induced by gas extraction in an area where no displacement had previously been reported. We find no agreement between temporal variations in b-values and results from laboratory experiments : there is no correlation between b-values and stress histories, nor between b-values and the spatial migration of seismicity. Aseismic slips introduce anomalies in b-value behaviour when seismic instabilities are a second order process compared to the whole deformation. These observations imply that when changes in b values have been used for earthquake forecasting, false alarms can be explained by the occurrence of aseismic displacement.

## Introduction

In b-value studies made using acoustic emission (AE) in laboratory experiments the characteristic parameters (degree of sample homogeneity, stress increase, space - time development of microfractures, size of microfractures, etc.) are known. In natural conditions it is difficult, even impossible, to have a precise knowledge of these parameters, because the size of the studied object cannot be clearly defined. Forecasting of major events using changes in b-values are thus either empirical or ineffective.

For earthquakes, the frequency of occurrence is a log-linear function of the magnitude [Gutenberg and Richter, 1949], corresponding to a power-law distribution of seismic moment or fault length. Mogi [1962, 1985] found the same frequency-magnitude relationship for microcracks as for earthquakes. Mogi [1980] and Hirata [1987], investigating b-values of microcracks in rock samples, noted that the b-value decreases before macroscopic failure caused by a constant or even a decreasing external stress.

Scholz [1968] repeated Mogi's initial laboratory experiments on rock deformation, using a larger frequency range, and showed that b-value is inversely proportional to the stress applied to the rock sample. A study of b-values in the Denver hydrocarbon field by Wyss [1973] local earthquakes in the vicinity of a water injection well, is consistent with Scholz's results [1968].

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McGarr [1976] using magnitude-frequency relationship of mine tremors argued that there is no simple correlation between b-value and stress changes up to 100 bars. He suggested that a change of 500 bars or more would be necessary to obtain significant results in his case study.

Main *et al.* [1990] show that b-values in laboratory experiments is anticorrelated to stress intensity K [e.g., Lawn and Wilshaw, 1975], which combines stress and crack length. They propose that it is a critical value of K rather than that of the stress which determines the rupture time, using a damage theory based on fracture-mechanics.

In this paper we study a cluster of seismicity located in a place that was historically aseismic before the beginning of a perturbation in the stress field due to gas extraction. The gas field is located 30 km north of a regional area of seismicity defining a narrow east-west strip along the North Pyrenean Fault. Spatial and temporal distributions of earthquakes in the gas field are correlated with the gas field deformation and pressure history [Grasso and Wittlinger, 1990]. Because of the gas extraction, the underground geomechanics is well known (boreholes associated with oil and gas exploration, seismic profiles, preexisting faults, effective stress variations deduced from gas pressure drop, etc.). All these data create a situation close to laboratory experimental conditions, but on a larger scale. We analyse spatial and temporal variations in b-values behaviour during a 17-year period (1974-1990) and test the models discussed above at an intermediate scale between laboratory experiments and regional crustal observations.

## Space-time Patterns of Stress and Induced Seismicity

Seismicity began in the Lacq gas field in 1969, 10 years after gas extraction started. No historic seismicity had been reported here for at least several centuries [Grasso and Wittlinger 1990]. The local induced seismic activity is well defined, 30 km away from the strip of the regional seismicity in the Western Pyrenees. The hypocenter distribution of induced events, deduced from the local seismic network operating since 1974, confirms that there is no seismically active zone connecting seismicity induced by fluid extraction in the Lacq area and that occurring on the North Pyrenean Fault system. The envelope of the induced hypocenter locations mimics the dome structure within the gas reservoir stand [Grasso and Wittlinger, 1990 ; Guyoton *et al.*, 1992]. All these observations allow us to consider the 10x10x10 km$^3$ volume defined by the envelope of the induced seismicity as acting as a closed system. This assumption is enhanced by the results obtained from numerous levelling profiles conducted in the area [Grasso and Feignier, 1990]. These profiles exhibit a subsidence of few centimeters correlated in both space and time with the gas pressure drop (Figure 1), implying that the regional deformation is elastic to first order. At a smaller scale Grasso *et al.* [1991] have shown that the behaviour of the subsidence is consistent with aseismic slip on three major faults. The sizes of the induced seismic sources, estimated by spectral analysis of seismic waves, show that seismic instabilities have only a second order effect on the displacement (maximum dislocations < 2 centimeters, maximum radius < 300 meters) [Feignier and Grasso, 1992].

Two phases of seismic activity have been observed. From 1974 to 1982 most of the seismic activity occurred within the stiffest part of the overburden, with highly diffuse locations (Figure 2a) [see also Grasso and Wittlinger, 1990]. Since

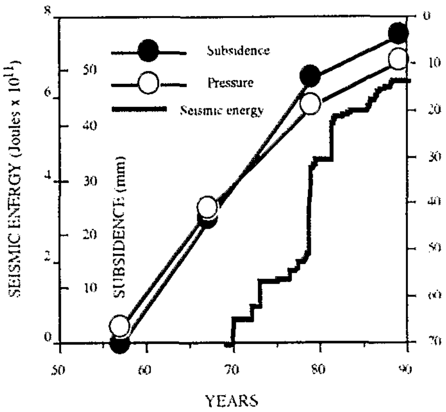


Fig. 1 : Temporal evolution of seismic energy, subsidence and gas pressure drop, since the beginning of gas extraction (1957). Subsidence is the maximum displacement measured at the top of the subsidence bowl. Gas pressure drop is averaged over several wells with a standard deviation ≤ 1 MPa.

1983, deep events have occurred below the reservoir and are more organized on large preexisting faults, defining several clusters (Figure 2b) [see also Guyoton *et al.*, 1992].

The forces applied at the boundaries of such a system are (1) tectonic forces and (2) gas reservoir pressure. Over a time scale of tens of years, we consider that tectonic forces are constant and that possible earthquake after-effects of regional seismicity are short-period phenomena [Grasso *et al.*, 1992]. Thus, the main perturbation of the medium over the 30-year time period of this study has been induced by the pore pressure decrease of the gas reservoir.

Stress changes in the overburden, as well as within the bed below the reservoir, were modeled using pororelastic stressing by Segall and Grasso [1991]. The basic result is that stress changes outside the reservoir (where seismic instabilities occur) are proportional to the pore pressure drop within the gas reservoir itself, despite the fact that there is no direct fluid connection between the reservoir and the overburden. The same critical stress threshold is necessary to explain both the rupture above and below the reservoir, except that due to the free surface a time delay to reach this threshold is two times smaller above the reservoir than below it. The calculated change in shear stress, which triggered shallow seismic activity in 1969 and deeper activity in 1983, is in both cases less than 1 MPa. Even if such stress changes are small, they are of the same order of magnitude as those recognized to drive seismic instabilities in the neighbourhood of artificial water reservoir [Roeloffs, 1988].

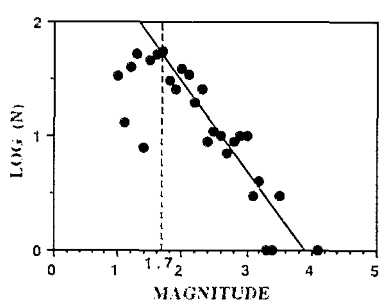


Fig. 2 : Temporal evolution of seismicity across the gas field. The events are plotted on vertical cross-sections. The coordinates of the end points of the cross section are indicated on both ends of the figure. (a) Activity from 1976 to 1982, [data from Grasso and Wittlinger, 1990]; (b) Activity from 1983 to 1989, [data from Guyoton *et al.*, 1992]. The distribution of the seismicity is diffuse during the first time period, while clusters appear during the second time period.

## Study of the magnitude frequency relationship

Since 1974, when a local network was installed, we have observed induced seismic activity and computed b-values in order to understand its physical meaning at a scale intermediate between (1) laboratory measurements on rock core samples, where each parameter is controlled, and (2) a large tectonic area where boundary conditions are difficult to isolate. The Lacq local network has operated continuously with only one interruption between July 1979 and December 1979. Starting in 1974, 4 stations were installed and 5 more stations were added between 1974 and 1979. Between 1974 and 1990, about 1000 induced events were detected with a local magnitude ranging from 1.0 to 4.2. Seismicity during the period 1974-1979, in numbers of both low and high magnitude earthquakes, was higher (while the network was smallest), than since 1979 (complete network). We can thus assume that during the first period no significant events were missed. The cutoff magnitude determined by the log-linear portion of the discrete frequency magnitude plot is 1.7 (Figure 3). About 500 events are selected according to this criteria.

As we have a magnitude range spanning less than 2 units, we used the general maximum likelihood estimation of b, where b'=b/log10 [e.g., Page, 1968] :

$$b = \log_{10} e \left[\overline{m}_l - \frac{m_{lmin} - m_{lmax} e^{-b'(m_{lmax} - m_{lmin})}}{1 - e^{-b'(m_{lmax} - m_{lmin})}}\right]^{-1}$$

We calculated the b-value using windows of N events. Windows of N events was preferred to windows of N months in which the difference in the number of events would give heterogeneous results because the seismic activity was not steady. We computed variations in b-values using several windows (between 60 and 100 events) and different increments (between 1 and the window value). Despite some high frequency variations, the general trend is the same. The smoothest variations in b-values were obtained with a window of 87 events. As we are interested in long term b-value variations, we chose this value for the study. Two distinct stages in the behaviour b-value are observed (Figure 4a and

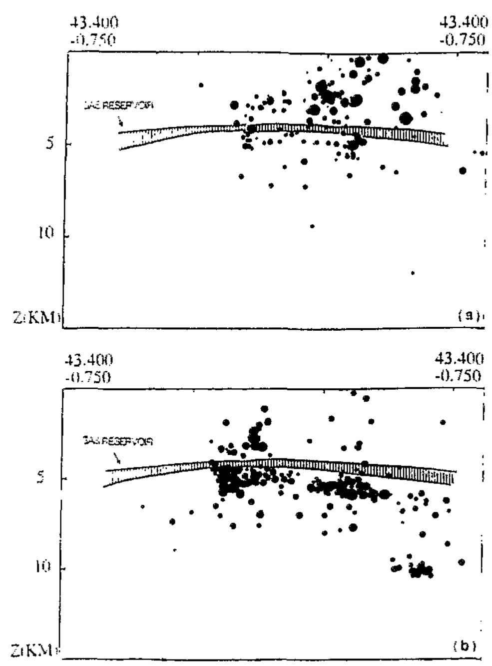


Fig. 3 : Discrete frequency distribution of magnitudes. The dotted line shows our cutoff magnitude ($M_l$ = 1.7).

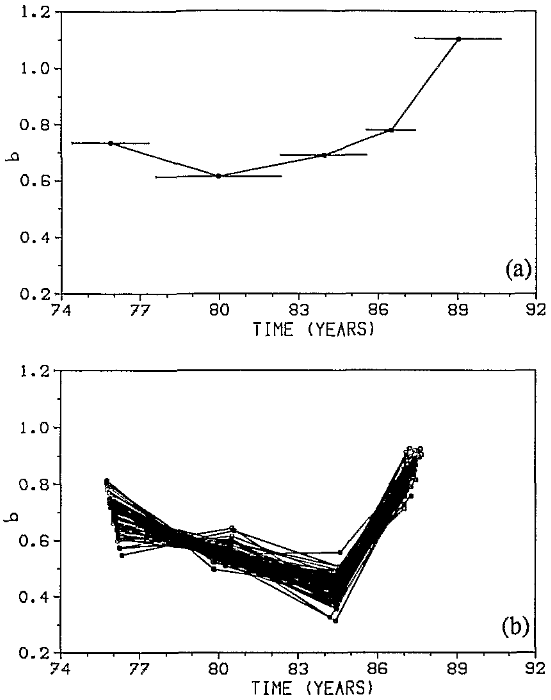


Fig. 4 : Temporal b-value evolution from 1974 to 1990. (a) using a window of 87 events, with an increment of 87 events. Horizontal bars represent the time necessary to obtain a sample of 87 events. (b) using a window of 87 events, with an increment of 87 events. The range of uncertainty on the magnitude is ± 0.1. To obtain the slope b, we added a random value ranging from -0.1 to 0.1 to the magnitudes.

Figure 4b) : a decrease between 1974 and 1982 followed by an increase from 1983 to 1990. The goal of this paper is to relate the b-value variations to the local stress and seismicity history.

## Discussion

The Lacq gas field area is considered as a well defined geological object, at a scale intermediate between a rock sample and a tectonic feature, with well defined boundaries and a recent seismological history. In this geological object five facts are observed : (1) a decrease of b-values between 1974 and 1982, followed by an increase from 1983 to 1990 ; (2) a decrease of the gas pressure between 1974 and 1990 that drives the induced stresses in rock above and below the gas reservoir, according to poroelastic stressing [Biot, 1941; Segall and Grasso 1991]; (3) a migration of the seismicity in depth starting in late 1982 - early 1983 ; (4) a change in hypocenter distribution, from diffuse to clustered and (5) a continuous subsidence starting in the 60's correlated with pressure drop, i.e. to stress increase. Mechanical modeling using the boundary elements technique shows that the shape of the subsidence profile can be controlled by aseismic slips on pre-existing faults [Grasso et al., 1991].

Observations (1) and (2) are inconsistent with Scholz's [1968] observations, which predict a continuous decrease of b-values with monotonically increasing stress. A possible explanation is that the stress increase is too small, although it is big enough to trigger and to sustain seismicity. There is thus a paradox : the motor of induced seismicity is pressure change (i.e. stress), but the observed b-value variations are not correlated to this key parameter.

As we observed a migration in the depth of the seismicity (observation 3), we have computed b-values separately for events located above ($z < 5$ km) and below the reservoir ($z > 5$ km). The differences between each data set are not significant. The observed b-value variations of the two data subset and that observed from the whole data set are the same. Thus, the contrast in geomechanical properties of rock matrices located above and below the reservoir, within which seismic fractures occur, does not affect the b-value. This is in contradiction to studies on rock samples [e.g. Mogi 1962, 1985], as well as Talwani's [1981] interpretation of b-values from earthquakes around an artificial reservoir, based on differences in rock properties.

Thus, given the first 3 observations, neither the increase of effective stress nor the rock mechanical properties can explain the temporal change in b-value behaviour.

The space-time evolution of the distribution of hypocenters exhibits two patterns (observation 4). During the first period (1974-1982), the seismicity is diffuse with neither aftershocks nor clustering, although the more energy is released during this period than later. In the second period (1983-1990) most of the seismicity in concentrated in several clusters. This change can be interpreted as pre-coalesence or coalescence of cracks. This process would imply a strong decrease in b, as predicted by studies of rock samples [Main *et al.* 1990]. This is opposite to what we observed.

Thus, the variations in b-values that we observed do not support published models obtained from rock mechanics experiments : we find no correlation with stress or stress intensity factor changes, nor with rock property changes. Nevertheless recent laboratory results [Main, personal communication 1992] propose that an increase in b-value might be associated with stable crack coalescence during experiments on wet core samples. They observed a stable b-value increased during crack coalescence and associated strain softening, due to a controlled drop in pore fluid pressure because of dilatant microcracking. The other difference between rock sample experiments and our study (the main difference is these variations in b-value) is the aseismic subsidence (observation 5) occurring simultaneously with seismicity [Grasso et al., 1991], while in rock mechanics experiments, acoustic emission from microcracks is mainly observed after elastic deformation is completed. Moreover, the subsidence is correlated with stress changes (figure 1) and accounts for a large part of stress release in the gas field area [Grasso and Feignier, 1990]. This could explain the discrepancies in b-value behaviour between rock experiments and our study. b-value variations alone would be insufficient to understand seismic instabilities because it does not include aseismic events, as also proposed, at a larger scale, by Robinson [1979]. In some case study b-value variations, deduced from seismic instabilitiy studies, a major event is predicted, but if this event is aseismic, a false prediction might result. On the other hand, when foreshocks of large seismic events (earthquakes) are mainly aseismic processes (slow and silent earthquakes), the b-value variations are inefficient to predict this kind of instability.

## Conclusion

Our study of b-value behaviour of induced earthquakes within an isolated geological object shows that laws deduced from rock sample experiments fail to explain our observations. Neither stress changes, although they trigger the seismicity, nor migration of seismic activity control b-value changes. A key factor to explain these discrepancies is the simultaneous occurrence of both aseismic slips and earthquake fractures. Depending of the mechanical behaviour of the studied area we can observe either aseismic foreshocks of earthquakes or earthquake fractures as foreshocks of large aseismic instabilities.

On the basis of both Scholz's report [1968] and observations of b-value decrease prior to large earthquakes [e.g. Suyehiro *et al.*, 1964 ; Li *et al.*, 1978], changes in b-values have been proposed as a precursor of major earthquakes, although correlation between b-value variations and occurrences of large earthquakes is not always observed. b-value interpretations for earthquake processes using results of rock sample analyses remain ambiguous if only stress interpretation is used, except of specific cases [e.g. Wyss, 1973]. In particular, false alarms of main earthquakes might be explained by occurrence of a main aseismic event. Therefore, in order to use variations in b-values to forecast major earthquakes in regions where significant aseismic deformation takes place, we should somehow include aseismic slip, slow earthquakes etc. in b-value calculations and certainly monitor other parameters ($Q^{-1}$, wich seems more simply related to creep than the b-value [Sato 1988; Jin and Aki, 1989]).

In our case study we now observe a clustering of seismicity, a decrease of seismic energy release and a locking of subsidence. From a classical seismological point of view this information can lead to a forecast of a major event. The question remains wether we know if the event will be seismic or aseismic. In addition, this information can be interpreted to forecast the end of gas production. In any case, the social and economical impact of these phenomena will be important : for instance for anticipating the shearing of wells in the first case (seismic or aseismic slip within the area where boreholes are situated), and for anticipating termination of the gas production and its social impact in the second case.

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