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Values of b and p: their Variations and Relation to Physical Processes for Earthquakes in Japan

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Synopsis

This work reviews some results obtained already for the variations of the seismicity parameters *b* and *p* in different seismogenic and tectonic regions in Japan. We bring as well new evidence that the time and space changes in seismicity parameters are correlating well with the crustal structure and/or some parameters of the earthquake process. In the first part of the paper we show that several seismicity precursors (clear *b*-value changes, quiescence and clustering) occurred about two years before the 1995 Kobe earthquake and they correlate well with other geophysical premonitory phenomena of the major event. The precursory phenomena occurred in a relatively large area, which corresponds probably with the preparation zone of the future event. In the second part, we analyze the *b* and *p* value spatial and temporal distribution for the aftershocks of the 2000 Tottori earthquake. The results indicate significant correlations between the spatio-temporal pattern of *b* and *p* and the stress distribution after the main shock, as well as the crustal structure. The swarm-like seismic sequences occurred in 1989, 1990 and 1997 showed significant precursory *b* and *p* values. In the third part of the paper we analyze the seismicity during the 1998 Hida Mountain earthquake swarm. The double-difference-relocated events are analyzed for their frequency-magnitude distribution and stress changes. While again the *b*-value is significantly different in south comparing with the north part of the epicentral area, the physical interpretation is difficult and complex. The changes in the Coulomb failure stress (ΔCFF) can explain the *b*-value distribution features, but the crustal structure may be also important. The seismicity distribution and migration, in relation with ΔCFF is also discussed. We refer as well to other world-wide studies.

Keywords: magnitude-frequency distribution, Omori law, seismicity, earthquake statistics, Coulomb stress changes, double-difference earthquake relocation

1. Introduction

The characteristics of the seismic activity in a certain region can give important and reliable information about the structure of the crust and the stress distribution associated with the earthquake occurrence. Besides, if the state of the stress or other physical parameters undergoes premonitory variations before a large event, the seismicity is, arguably, the "first candidate" which should show precursory changes, as it is the most obvious "product" of the stress distribution in a crustal volume. Closely monitoring the micro-earthquake activity, in particular, might be an effective tool to detect premonitory changes.

The frequency-magnitude distribution, which mainly expresses the relative amount of smaller to larger earthquakes through its "slope-parameter" *b*, has been extensively studied in laboratory experiments, as well as for synthetic and real seismicity. There are three main natural factors, as inferred from laboratory studies,

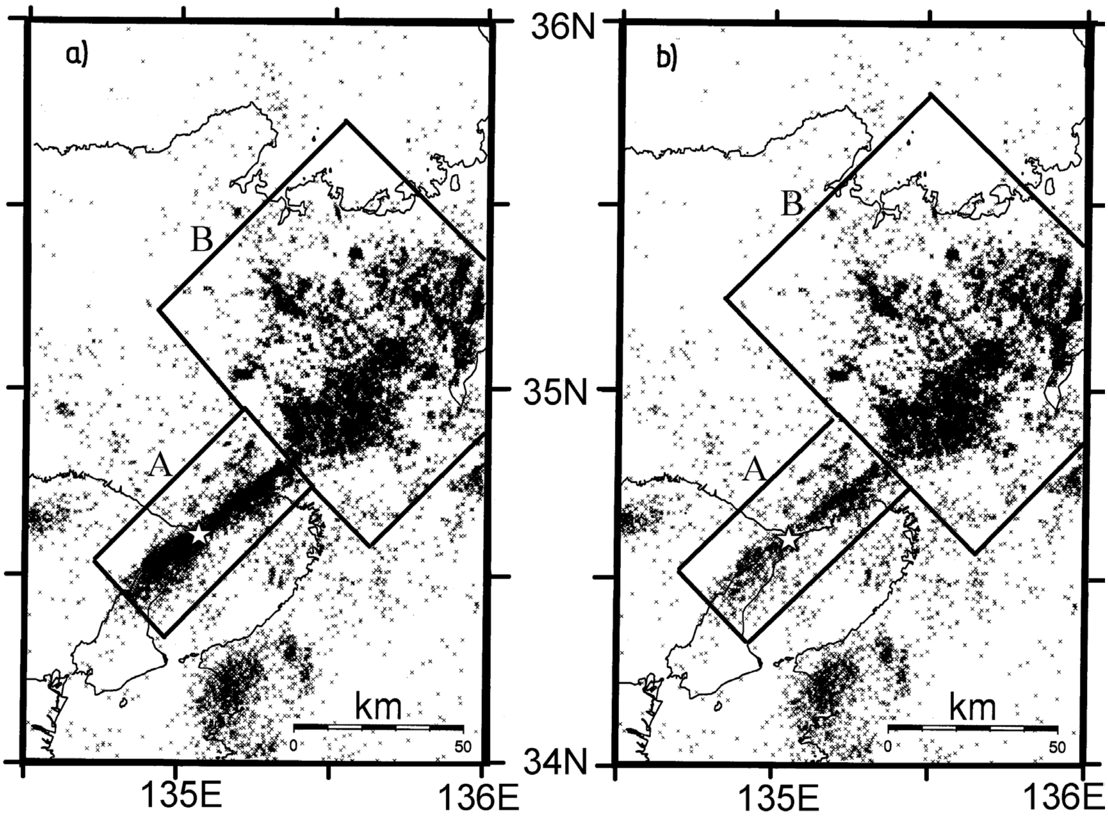

Fig. 1 Clustered (a) and undeclustered (b) seismicity (1976-1998, M ≥ 1.5) in a large region surrounding the epicenter of the 1995 Kobe earthquake. A – epicentral region, B – Tanba region. Star – Kobe earthquake.

that can cause significant changes of the frequency-magnitude distribution from an average, "normal" value of 1.0: a) an increased material heterogeneity increases the *b*-value (Mogi, 1962); b) a shear stress or effective stress increase decreases the *b*-value (Scholz, 1968; Wyss, 1973); c) an increase in the thermal gradient may cause an increase in the *b*-value (Warren and Latham, 1970).

In volcanic areas, high *b*-values are observed near magma chambers and highly cracked volumes; in creeping sections of faults high *b*-values are found, whereas asperities exhibit a low *b*-value (see a review, Wiemer and Wyss, 2002). The *b*-value is also found to decrease with depth in strike-slip fault zones (Mori and Abercrombie, 1997), probably as a result of the increased applied stress. There are as well many reports of *b*-value increase and/or decrease prior to large earthquakes (for example, Smith, 1981).

The occurrence rate of aftershock sequences in time is empirically well described by the Modified Omori law (Utsu, 1961): $n(t) = K/(t+c)^p$, where $n(t)$ is the frequency of earthquakes per unit time, at time t after the main shock, and K, c, p are constants. The characteristic parameter p, which is a measure of the decay rate of aftershocks, changes from 0.9 to 1.5 and its variability may be related to the structural heterogeneity, stress and temperature in the crust (Utsu et al., 1995).

The spatial distribution of well-located events can give important and reliable information about the fault structure, cracks distribution, earthquake migration and the state of stress in the crust. In sub-chapter 3.3 of this study we used the double-difference relocation algorithm in order to improve the accuracy of hypocentral parameters for the 1998 Hida Mt. earthquake swarm. The stress changes associated with the largest events in the sequence are determined and compared with the earthquake epicentral distribution.

2. Method

We first give a brief description of the methods and algorithms used in the paper and refer the reader to several references where one can find more detailed information.

For the computation of the relevant parameters in the frequency-magnitude distribution and Omori law, we used a maximum likelihood procedure (Utsu, 1965, 1992; Ogata, 1983). The probability that two different frequency-magnitude distributions come from the same population was determined by using Utsu's test (Utsu, 1992). The spatial (2D) distribution of *b*- and *p*- values was determined by superposing a dense grid of points on the seismicity map and computing *b* and *p* for each node of the grid by using the closest 100-200 earthquakes. The magnitude of completeness (M_C), i.e. the magnitude above which the earthquake catalog is complete, was checked for each node of the grid (Wiemer and Katsumata, 1999; Enescu and Ito, 2002a). For the determination of the temporal change of parameters a simple moving-window technique was applied.

As we mentioned already a double-difference relocation technique (Waldhauser and Ellsworth, 2000) was applied to the events of the 1998 Hida Mt. earthquake swarm. The method minimizes the residuals between observed and theoretical travel-time differences (or double-differences) for pairs of earthquakes at each station while linking together all observed event-station pairs. A least-squares solution is found by iteratively adjusting the vector difference between hypocentral pairs. We have used only absolute P-and S-wave travel-time differences, even the location method can incorporate also cross-correlation differential travel-time measurements.

In sub-chapter 3.3 we investigate as well the Coulomb stress change associated with the largest earthquakes during the 1998 Hida Mountain earthquake swarm. Our aim is to correlate the stress changes with the seismic activity and, finally, the *b*-value changes. In the Coulomb criterion, failure occurs on a plane when the Coulomb stress σf exceeds a specific value:

$$\sigma_f = \tau_\beta - \mu'\sigma_\beta \qquad (1)$$

where τ_β is the shear stress on the failure plan, σ_β is the normal stress and μ' is the effective coefficient of friction, which incorporates as well the effects of pore fluid pressure (e.g. King et al., 1994, Beeler et al., 2000). The change in the Coulomb stress is defined by the difference between the Coulomb failure stresses after and before an earthquake.

For computations we used the computer programs: Zmap (Wiemer, 2001), the Coulomb failure stress modeling tools GNStress 1_5 (author: R. Robinson), CFF (author M. Hashimoto) and DLC (author B. Simpson). Some of the routines we wrote or adapted for aftershock analysis (Enescu and Ito, 2002a) are available in the new release (version 6) of the seismicity analysis software Zmap. A new toolbox for stress analysis, written in Matlab and integrated with the Zmap software, will be available in early August from B.E. web page.

3. Characteristics of b and/or p-values for three recent earthquakes in Japan

3.1. 1995 Kobe earthquake: seismic quiescence, b-value and fractal dimension premonitory changes. Correlation with other geophysical precursors.

Enescu and Ito (1998, 2001) and Ito and Enescu (2001a) present in detail the DPRI catalog analyzed for significant seismicity changes before the 1995 Kobe earthquake. The catalog consists of 19604 events with $M \geq 1.5$, occurring in the area delimited by 34^0-36^0N and 134^0-136^0E, from 1976 to 1998 (fig. 1). The catalog is firstly declustered (aftershocks are removed), by using a method by Resenberg (1985). Then, the completeness of the catalog is checked by using a method of Rydelek and Sacks (1989) based on a day-to-night modulation scheme.

By analyzing the declustered catalog, Enescu and Ito (1998, 2001) found a significant quiescence anom-

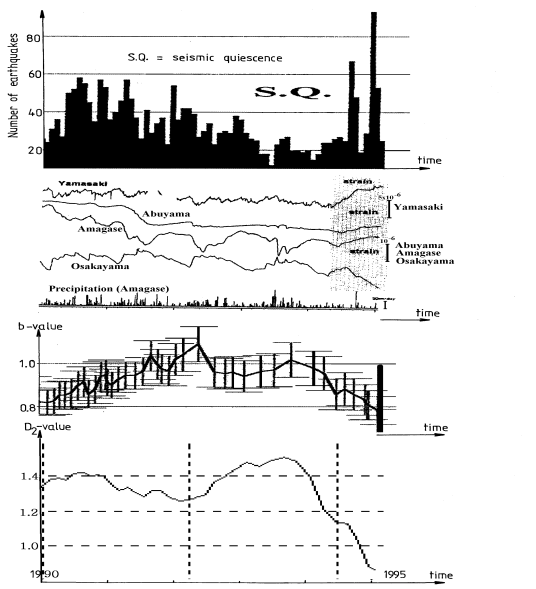

Fig. 2 From up do down: seismicity rate changes, with a clear seismic quiescence pattern (SQ) followed by an increase in seismic activity (ISA), several months before the Kobe earthquake; b) strain variation and precipitation (DPRI); c) b-value changes: increase and decrease before Kobe earthquake; d) changes in the fractal dimension (D_2) of the epicentral distribution of earthquakes: large decrease associated probably with ISA, correlates well with the decrease in b-value.

aly, which started about 2 years before the main shock, followed by a significant increase of the seismicity rate, several months before the 1995 large event. The *z*-value statistical test, which can quantify a relative decrease or increase of the seismicity rate (Habermann, 1983), was applied to infer the spatial pattern of the quiescence anomaly, by using a gridding technique. It was found that the quiescence anomaly occurs in a relatively large region surrounding the epicenter of the main shock as well as in the source area. The quiescence anomaly is clearly seen, in particular in the Tanba region (fig. 1). The analysis of the declustered catalog revealed as well that the background seismicity rate in the Tanba region, situated NE from the aftershock region of the 1995 Kobe earthquake, increased about 6 times after the large event, probably due to static stress triggering (Hashimoto, 1996; Toda et al., 1998; Nakamura, 1999).

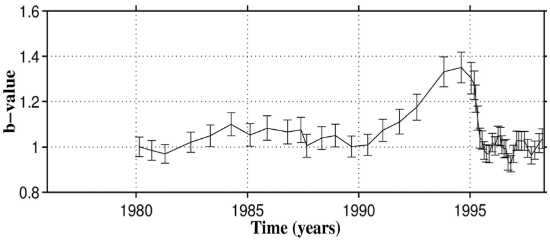

Fig. 3 b-value change in the Tanba region. The graph shows that before the large event the b-value in the Tanba region had a very clear increase. The b-value after the occurrence of 1995 Kobe earthquake returns to an average of 1.0.

Because by applying the declustering procedure (required for the application of the *z*-value statistical test) we may have eliminated as well some useful premonitory information, the original catalog (undeclustered) was also analyzed. The *b*-value and fractal dimension, *D2*, variation with time showed clear premonitory patterns, while the quiescence anomaly is seen as well. Figure 2 shows the time evolution of the seismicity rate, *b*-value and fractal dimension, five years before the main shock. One can easily recognize the discussed precursors. The variation of strain and precipitation with time (DPRI, Kyoto University) are also presented in Fig. 2. One can see that the precursory variation of seismicity parameters corresponds well to an increase in stress at the crustal deformation observation stations Yamazaki, Abuyama and Amagase. Ito and Enescu (2001a) discuss other reports of long-, intermediate- and short-term geophysical precursors occurred before the 1995 Kobe earthquake.

Figure 3 presents the evolution of *b*-value, from 1976 to 1998, in Tanba region, situated NE from the main-rupture area of the 1995 Kobe earthquake. A clear increase in *b*-value started about 2.5 years before the major event. This increase in *b*-value suggests that the quiescence anomaly discussed above is also an “energetic anomaly” (a relative increase in the number of smaller-size events, and relative decrease in the number of larger ones). As one can notice, *b*-value “recovers” after the l995 Kobe earthquake.

3.2 2000 Western Tottori earthquake – spatial variation of b- and p-values and its correlation with the stress changes and the crustal structure.

Enescu and Ito (2002a) studied in detail the distribution of *b*- and *p*-values for the aftershocks of the 2000 Western Tottori earthquake. Firstly, they have divided the aftershock area in several sub-regions, where the *b*- and *p*-values have been computed. The values of the parameters are significantly larger near the epicenter, comparing with the north and south ends of the aftershock region. This significant spatial variation of both *b*- and *p*-values is more clearly seen when the parameters are mapped at smaller scales, by using the gridding technique. Figures 4a and b maps the spatial variation of *b*- and *p*-values, respectively. One can notice that besides the regional features, the *b*- and *p*- values show also important “local” variations. Figures 5a and b compare the *b*- and *p*-values, respectively, in some selected areas characterized by small or large values of the parameters. The graphs themselves and several statistical tests suggest that the difference in *b* and *p* is significant.

The rather high *b*- and *p*- values found in the main shock area can be interpreted in terms of stress changes associated with the main shock and the 1989, 1990 and 1997 seismic swarms (Enescu and Ito, 2002a). Besides, as the spatial analysis did not include the very beginning of the aftershock sequence, when the catalog incompleteness was relatively high, the three M4.5 aftershocks occurred immediately after the main shock, in the main shock region, are probably also responsible for the higher values of *b* and *p* observed in the epicentral area (Fig. 4). The reduced stress and/or increased fracturing due to the occurrence of all these larger events could explain the higher *b* and *p* values. The lower *b*- and *p*-values, on the other hand, are found in areas that did not experience significant rupture. The NW region

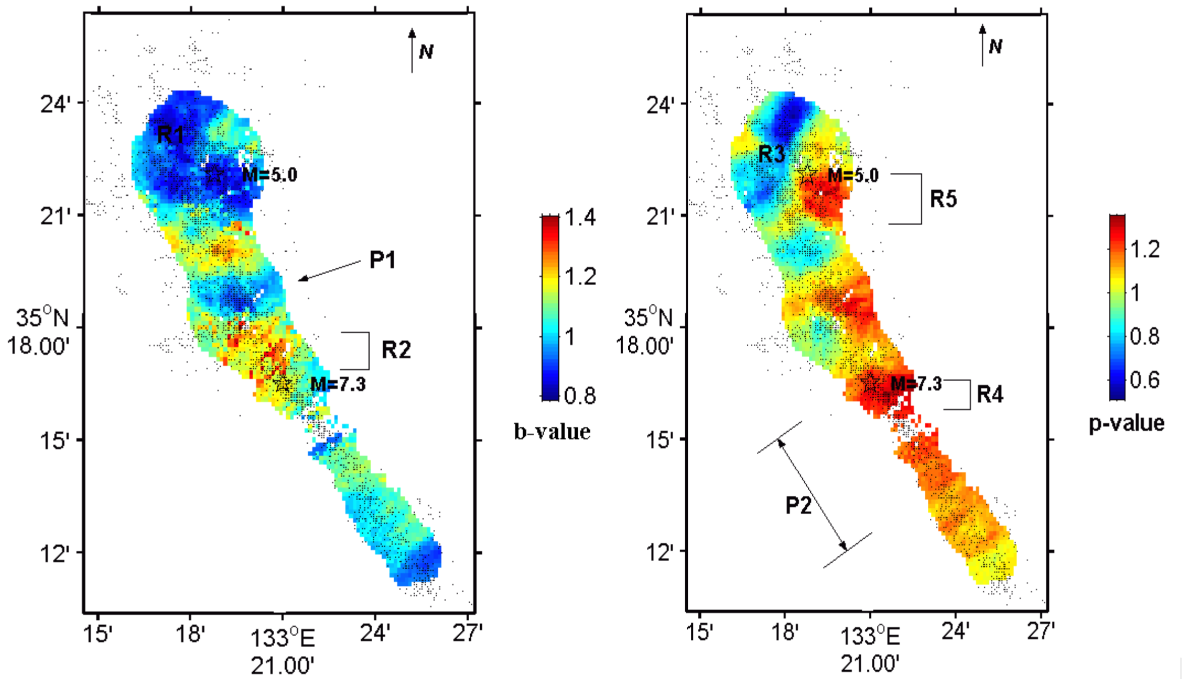

Fig. 4 a) b-value map for the aftershocks of 2000 Western Tottori earthquake. One can notice that the b-value is larger (≥ 1.0) for a large area extending about 10 km on each side of the main shock. There is an exception, the patch P1, where b-value is smaller. The both ends of the aftershock region (NW and SE) are characterized by a small b-value. In order to demonstrate the difference of b-value clearly we have chosen two regions (R1 and R2) and compute their frequency-magnitude distribution. The result is presented in the next figure.

b) p-value map for the aftershocks of 2000 Western Tottori earthquake. The area P2 shows large values of p. The area R5, near M5.0 aftershock epicenter, has also large p-values. There is a relatively large region in NW characterized by small p values. The areas R3, R4 and R5 were tested for the difference in p-value.

did not rupture for more than 20 years before the occurrence of the 2000 Tottori earthquake. Similar results for different earthquakes in Japan and the United States were obtained by Wiemer and Katsumata (1999).

As shown by Ito and Enescu (2001b, 2002a) the heterogeneous structure of the crust can offer an alternative explanation for the spatial patterns of *b* and *p* values. The heat flow distribution is shown to be an important element that controls seismicity.

Shibutani et al. (2002) and Enescu and Ito (2002a) showed that the *b*-value for the earthquake swarms in 1989, 1990 and 1997 was relatively low (around 0.6). This observation suggests an increased stress in the area where the future large event occurred. Moreover, Enescu and Ito (2002a) found that the *b*-value has a significant increasing trend, from values of about 0.6 for

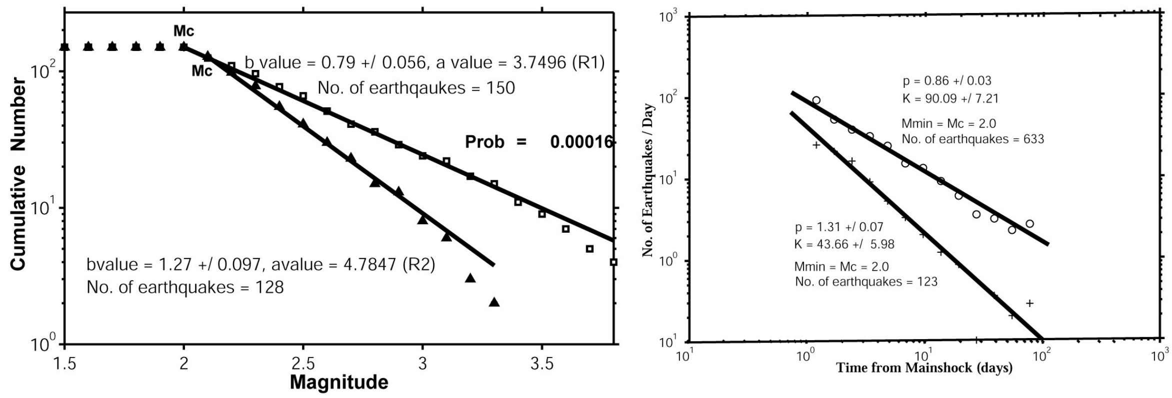

Fig.5 a) Comparison of the frequency-magnitude distribution for the Region R1 and R2 in Fig.4a. The probability, Prob, that the two distributions come from the same population is very small, as indicated by Utsu's statistical test (Utsu, 1992). b) Comparison of p-value for the regions R3 and R4 in Fig.4b. The decay of aftershocks in the two regions is significantly different. We have obtained a similar result, when comparing the R3 and R5 regions.

the first swarms to 1.3-1.4 after the Tottori main shock, in the main-rupture area.

3.3. The 1998 Hida Mountain earthquake swarm: double-difference event relocation, frequency-magnitude distribution and Coulomb stress changes.

The discrimination between different physical mechanisms responsible for the occurrence of earthquakes in a certain region is certainly a difficult task. Anyway, a reliable answer depends heavily on the quality and accuracy of the data to be analyzed.

The 1998 Hida Mountain earthquake swarm started on August 7 and its most active period lasted for about two months. In this period, 18 shocks with $M \geq 4.0$ occurred, the largest one ($M = 5.0$) being recorded on August 16. The swarm started in the south and migrated toward the north, reaching the northern most point in about one month. Most of the largest events in the sequence (fifteen) occurred during this south to north-migration period. For the following month, an opposite, north to south-migration of seismicity was observed. The average speed of migration was of about 1-2 km/day (Wada et al., 2000).

As a first step in analyzing our data we have relocated the events of the swarm recorded by DPRI (Kyoto University), by using the double-difference relocation algorithm (Waldhauser and Ellsworth, 2000). The velocity structure used for this analysis is the same as the one used to determine hypocenter locations at the Kamitakara observatory, at distances of about 30 km from epicenters. Both P- and S-wave catalog travel time data were used. Figure 6 presents the epicentral map of the relocated events of the sequence having $M \geq 1.7$. The estimated horizontal location error is less than 200m, while the depth error is less than 600m. The events occurred in the south (*S* region in Fig. 6), during a period of intense observation (August 25 – November 10), are most accurately determined, due to a better coverage with seismic stations (27 stations, in total). By examining fig. 6, one can notice two main directions defined by the epicenters: one oriented N-S, corresponding to the main direction of earthquake migration, and another one oriented approximately E-W; both of them probably reflect the directions of the regional stress.

Several studies (Enescu and Ito, 2001, 2002b; Sakai et al., 2002) found a relatively large *b*-value in the south

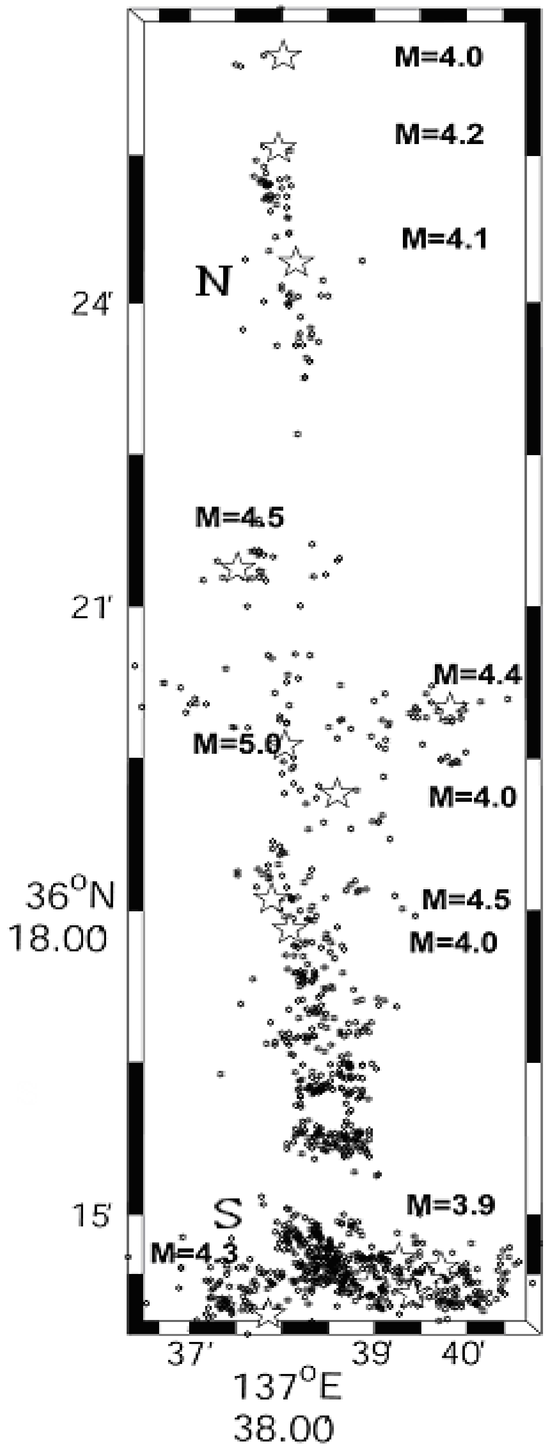

Fig.6 Relocated seismicity for Hida Mt. earthquake swarms ($M \geq 1.7$). N and S refer to the north south regions that are compared for their frequency magnitude distribution in Fig.7. Star show events with M>=4.0.

part of the Hida Mt., near the Yakedake volcano, comparing with the northern part. Besides, in our previous studies, we found an increase of *b*-value with depth.

By using the relocated events, we have computed the *b*-values for the north and south clusters, indicated by N and S in fig. 6, during the campaign of intense

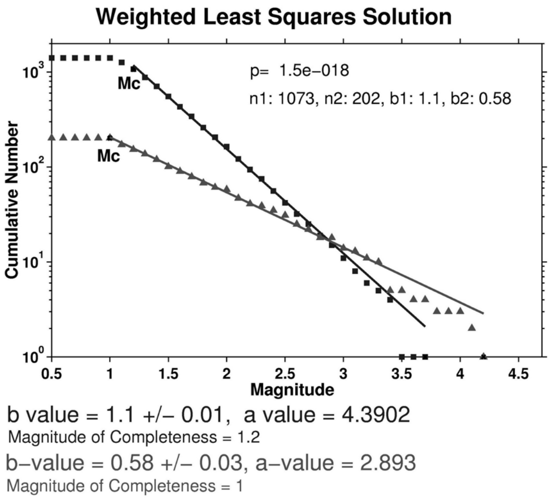

Fig. 7 Frequency-magnitude distribution in the north (*N* region in fig. 6) and south (*S* region in fig. 6) of Hida Mountains. The small p indicates here the probability that the two distributions come from the same mother population, by using Utsu's statistical test (1992). Squares – south region; triangles – north region.

observation (August 25 – November 10), when both the location and the magnitude of completeness are good enough. The result of the analysis is presented in fig. 7. As one can notice, the two frequency-magnitude distributions are significantly different, with particularly small *b*-values in the north, where three earthquakes larger than or equal to 4.0 occurred. Similar significant differences between north and south were obtained when using JMA (Japanese Meteorological Agency), NIED (National Inst. for Earthquake Disaster) and JUNEC (a joint catalog of several research institutions in Japan). Therefore, we believe that the difference in b-value between south and north is highly reliable. On the other hand the significant variation of *b*-value with depth found in our previous study could not be confirmed. The previous result, showing a large *b*-value (1.5-1.7) at depth (4-7 km), near Yakedake volcano, was probably due to the miss-location of small-size events. In fact, Enescu and Ito (2001) recognized this possibility and suggested the relocation of events, for a reliable depth determination.

In order to understand better the differences in the frequency-magnitude distribution, we analyzed also the *b*-value change with time in the south part of Hida Mt., from the beginning of the swarm until the end of 1999. Because this time we are not interested in a precise location, we have used the catalog data by Kyoto University, DPRI. Figure 8 presents the time variation of *b*-value for the south seismicity cluster. One can notice that *b*-value is varying significantly with time: small values (≅ 0.7) at the very beginning, followed by a relative increase and then, after about 2.5 month a decrease toward a stable b-value of about 0.8. Fig. 9 shows the frequency-magnitude distribution for the first 120 events occurred in the south cluster. The well-defined frequency-magnitude graph suggests that the small *b*-value observed at the beginning of the swarm is reliable and not a consequence of magnitude incompleteness. Due to the relatively small number of events occurred in the north (N) cluster, a similar temporal analysis could not be performed.

The static stress changes associated with the 18 largest ($M \geq 4.0$, JMA data) earthquakes in the swarm may offer important clues for understanding the rather complex *b*-value spatio-temporal "puzzle". Aoyama et al. (2002) performed an in-depth analysis of the static and dynamic stress changes for the 1998 Hida Mt. earthquake swarm. We do not intend to repeat their detailed analysis, but just to reveal some characteristics of interest for the present study.

We have considered the focal mechanism of the largest 18 events of the swarm and infer their source extent and average slip by using the relationship between focal dimension and seismic moment (Kanamori and Anderson, 1975), in a similar manner to Aoyama et al. (2002). The fault planes of the largest events of the swarm are oriented in two main directions, EW or NS. Assuming that these are the directions of the optimum failure planes, one can estimate the effective coefficient of friction, μ' (Aoyama et al., 2002, Iio, 1997). For this study the coefficient of friction was found to be about 0.2. This value can be interpreted as an argument for a weak fault-zone, but only a detailed analysis (e.g. Iio, 1997) can offer a reliable explanation. As it is known the small value of μ', could be also a consequence of a large fluid pore pressure.

Figure 10 presents the Coulomb stress changes, calculated on optimum failure planes, as a result of the 18

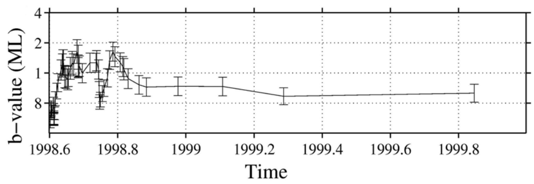

Fig. 8 Variation of b-value with time, for the south region (*S* in figure 6) of Hida Mt., when using all the earthquakes with $M \geq 1.2$ located by DPRI, Kyoto University. There are 200 events in a moving window.

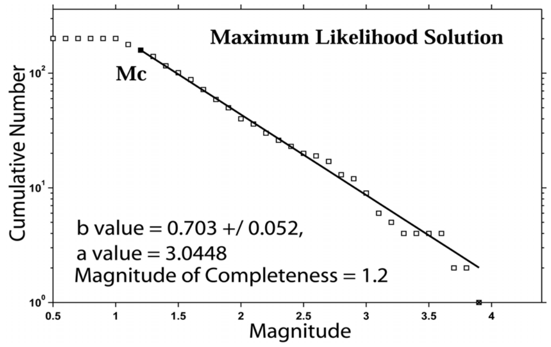

Fig. 9 b-value obtained by the maximum likelihood method for the very beginning of the swarm sequence in the southern cluster (*S* in figure 6).

largest earthquake in the sequence. We considered a general case, without any "a-priori" assumption made about the optimum failure planes or type of faulting. The orientation of the regional principal stresses is proved to be an important element to control the pattern of the stress changes. We have chosen a maximum compressional stress oriented N35^0 W, which is in good agreement with observation data (Wada et al., 2000). The relocated events with M ≥ 2.5 were superposed on the Coulomb stress map; due to the main south-north migration of seismicity the superposition shows correctly the correlation between seismicity and stress changes. About 80% of earthquakes occurred in regions with a positive stress change. The probability to obtain this by chance is less than 0.1.

As the swarm progressed from south to north, the stress changes advanced as well in the same direction. The small values of *b* could be well explained by an

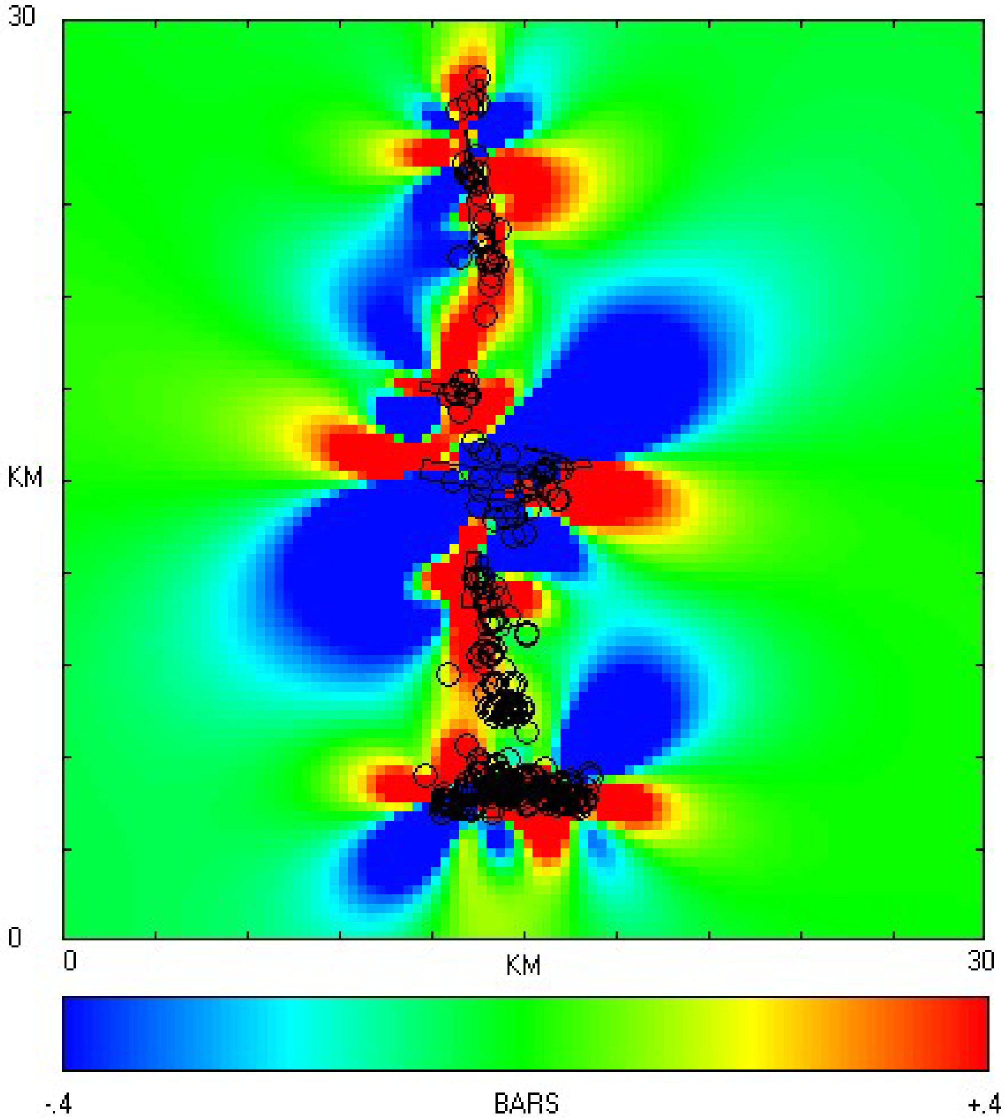

Fig. 10 Coulomb failure stress change on optimum oriented failure planes, as a result of the largest 18 events with M ≥ 4.0 occurred during a time period of about two months since the beginning of the Hida Mt. swarm. The red and blue colors indicate an increase or, respectively, a decrease in Coulomb stress. The regional stress has a WNW-ESE direction. Double-difference relocated events (M ≥ 1.7) are superposed on the stress map.

increased stress at the beginning of the earthquake activity in the south or north part of the swarm area. The release of stress was associated with a gradual increase of *b*-value for the south cluster. Thus, it seems reasonable to assume that the changes in stress may explain, at least in part, the *b*-value changes. On the other hand, the structure of the crust can not be excluded as a significant factor for the frequency-magnitude changes in the Hida Mt. region.

4. Conclusion

Our studies revealed significant variations in *b* and *p* values in different tectonic regions in Japan. Generally we observed that the lower values of *b* and *p* are correlating well with regions under increased stress or decreased crustal heterogeneity, as the laboratory studies predict. A high *b*-value may indicate an increased material heterogeneity.

In the case of the 1995 Kobe earthquake, we found good evidence of *b*-value precursory change, starting 2-3 years before the main shock. We pointed out that this precursory change correlates well with several other precursors.

The *b*- and *p*-value distributions for the aftershocks of the 2000 Western Tottori earthquake showed significant spatial changes, with larger values close to the main shock and with smaller values especially in the northern part of the aftershock area. The stress changes and/or crustal structure can explain the observed spatial pattern. The 1989, 1990 and 1997 seismic swarms, associated with moderate seize events, are characterized by a relatively small *b*-value, which increases with time. The *p*-value for these swarms follows a similar increasing trend.

In the case of the 1998 Hida Mountain earthquake swarm, a significant difference in *b*-value was found between north and south clusters. Precise earthquake location and the computation of the Coulomb stress changes associated with the largest events of the sequence were used to reliably interpret the seismicity pattern and *b*-value changes.

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ｂ値とｐ値，その変化と地震の物理過程との関連

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要　旨

最近の日本における３つの地震群に対して解析したｂ値およびｐ値についてのレビューを行い，さらに，新しい解析結果を付け加えて，それらのパラメータと地下構造および震源過程との関連について調べた。1995 年兵庫県南部地震については，約２～３年前からｂ値が変化したが，この変化は他の観測地の変化と対応している。また，この前兆的なｂ値の変化は震源域だけでなく，広い範囲で観測された。2000 年鳥取県西部地震については，余震のｂ値とｐ値は本震付近で大きく，余震域の北部で小さい。この変化は地域的な構造の相違か応力変化のいずれかだと思われる。また，1989, 1990，1997 年に起きた前駆的地震活動ではｂ値は 0.6 程度で小さかった。これらのｂ値はｐ値とともに時間とともに増加した。1998 年飛騨山地の群発地震については，相対的震源決定を行い，深さを含めて地域的なｂ値の変化を調べた。さらに，その結果，ｂ値は山脈の北部で小さく南部で大きい。また，18 個の大きな地震についてクーロン応力を計算したが，地震の 80%は応力が増加した地域で発生している。ｂ値の変化は応力変化と対応するように思われる。

キーワード：マグニチュード頻度分布，余震についての大森公式，地震活動，地震統計，クーロン破壊関数，相対震源決定