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Temporal Change of Q_s^{-1} in the 1999 Chi-Chi Earthquake Fault Area, Taiwan

Kuang-Jung Chen[1,*], Kevin Chao[1], Yung-Tang Tseng[1], Cheng-Horng Lin[2], and Chien-Hsin Chang[3]

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ABSTRACT

On the basis of magnitude deviation analysis, we investigated the temporal change of Q_S^{-1} in the 1999 Chi-Chi earthquake fault area in central Taiwan, by analyzing the seismograms of events occurring in a volume in the focal and aftershock areas. The temporal change in attenuation intensity around the fault area was measured. Attenuation change was estimated by analyzing the deviations between the station magnitude and the average local magnitude. The station magnitude M_L was calculated (earthquake report of Central Weather Bureau) from the simulated Wood-Anderson seismogram at each station and the average local magnitude was the mean of the station magnitude at the stations near the main shock. The results show that the deviations decreased at five stations, increased in one station in the vicinity of the focal area after the main shock compared with that before. Q_S^{-1}, derived from magnitude deviation analysis, in the upper crust in the close vicinity of the focal region increased during 1 day to 4 months following the main shock and decreased to its normal value after then. The simplest interpretation of the results is that the attenuation intensity in the upper crust in the close vicinity of the focal region increased after the main shock for frequencies from 4 to 25 Hz: $\Delta Q_S^{-1} = 1.378 \times 10^{-2} \times f^{-1}$.

(Key words: Chi-Chi earthquake, Attenuation, Q_S^{-1})

1. INTRODUCTION

Temporal changes of attenuation intensity in the vicinity of focal area - of interest to seismologists for the past decade - are of particular importance to better understanding of the change in rock properties during the whole process of preparation and generation of an earthquake. However, such a study is difficult to carry out because the foreshock sequence

[1]Dept. of Earth Sciences, National Taiwan Normal University, Taipei, Taiwan, ROC

[2]Institute of Earth Sciences, Academia Sinica, Taipei, Taiwan, ROC

[3]Central Weather Bureau, Taipei, Taiwan, ROC

[*]*Corresponding author address*: Prof. Kuang-Jung Chen, Dept. of Earth Sciences, National Taiwan Normal University, Taipei, Taiwan, ROC; E-mail: kjchen@cc.ntnu.edu.tw

cannot be observed on routine coarse networks and the portable array cannot be installed prior to the main earthquake in the absence of accurate predictions. Observations for the aftershock stage can be easily and accurately made.

Recently, temporal changes in attenuation intensity have attracted increasing attention for both observational and theoretical seismologists as a useful tool for earthquake prediction (Aki 1983). Gusev and Lemzikov (1985) found that the decay rate of S coda amplitude at stations near the aftershock area became larger than usual for a 1-year period preceding the main large earthquake in Kurile and the Kamchatka earthquake (M_s=8.1) of 1963; the Ust-Mamchatsk earthquake (M_s=7.8) of 1971; and the Iturup earthquake (M_s=8.0) of 1978. Jin (1981) discovered that S coda durations at stations close to the aftershock area were shorter than usual for earthquakes preceding the Tanshan earthquake (M_s=7.8) of 1976 in China. Jin and Aki (1986) found that coda Q^{-1} in the close vicinity of the main shock in a 3-year period preceding the Tangshan earthquake was about 3 times larger than before or after that period. Sato (1986) made a more detailed analysis for the temporal change in attenuation intensity associated with a crustal earthquake of magnitude 6 in the vicinity of the main shock. The attenuation intensity after the main shock was found to have become smaller than that before the main shock on the basis of the deviations of station magnitude. He also found an increase of S coda duration after the main shock at stations close to the main shock epicenter.

Some studies based on spectral analysis have shown that drastic changes in attenuation seem to be associated with seismic activity. Some of the papers so far published report a decrease in Q^{-1} for the aftershock stage (Novelo-Casanova et al. 1985; Jin and Aki 1986; Sato 1986). On the other hand, stronger attenuation for the aftershocks in the vicinity of the main shock is reported by Tsukuda (1985) and Peng et al. (1987). As for the preseismic stage, both an anomalous increase (Gusev and Lemzikov 1985) and an anomalous decrease (Sato 1987) of attenuation intensity are reported based on analysis of coda decay. According to the review by Gao (1985), both increase and decrease of Q^{-1} are reported for Chinese earthquakes of magnitude 4.7-7.8. Reliable field studies are still needed for getting more insight into the attenuation change.

In the present study, we investigate the temporal change of Q_S^{-1} associated with the Chi-Chi earthquake in Taiwan of magnitude 7.3 in the vicinities of the main shock and the aftershock, by measuring the deviation of the station magnitude. The attenuation intensity after the main shock was found to have become smaller than that before the main shock on the basis of deviation of station magnitude from the average magnitude calculated by routine observations.

2. DATA

An earthquake of magnitude 7.3 (M_L=7.3, M_W=7.7) occurred on 21 September 1999, near the small town of Chi-Chi in Nantou County in central Taiwan (Shin 2000). The epicenter of the main shock was located at 120.82ºE, 23.85ºN with a focal depth of 8.0 km. The focal mechanism was a thrust type with strike 5º, dip 34º and rake 65º (Chang et al. 2000). The distribution of aftershocks extends about 40 by 100 square kilometers horizontal and from 15 to 25 km in depth (Shin 2000). Most of the aftershocks occurred on the eastern side of the

Chelungpu fault because the fault is an east-dipping thrust fault. In Fig. 1, the epicenter distribution is shown for 20 months before and 9 months after the Chi-Chi earthquake. These earthquakes were recorded by the seismological network of the CWB (Central Weather Bureau). The daily number decreased monotonically in the aftershock stage and by 1 or 2 months after the main shock was at nearly the same level as before the main shock. We restricted the spatial distribution of hypocenters studied to get accurate measurements; however, the region should cover a sufficient number of aftershock hypocenters.

The earthquake data 20 months prior to the main shock and 9 months after the main shock in a volume (23º18'~24º24'N, 120º24'~121º30'E, depth shallower than 30 km) were selected as the database for this study. We divide the earthquake sequence into three periods: **I**, 1 January 1998 to 20 September 1999; **II**, 21 September 1999 to 31 December 1999; **III**, 1 January 2000 to 31 May 2000. We restricted the spatial distribution of focal depths in the region of 1 to 30 km. Totally, 388 events were selected, and its epicenter distribution is shown in Fig. 1.

In the routine data processing, the *i*th station magnitude $M_L(i)$, is calculated from the maximum amplitude (A), from the simulated Wood-Anderson seismograms, and the distance correction term, $logA_o$(D):

$$M_L(i) = logA - logA_o(\mathrm{D}),$$

where D is the epicentral distance in kilometers. Considering the focal depth of earthquakes in the Taiwan area, the attenuation function $logA_o$(D) are (Shin 1993):

$$\log A_o(\mathrm{D}) = \begin{cases} \textit{for focal depth (h)} \le 35\ km: \\ \quad 0.00716\, R + \log R + 0.39\ (0\ km \le \mathrm{D} \le 80\ km) \\ \quad 0.00261\, R + 0.83 \log R + 1.07\ (80\ km < \mathrm{D}) \\ \textit{for focal depth (h)} > 35\ km: \\ \quad 0.00326\, R + 0.83 \log R + 1.01 \end{cases}, \tag{1}$$

where $R = \sqrt{\mathbf{D}^2 + h^2}$ is the hypocentral distance and h is focal depth.

The average magnitude $\overline{M_L}$, usually called simply "magnitude", is the average of station magnitude over N stations:

$$\overline{M_L} \equiv \frac{1}{N}\sum_{i=1}^{N} M_L(i), \tag{2}$$

where an overbar denotes the station average.

3. DEVIATION OF STATION MAGNITUDE FROM THE AVERAGE MAGNITUDE

The station magnitude calculated for each station usually differs from the average magnitude. As described in the paper written by Sato (1986), station magnitude at the *i*th

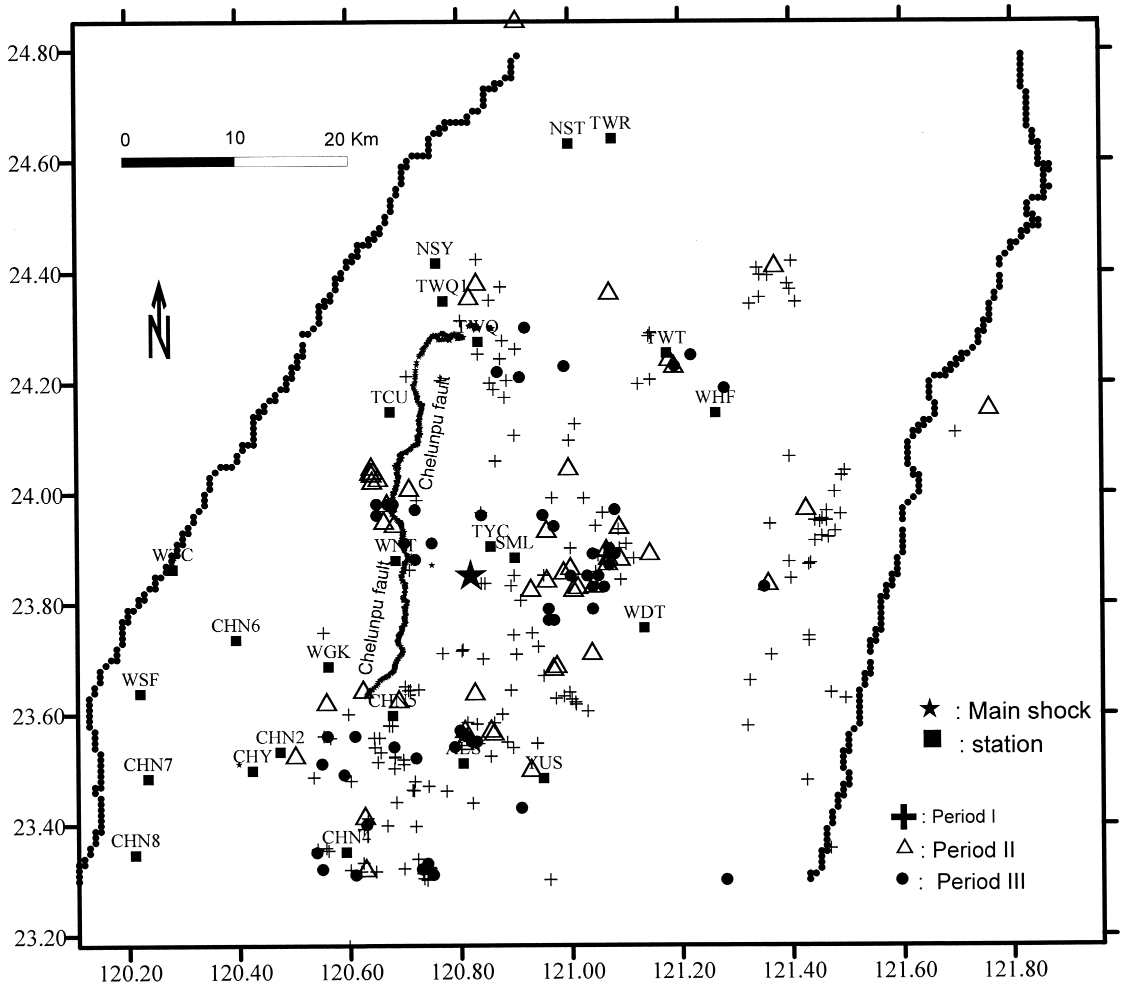


Fig. 1. Main shock of the Chi-Chi earthquake (purple *) and epicenter distribution of selected earthquakes 20 months before and 9 months after the Chi-Chi earthquake, that were analyzed in this study. These earthquakes are recorded and located by CWB (Central Weather Bureau) seismological stations (denoted by triangles) in Taiwan area.

station, *M(i)*, may be written as a sum of "true magnitude" corresponding to radiated seismic energy, M^o, radiation pattern effect $\delta M^R(i)$, "static" local site effect $\delta M^S(i)$ mostly caused by differential amplification, and "time dependent" correction term on the ray path $\delta M^T(i)$:

$$M(i) = M^o + \delta M^R(i) + \delta M^S(i) + \delta M^T(i). \tag{3}$$

Substituting Equation (3) into (2), we obtain

$$\overline{M} = M^o + \overline{\delta M^S} + \overline{\delta M^T}, \quad (4)$$

where we assume that $\overline{\delta M^R} = 0$. Only the difference $\delta M(i)$ between $M(i)$ and $\overline{M}$ is measurable for each station

$$\begin{aligned}\delta M(i) &\equiv M(i) - \overline{M} \\ &= \delta M^R(i) + \delta M^S(i) + \delta M^T(i) - \overline{\delta M^S} - \overline{\delta M^T},\end{aligned} \quad (5)$$

and $\overline{\delta M} = 0$. For one earthquake, in each station (i) which δM can be calculated by subtracting the station magnitude [$M(i)$], measured by Equation (1), with the average magnitude ($\overline{M_L}$), estimated by Equation (2). The 388 δM calculated from the events occurred within the study period (from 1 January 1998 to 31 May 2000) at 6 stations (WNT, SML, WGK, ALS, YUS, and WDT) in the vicinity of the focal area and 4 farther stations (CHN8, CHN4, NST, and WSF) are plotted in Fig. 2. We let angle brackets $\langle \delta M(i) \rangle$ mean a time average in each period. Assuming that $\Delta\langle \delta M^S(i) \rangle = \Delta\langle \overline{\delta M^S} \rangle = 0$, we can calculate the temporal change in $\langle \delta M(i) \rangle$ from period **I** to Period **II**:

$$\begin{aligned}\Delta\langle \delta M(i) \rangle_{II-I} &\equiv \langle \delta M(i) \rangle_{II} - \langle \delta M(i) \rangle_{I} \\ &= \Delta\langle \delta M^R(i) \rangle + \Delta\langle \delta M^T(i) \rangle - \overline{\Delta\langle \delta M^T \rangle},\end{aligned} \quad (6)$$

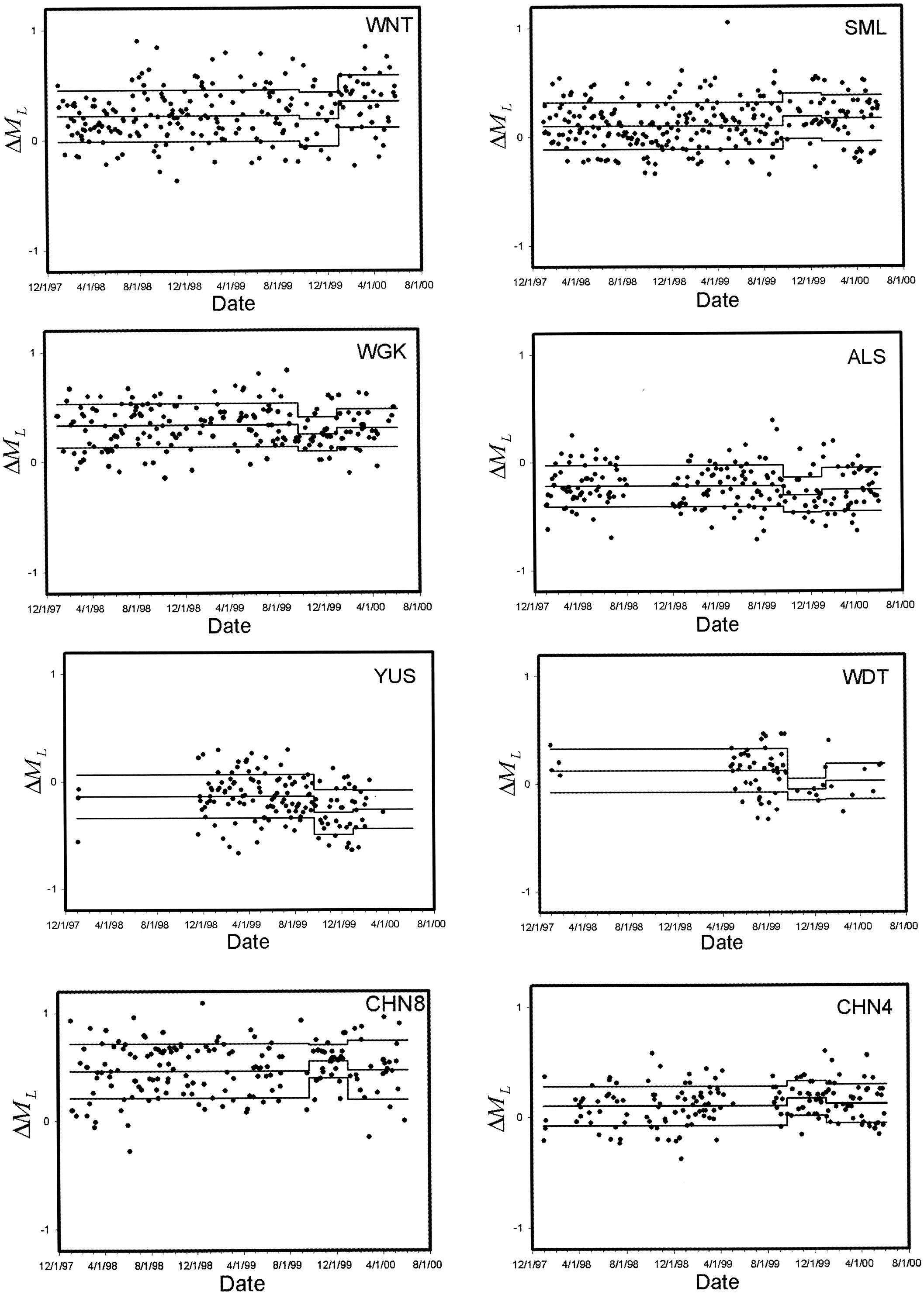
where $\langle \delta M(i) \rangle_I$ and $\langle \delta M(i) \rangle_{II}$ is the mean of $\langle \delta M(i) \rangle_{II}$ over period **I** and **II**, respectively. The steplike changes are apparently negative at five stations and positive at one station (SML) in the vicinity of the focal area. According to the study of Ukawa et al. (1984), at the station with an epicenter distance greater than 30 km will not be affected by the main shock. The changes $\Delta\langle \delta M(CHN4) \rangle$ = 0.071, $\Delta\langle \delta M(CHN8) \rangle$ = 0.0892, $\Delta\langle \delta M(NST) \rangle$ = 0.06425, and $\Delta\langle \delta M(WSF) \rangle$ = 0.067 are positive, since they are not deviations from the true magnitude but from the average magnitude. Substituting $\Delta\langle \delta M^R(i) \rangle$ = $\Delta\langle \delta M^T(i) \rangle$ = 0 at these farther stations into Equation (6), we obtain

$$\Delta\overline{\langle \delta M^T \rangle} = -0.09 \sim -0.06. \quad (7)$$

Substituting $\Delta\overline{\langle \delta M^T \rangle}$ = -0.075 and $\Delta\langle \delta M^R(i) \rangle$ = 0 into Equation (6), we get $\Delta\langle \delta M^T \rangle$ = -0.16 at ALS, -0.226 at YUS, -0.203 at WDT, -0.16 at WGK, -0.108 at WNT, and 0.013 at SML. We get the change on the average over six stations in the vicinity of the focal regions,

$$\Delta\overline{\langle \delta M^T \rangle}_{6\ stations} = -0.141. \quad (8)$$

We now try to estimate the change in Q_S^{-1} between periods **I** and **II** from the station magnitude fluctuation (8). Assuming that the maximum amplitude phase is **S** wave, we may write the maximum vertical component amplitude as


WNT
SML
WGK
ALS
YUS
WDT
CHN8
CHN4
ΔM_L
1
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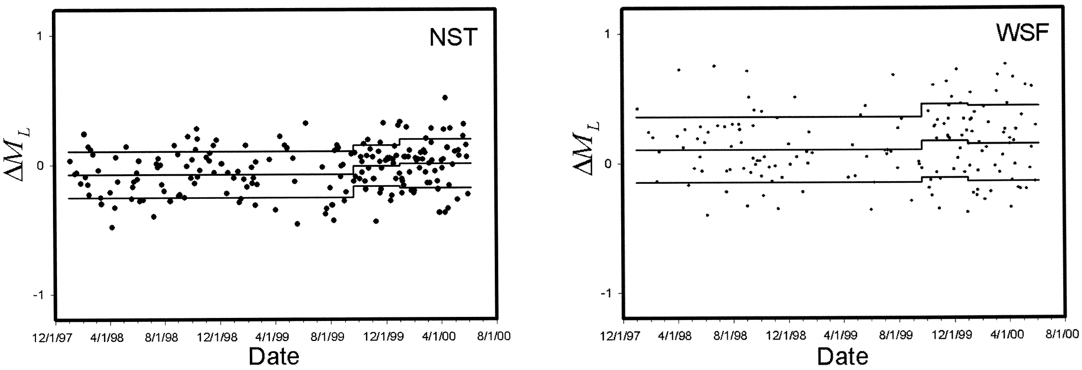

Fig. 2. Temporal changes in the deviation of station magnitude from average magnitude, δM, at 6 stations (WNT, SML, WGK, ALS, YUS, and WDT) are in the vicinity of the focal area and 4 farther stations. The mean (and standard deviation), corresponding to each period (**I**, **II**, and **III**), was a time average over each whole period. Bold lines represent mean and standard deviation.

$$A_v=(CR_oS_o/r)\exp(-Q_S^{-1}\pi f t_s), \tag{9}$$

where Q_s is the quality factor for S wave, R_o is the radiation pattern, S_o is the source factor as a function of the true magnitude M^o, r[km] is the hypocentral distance, t_s[s] is the S wave travel time, f[Hz] is frequency, and C is the local amplification factor. Substituting Equation (9) into (1), we get

$$M(i) = -Q_S^{-1}\pi f t_s \log e + \log(CR_o) + \log S_o(M^o) + 0.00716R + 0.39. \tag{10}$$

The temporal change in the deviation of station magnitude from the true magnitude in the vicinity of the focal area is produced by the change ΔQ_S^{-1} in the hemisphere:

$$\Delta\overline{\langle\delta M^T\rangle}_{6\ stations} = -\Delta Q_S^{-1}\pi\ f\ \overline{t_S}\log e. \tag{11}$$

The predominant frequency of the maximum amplitude phases is from 4 to 25 HZ (Shin 1993) on seismograms at the fifteen stations investigated. Substituting Equation (8) and average travel time of the six stations near the focal area, $\overline{t_S}$ =7.5 s, log e =0.4342942, into Equation (11), we have

$$\Delta Q_S^{-1}{}_{II-I} f = 1.378\times10^{-2}[s^{-1}] \qquad \text{for } 4 \sim 25 \text{ Hz}. \tag{12}$$

In order to avoid the effect by the major shock, we average the change in $\langle\delta M(i)\rangle$ at three stations (CHN4, CHN8, and WSF), that are far away from the focal area, to get the mean deviation in this area $\Delta\langle\delta M^T\rangle$ = 0.05. Substituting it and $\Delta\langle\delta M^R(i)\rangle$ = 0 into Equation (6), we obtain that $\Delta\langle\delta M_L^T(ALS)\rangle = 0.1$, $\Delta\langle\delta M_L^T(WDT)\rangle = 0.122$, $\Delta\langle\delta M_L^T(WGK)\rangle = 0.106$, $\Delta\langle\delta M_L^T(WNT)\rangle = 0.212$, $\Delta\langle\delta M_L^T(YUS)\rangle = 0.078$, and $\Delta\langle\delta M_L^T(SML)\rangle = 0.03$. Also, we can make a similar estimation of the temporal change in Q_S^{-1} in the hemisphere:

$$\Delta Q_S^{-1}{}_{II-I}\, f = 1.056 \times 10^{-2}\ [\mathrm{s}^{-1}] \qquad \text{for } 4 \sim 25 \text{ Hz.} \tag{13}$$

Calculated the difference of magnitude deviation in the study area between period **I** and period **II** ($\Delta\langle\delta M\rangle_{II-I}$) and between period **III** and period **II** ($\Delta\langle\delta M\rangle_{III-II}$) at fifteen stations near the main shock and aftershock area, we can make a contour map of the spatial distribution of the $\Delta\langle\delta M\rangle$, shown in Figs. 3 and 4. Figure 3 shows that the change of $\Delta\langle\delta M\rangle$ after the main shock is negative in most of the study area, with an small area (about 10×10 km) in the east of the main shock (near the stations SML and TYC) having positive change of $\Delta\langle\delta M\rangle$. The negative change area turns to positive three months after the main shock, while the small area with positive change near the focal area still stays positive (Fig. 4).

4. DISCUSSIONS

It is hard to ascertain whether a change in focal mechanisms will affect the station magnitude or not. We calculated the changes $\Delta\langle\delta M\rangle$ for the whole Taiwan area with estimation similar to that mentioned above. The results show that the changes $\Delta\langle\delta M\rangle$ between period **I** and **II** ($\Delta\langle\delta M\rangle_{II-I}$), and between **II** and **III** ($\Delta\langle\delta M\rangle_{III-II}$), seem to be very small, especially near the Chelungpu fault area and the focal area of the Chi-Chi earthquake (Fig. 5 and Fig. 6). Although two remarkable changes in southern and northeastern Taiwan, respectively, can be found are far away from the study area. They may be affected by the shallow local anomaly of attenuation intensity under those areas more than in other areas. The temporal unchanged pattern (comparing Fig. 5 and Fig. 6) implies that the changes $\Delta\langle\delta M\rangle$ cannot be explained by a systematic change in focal mechanisms only.

Changes in the density, distribution, and saturation of cracks are the probable driving mechanisms for most earthquake precursors. The dilatancy diffusion hypothesis (Scholz et al. 1973; Aggarwal et al. 1973) may explain the behavior in the immediate vicinity of an impending fracture. Crampin and Evans (1984) proposed that subcritical crack growth at low stress and low strain rates produces extensive dilatancy anisotropy through earthquake preparation zones; the cracks tend to grow parallel to the maximum compressive stress. In such a cracked medium, scattering by open cracks (Kikuchi 1981) and interaction with viscosity of water in thin cracks (Fehler 1977) are important mechanisms of temporal change in attenuation intensity. From the molecular level analysis, Tittmann et al. (1980) proposed a stress-induced diffusion model based on the interaction between adsorbed layers of volatiles, notably water, and the solid surface of rock minerals in terms of thermally activated motions in the adsorbed film. Suzuki (1971) and Chen (1998) found that the location of the high Q_S^{-1} region coincided with the seismically most active region. Chen et al. (1996) also found that high Q_p^{-1} regions have higher seismicity. These analyses show that high Q_p^{-1} or high Q_S^{-1} regions might reveal fractured media.

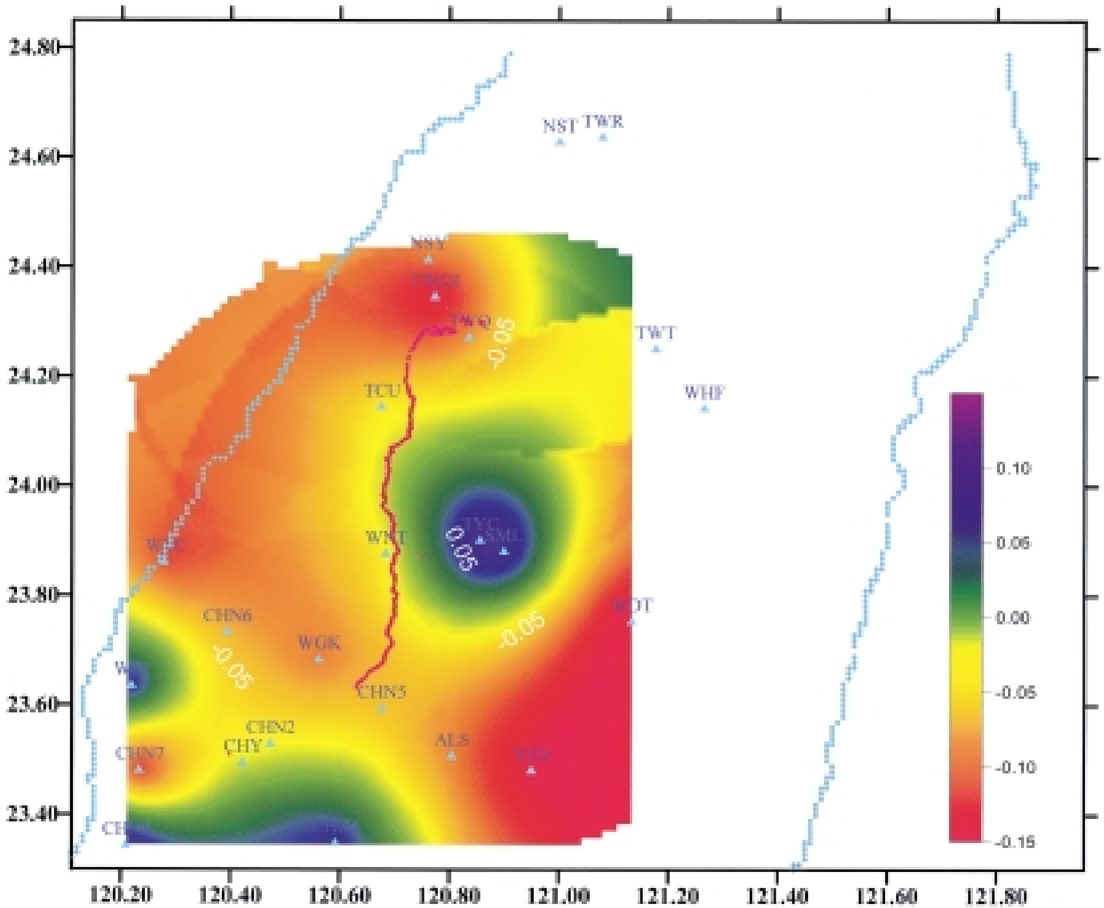

Fig. 3. Contour map of $\Delta\langle\delta M\rangle_{II-I}$ (difference of station magnitude deviations in the study area between period **II** and **I**), calculated by subtracting the distribution of station magnitude deviations in period **I** from that of period **II**. Values (are unitless) in contours and color chart denotes the variation of magnitude.

The change in attenuation $\Delta Q_S^{-1} = 1.378 \times 10^{-2} f^{-1}$ here estimated amounts to 60% of $Q_S^{-1} = 2.3 \times 10^{-2} f^{-1}$ in the lithosphere under the Kanto district (Sato and Matsumura 1980). Sato (1984) previously proposed a model for explaining frequency dependent Q_S^{-1} in the lithosphere by scattering loss due to randomly inhomogeneous elastic structure. According to this theory, the maximum value of Q_S^{-1} is roughly equal to the mean square of the velocity fractional fluctuation. Formation of new cracks, reopening of existing closed cracks, and water movement through cracked media are the most viable mechanisms for increasing Q_S^{-1} prior to the main shock in a shallow portion of crust; they will directly and indirectly affect attenuation intensity through changes in velocity in dilation region.

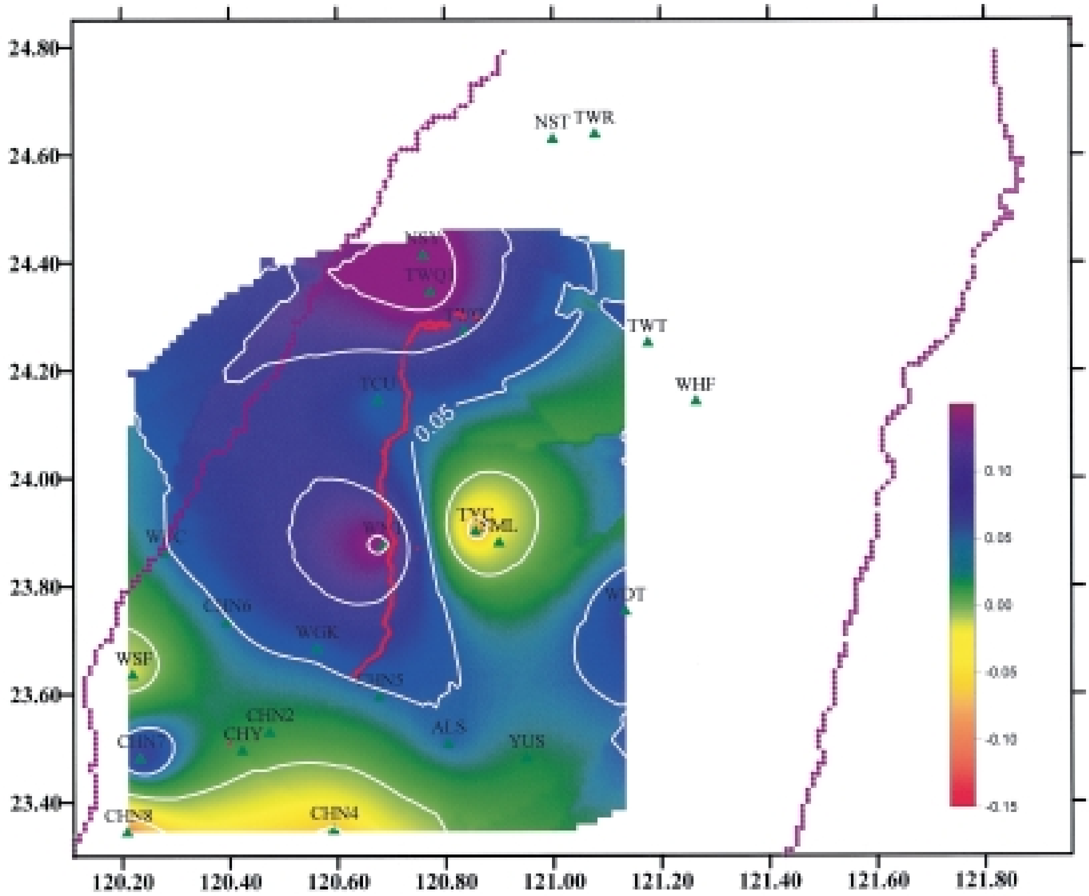

Fig. 4. Contour map of $\Delta\langle\delta M\rangle_{III-II}$ in the study area.

5.CONCLUSIONS

An earthquake of magnitude 7.3 (M_L=7.3, M_W=7.7) occurred on 21 September 1999, in central Taiwan. Earthquakes occurring in a volume (mentioned in Data section) in this earthquake fault region were studied. Seismograms recorded at the surrounding 15 stations were analyzed for a period from 20 months before to 9 months after the main shock. It was found that the deviation of station magnitude from average magnitude decreased after the main shock at five stations, but increased in one station in the vicinity of the focal area, after the main shock compared with that before. The simplest interpretation is that the attenuation factor increased as $\Delta Q_S^{-1}=1.378\times10^{-2}\times f^{-1}$ for frequencies within 4 ~ 25 Hz after the main shock in the vicinity of the focal region.

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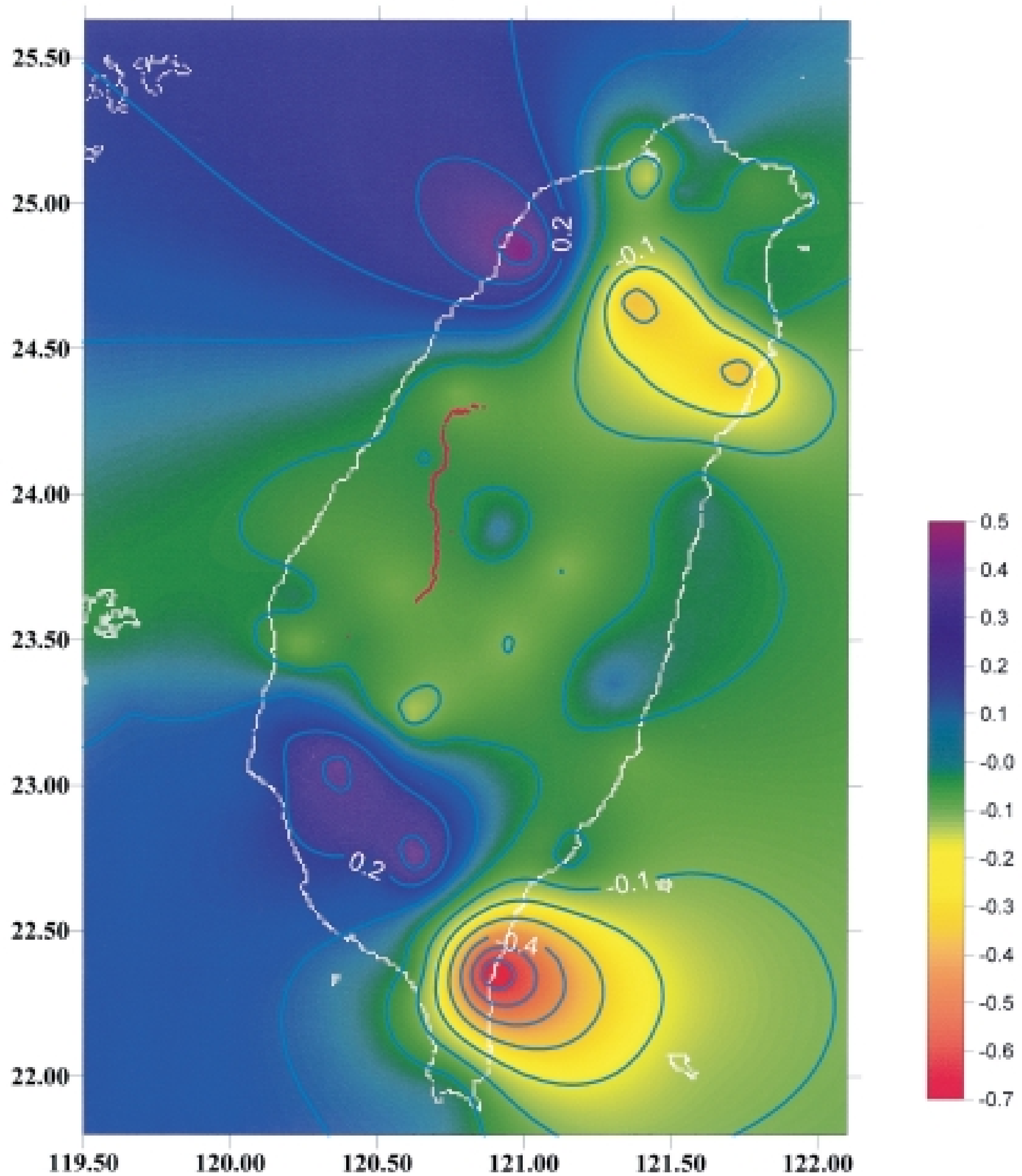

Fig. 5. Contour map of $\Delta\langle\delta M\rangle_{II-I}$ (difference of station magnitude deviations for the whole Taiwan area between period **II** and **I**), calculated by subtracting the distribution of station magnitude deviations in period **I** from that of period **II**. Values (are unitless) in contours and color chart denotes the variation of magnitude.

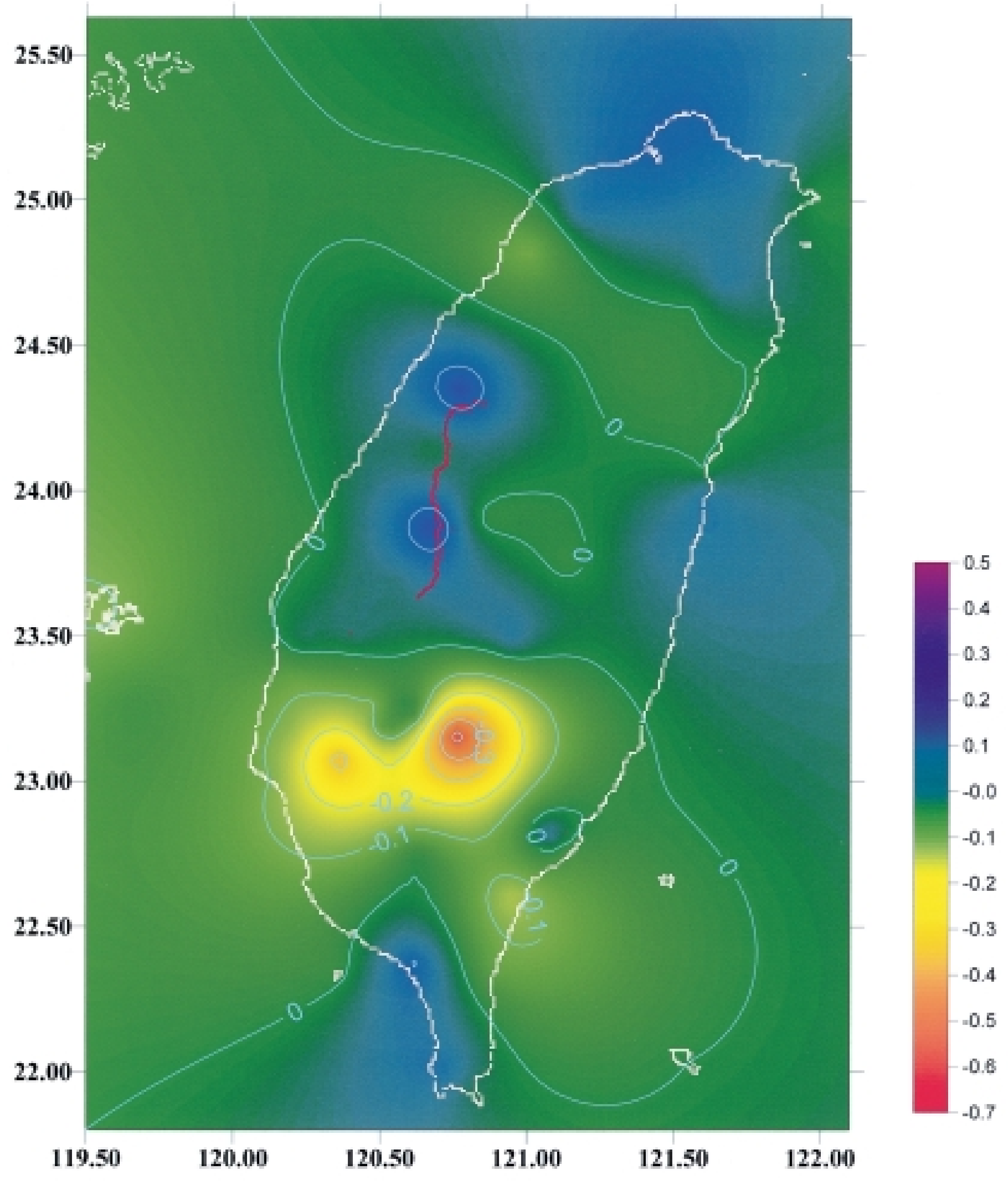

Fig. 6. Contour map of $\Delta\langle\delta M\rangle_{III-II}$ for the whole Taiwan area.

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