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Title: Mechanism of Different Coseismic Water-Level Changes in Wells with Similar Epicentral Distances of Intermediate Field

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Abstract: Water level changes at different monitoring stations are observed during the Wenchuan earthquake (Ms8.0) in the Chinese mainland. In the intermediate field, we observed co-seismic water level changes of different amplitude in wells with similar epicentral distances. In order to study about the mechanism of those co-seismic water level changes, we calculated the static strain change with the Okada's dislocation model. Compare the calculated co-seismic water level change based on the poro-elastic theory with the observed water level change, we can judge whether the poro-elastic theory can be applied to the aquifer of the well, from which we find that: When the water level change of those wells can be explained by the poro-elastic theory (from our research we find the poro-elastic theory can be applied to the area with epicentral distance < ~1.5 fault rupture length), the difference of the water level change in wells with similar epicentral distances is mostly related to the difference of the Skempton's coefficient B. Otherwise, the water level change may be induced by the transition of the seismic waves, since it is usually larger than the one induced by the un-drained dilatation and consolidation, and changes more gradual.

Suggested Reviewers: Chi-yuen Wang
chiyuen@berkeley.edu
he is an expert in the region we studied in this paper

Opposed Reviewers:

Response to Reviewers: I will include this information in files that will be uploaded

Reply:

We have changed part of the paper (“Abstract”, “Introduction”, “Mechanism analysis”, “discussion” , “part of Table 2” and “Figure 6”), and use the highlighted yellow color to show those changes in the “Response to reviews”.

Reviewers' comments:

Reviewer #3: Review of Zhang and Huang

This manuscript presents some interesting and unusual data of water level responses to a major earthquakes where the wells span a large range of epicentral distances. I did not review the earlier versions and did not spend a great deal of time analyzing the earlier referees comments, but it seems that the revised sections and interpretations have improved significantly. The separation of the wells into groups consistent and inconsistent with the poroelastic response is a good innovation. In fact, I would emphasize this aspect of the paper more strongly in the introduction and downplay the Skempton coefficient variation, which is not in my opinion the strongest or most interesting part of the paper. That said, there are still a few technical inaccuracies that need to be addressed before publication.

1. As a previous reviewer commented, there is a misrepresentation of the current models of hydrological changes from earthquakes in the introductory paragraph. The sentence that "At even greater distance ? only transient oscillations of the water level have been documented" is incorrect. Sil (2006) document sustained steps at teleseismic distances although the data is messy. The Brodsky et al. 2003 reference focuses on explaining sustained farfield water level changes by using information from the transient waves. Fig. 5.6 of the book by Wang and Manga, "Earthquakes and Water" (Springer, 2010 shows several other examples.

Reply: We have modified in the introductory paragraph, and changed the sentence that "At even greater distance only transient oscillations of the water level have been documented" into “In the far field, transient oscillations of the water level have been documented, and sustained water level changes also have been observed”, as shown in page 2.

2. The explanation of the physical interpretation of a low Skempton coefficient on p. 5 as "a low value of B indicates a poorly confined aquifer system" is inaccurate. (Similar comments apply on p. 11). A Skempton coefficient is defined based on the undrained

state (eq. 3). Therefore, the degree of draining (confinement) does not affect its value. The comment cited in Sil (2006) apparently refers to a low effective Skempton coefficient inferred from a tidal model, which really means that the undrained tidal model is insufficient. That same comment may be applicable here as well, but in this case, the comment should be clarified to indicate that the true physics is that the tidal model is inaccurately capturing the Skempton coefficient. An actual low Skempton coefficient is better interpreted as a consquence of a stiff rock matrix that supports the load without significant coupling to the fluid (Nur and Byerlee, 1971).

Reply: This is really a good suggestion, and I have learned the true physics of the Skempton's coefficient from the reviewer's comments. We have changed the description into"a low value of B indicates the stiff rock matrix that supports the load with low coupling to the fluid", "When B-values are high, the stiff rock matrix supports the load with high coupling to the fluid" as shown in page 5 and page 10.

3. I am unconvinced that the water level changes in wells 1-12 are more abrupt than the changes in wells 13-27. Wells 14, 15 and 25 certainly look abrupt to me. Therefore, the conclusion that the poroelastic change results in abrupt steps and the transient wave changes results in gradual steps does not seem well supported by the data.

If the authors disagree and still think that there is a difference in the abruptness, can they please quantify the speed of the water level change and produce a graph how well the poroelastic model works (q) as a function of abruptness. For instance, they can measure the time that the water level change takes to reach 80% of its maximum anomaly and plot this versus q. If their assertion of a correlation between abruptness and poroelasticity is correct, it should be clear in such a figure.

Reply: According to the reviewer's comments, we have modified the paper, and just use the reviewer's suggestion--------"in the relatively near field data (<~ 1.5 fault rupture lengths) may be consistent with a poroelastic model. Beyond this distance, the data is demonstrably inconsistent with a poroelastic model and the seismic waves likely play a role in determining the water level change." We do agree that is a good suggestion. The modification is shown in the high lightened part (in Abstract, Introduction and Conclusions).

4. The statement on p. 10 that the orientation of the receiver fault is important for calculating the poroelastic change is not consistent with the volumetric strain change that

is used for all interpretations in this paper (Table 2). Although eq. 1 certainly has orientation dependent terms, the homogeneous, isotropic poroelastic model produces a volumetric strain at a point that is invariant to any receiver fault plane orientation. The volumetric strain is used in the unnumbered equation for delta h on p. 8 and therefore the receiver fault orientation is irrelevant.

Reply: as indicated by the reviewer, we recognize this problem, and "the receiver fault theory" seems contradict with "the stress calculated in one point according to the Okada's dislocation model, which just suppose the whole land is homogeneous, isotropic", so the receiver fault orientation is irrelevant, we have deleted all those descriptions relating to the receiver fault.

All of these issues should be remediable with a revision, although #3 may require a major change in interpretation.

I should say that I would be tempted to write a somewhat different paper with this data. I would have concluded based on the data shown that for the most part the relatively nearfield data (<~ 2 fault rupture lengths) may be consistent with a poroelastic model. Beyond this distance, the data is demonstrably inconsistent with a poroelastic model and the seismic waves likely play a role in determining the water level change. This relatively simple division of the data seems to be well-supported by the observations and analysis and emphasizing the separation by distance would be a useful contribution to BSSA. Of course, the authors may view things differently and as long as the technical points above are fully addressed, their interpretation can be published. I simply include my own interpretation in the hopes that it might be a helpful suggestion.

Reply: Thanks very much for the reviewer's valuable suggestions. We have taken the good suggestion, and modified the paper accordingly (see the yellow highlighted part).

Mechanism of Different Co-Seismic Water Level Changes in Wells with Similar Epicentral Distances of Intermediate Field

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Abstract

Water level changes at different monitoring stations are observed during the Wenchuan earthquake (Ms8.0) in the Chinese mainland. In the intermediate field, we observed co-seismic water level changes of different amplitude in wells with similar epicentral distances. In order to study about the mechanism of those co-seismic water level changes, we calculated the static strain change with the Okada's dislocation model. Compare the calculated co-seismic water level change based on the poro-elastic theory with the observed water level change, we can judge whether the poro-elastic theory can be applied to the aquifer of the well, from which we find that: When the water level change of those wells can be explained by the poro-elastic theory (from our research we find the poro-elastic theory can be applied to the area with epicentral distance < ~1.5 fault rupture length), the difference of the water level change in wells with similar epicentral distances is mostly related to the difference of the Skempton's coefficient *B*. Otherwise, the water level change

may be induced by the transition of the seismic waves, since it is usually larger than the one induced by the un-drained dilatation and consolidation, and changes more gradual.

Introduction

Several types of earthquake induced groundwater level changes and corresponding mechanisms have been recognized for decades. In the near field (generally, epicentral distance D between 0-100 km), most documented water level shows abrupt (step-like) coseismic changes (Wakita 1975; Quilty and Roeloffs, 1997; Wang et al., 2001, 2004; Chia et al., 2001; Wang and Chia, 2008). Undrained dilatation and consolidation of the sediments may be responsible for the step-like water level changes in the near field, and can often be quantitatively related to the poroelastic response to the earthquake's static strain. In the intermediate field (epicentral distance D between 100-1000 km), most documented changes are gradual and can persist for days or weeks. These are coined by Roeloffs (1998) as the 'sustained' water level changes, and an earthquake-enhanced permeability may be responsible for this intermediate field phenomenon (Wang and Chia, 2008). In the far field (epicentral distance D larger than 1000 km), transient oscillations of the water level have been documented, and sustained water level changes also have been observed. There are several existing models for far-field coseismic pore pressure changes: mobilization of gas bubbles, (Roeloffs, 1998), shaking induced dilatancy (Bower and Heaton, 1978), fracture of an impermeable fault (King et al., 1999), fracture clearing (Brodsky et al., 2003), and shaking induced by surface waves (West et al., 2005; Sil and Freymueller, 2006).

Investigation of coseismic water level changes has been of scientific interest for decades (Wang and Manga, 2010). Groundwater level changes following earthquakes can affect water supply; seismic waves can affect oil well production, and it has been suggested that in some cases the induced seismicity can stimulate oil production (Beresnev and Johnson, 1994). Earthquake-induced fluid pressure changes are hypothesized to control the timing and/or location of the aftershocks and trigger seismicity (Hill et al., 1995; Gomberg, 1996). Finally, these groundwater level changes could also be related to the hydrologic earthquake precursors (Roeloffs, 1998).

In this paper we calculate Skempton's coefficient *B* from the poroelastic relationship between water level changes and tidal strain using data prior to the earthquake. Further analysis of the water level data from the Groundwater Monitoring Network (GMN) (see Data and Resources Section) is done during the Wenchuan earthquake for intermediate field. A relation between the amplitude of the water level , the earthquake magnitude and distance is developed by Roeloffs (1998) for the "sustained" water level changes. To develop this relationship, different intermediate field earthquakes are used. Several authors have obtained similar empirical relations between water level change, epicentral distance, and the earthquake magnitude (Matsumoto et al., 2003; Yang et al., 2005; Sil and Freymueller, 2006).

In addition to the above observation, we find that the size of the water level change at GMN stations in the intermediate field is not only related to the earthquake magnitude and the epicentral distance. Several wells with similar epicentral distances have different amplitude of co-seismic water level changes, and some of those wells even stay close to each other in one fault. We calculated the static strain change with the Okada's dislocation model. Supposing the poro-elastic theory can be applied to all of those wells, based on that we calculated the Skempton's coefficient *B* of all those wells. With the calculated static strain change and the Skrmpton's coefficient *B*, we derived the co-seismic water level changes. Compare the calculated co-seismic water level change based on the poro-elastic theory with the observed water level change, we can judge whether the poro-elastic theory can be applied to the aquifer of the well, so as to study about the mechanism of those co-seismic water level changes.

In this paper, we find that: the poro-elastic theory can mainly be applied to the area with epicentral distance < ~1.5 fault rupture length, during this area the coseismic water level changes are mostly induced by the undrained dilatation and consolidation of the sediments, and the difference of the water level change in wells with similar epicentral distances is mostly related to the difference of the Skempton's coefficient *B* of those wells. Large *B*-values come with large changes in water level. This phenomenon is in accordance with the poro-elastic theory. When the poro-elastic theory can only be applied to one of the wells with similar epicentral distances, usually the water level change of the other well is more gradual and with much larger amplitude, we may infer it is induced by the earthquake shaking, which is caused by the transition of the seismic waves.

Theory

Skempton's coefficient *B* is a significant pore-fluid parameter in poroelastic theory. A poroelastic material consists of an elastic matrix containing interconnected fluid saturated pores. Fluid saturated crust behaves as a poroelastic material to a good degree of approximation.

Rice and Cleary (1976) summarized the following equations for a linearly elastic isotropic porous medium, which are the building blocks of the poroelastic theory:

$$2G\varepsilon_{ij} = \sigma_{ij} - \frac{\nu}{1+\nu}\sigma_{kk}\delta_{ij} + \frac{3(\nu_u - \nu)}{B(1+\nu)(1+\nu_u)} p\delta_{ij}, \qquad (1)$$

$$m - m_0 = \frac{3\rho(\nu_u - \nu)(\sigma_{kk} + 3p/B)}{2GB(1+\nu)(1+\nu_u)}. \qquad (2)$$

Here m-m_o is the change of the fluid mass, ε_{ij} is the strain tensor, σ_{ij} is the stress tensor, δ_{ij} is the Kronecker delta function, G is the shear modulus, ρ is the density of the fluid, *B* is the Skempton's coefficient, p is the pore pressure, ν is the Poisson's ratio, and ν_u is the "undrained" Poisson's ratio. Rice and Cleary (1976) describe equation 1 as a stress balance equation and equation 2 as a mass balance equation.

For the undrained condition, the poroelastic effect on the crust can be obtained by putting m-m_o =0 in equation 2, and therefore we obtain:

$P = -B\sigma_{kk} / 3$ or $\Delta p = -B\Delta\sigma_{kk} / 3$. (3)

Equation 3 says under "undrained" condition, the change in fluid pressure (Δp) is proportional to the change in mean stress ($\Delta\sigma_{kk}$/3). This is the mechanism of water level changes for poroelastic material. (p=ρg h, where h is the water column height, g is the acceleration due to gravity and ρ is the density of water).

According to equation 3, Skempton's coefficient *B* can be qualitatively defined: In the "undrained" condition, *B* is the ratio of the induced pore pressure divided by the change in mean stress (Wang, 2000). *B* governs the magnitude of water level changes due to an applied stress since pore pressure is directly proportional to water level. The value of *B* is always between 0 and 1. When *B* is 1, the applied stress is completely transferred into changing pore pressure. *B* equals 0 indicates no change in pore pressure after applying the stress. Thus a low value of *B* indicates the stiff rock matrix that supports the load with low coupling to the fluid (Nur and Byerlee, 1971). Laboratory studies indicate the value of *B* depends upon the fluid saturated pore volume of the sample (Wang, 2000).

Equation 3 can be expressed in terms of tidal strain as well (Roeloffs, 1996):

$$\Delta h = -\frac{2GB(1+\nu_u)}{3\rho g(1-2\nu_u)}\Delta\varepsilon_t \quad (4)$$

Equation 4 shows that water level changes proportionally in a poroelastic material under the influence of tidal strain (ε_t). Here Δh is the change in height of water level, and $\Delta\varepsilon_t$ is the corresponding tidal strain change (Sil, 2006).

From equation 4 we obtain:

$$B = -\frac{3\rho g(1-2\nu_u)}{2G(1+\nu_u)}\frac{\Delta h}{\Delta\varepsilon_t} \quad (5)$$

With equation (5) we can get the value of *B* with water level and tidal strain. However, the calculation must be on the strict premise of the undrained condition, the good correlation ship between the water level and the tidal strain and should not be influenced by the other factors.

For the effect of the solid tide on the crust, when the wavelength of the tidal strain is much larger than the size of the aquifer, we can suppose the aquifer system is undrained (Huang, 2008). The wavelength of the M2 wave is about 2 406 329 km ($\lambda=\omega\times r\times T$, $\omega=1.4\times10-4$/s is the angular frequency of M2 wave, r=384 400 km is the distance from the earth to the moon, T=745.236 min is the period of the M2 wave), which is much larger than the size of the radius of the Earth, and is definitely much larger than the thickness of the aquifer systems of those wells. Thus, the effect of the M2 wave in the crust can meet with the undrained condition (Zhang et. al, 2009). Besides, those wells can record clear tidal strains and as we calculate the phase lags between the water levels and the tidal strains are small, thus the wells can meet with the undrained condition well. In the M2 wave frequency domain the water level and the tidal strain have a good correlationship, we just set the Changping station as an example to see the correlationship clearly (Figure 1). We can see in the M2 wave frequency domain the correlationship between the tidal strain and the water level approaches 1, which means a good correlationship between them. Besides, the M2 wave is hardly influenced by atmospheric pressure. Since that, we distill the frequency domain of the M2 wave from the water level and the tidal strain by using band-pass filter (the frequency of the M2 wave is 0.0805114 h^{-1}) to calculate the Skempton's coefficient *B* (Figure 2). Disposing the obtained frequency domain of the M2 wave by IFFT (inverse fast Fourier transform) and adjusting their phase, through the least square fit and putting the results into equation (5), we can finally derive *B*. More details of the method are explained by the paper "Research on Skempton's coefficient *B* based on the observation of groundwater of Changping station" (Zhang et. al, 2009). All the Water level observations come from the sensor of water level, while tidal strain data are calculated via Mapsis software, which is programmed by Shengle Li.

Methods

Water level changes at different monitoring stations are observed during the Wenchuan earthquake (Ms8.0) in the Chinese mainland. We aim at exploring the mechanism of those co-seismic water level changes of different amplitude in wells with similar epicentral distances.

We only find 27 wells which can form groups that have similar epicentral distance (within a range of less than 0.15 degrees or 16.68 km) in the intermediate field of mainland China (Figure 3). One well (Weinanshuangwang) has been deleted since we can not confirm the range of the shear modulus of its lithology (Sand clay). We divided those 27 wells into twelve groups (group a to group l), each group has a specific range of epicentral distance (Table 2). As show in Figure 3, wells in group a (well 1, 2), b (well 3, 4), c (well 5, 6) , h (well 15, 17) and k (well 24, 25) stay close with each other.

First of all, we suppose the poro-elastic theory can be applied to all of those 27 wells. We apply the method of *B*-value calculation to those 27 wells. Pre-earthquake analysis is carried out using data from May 2, 2008 to May 10, 2008 to obtain the $B*$ values (Table 2). Calculation is performed using $\rho = 1000 kg/m^3$, $g = 9.8 m/s^2$, and $\nu_u = 0.29$. Since the shear modulus will change with the change of the stress, we can hardly get the in suit value of the shear modulus of those wells by experiment, which is as hard as getting the in suit Skempton's coefficient B. We have investigated the geology of each well and referred to the Rock Mass Mechanism (Liu and Tang, 1998), using the dynamic elastic modulus and dynamic Poisson's ratio to estimate the range of the shear modulus of those rocks, and approximately choose the mean value (Table 1).

Then, we must check if the prediction of the poro-elastic theory is consistent with the observed water level changes, so as to check whether the poro-elastic theory can be applied to the aquifer of the 27 wells. Since that, we show the co-seismic volume strain changes in Table 2, which is already calculated by Fuqiong Huang in her PhD Dissertation with Okada's dislocation model (Huang, 2008). We have plotted those wells with the spatial distribution of the static volume strain change of Wenchuan earthquake (Figure 4), and also plotted the original water level change of those 27 wells in Figure 5. From equation (3) we can obtain $\Delta h^* = -\frac{B^* \cdot \Delta\sigma_{kk}}{\rho g} = -\frac{B^* \cdot (E^* \cdot \Delta\varepsilon_{kk})}{\rho g}$, we calculated the water level change from B^* and the static stress change $E^* \cdot \Delta\varepsilon_{kk}$. We can judge whether the well aquifer can fit for the poro-elastic theory just by comparing the observed water level change Δh and the water level change Δh^* calculated from the poro-elastic theory. Define $q = \frac{\Delta h}{\Delta h^*}$, we calculated q values of those 27 wells (Table 2). As show in Table 2, when the value of q is too large (it means there are huge differences between the theory value Δh^* and the real value Δh) or $q < 0$ (it implies the sign of the water level change is not consistent with the direction of the volume strain change, and is not caused by the un-drained consolidation or dilatation) , the well aquifer may not fit for the poro-elastic theory, and we should not use the poro-elastic theory to explain the mechanism of water level change. Since that, the B^* value which is calculated based on the poro-elastic theory will be invalid.

Generally, according to the q and static strain change values in Table 2 and also take the patterns of those co-seismic water level changes in Figure 5 into consideration, we take q=30 as the threshold value, when $q<30$ we suppose the poro-elastic theory can be applied to the well aquifer, otherwise if $q>30$ the mechanism of the water level change may not be the static strain change, thus the poro-elastic theory may not be applied to the well aquifer.

Firstly, as show in Table 2, except for well 7, q values of well 1 to well 9 are all smaller than 30, they are much smaller than those q values of well 10 to well 27 (except well 14). The mean q value of those wells (well 1 to 9, discarding well 7) is $\bar{q} = 7.90$, which is relatively acceptable (The Okada's dislocation model is based on the assumption that the whole land is isotropic and homogeneous, and does not consider about the geology conditions. However, there are several faults between those regions and the epicenter, so the medium is not uniform, and the volume strain change $\Delta\varepsilon_{kk}$ got from this model will definitely have some differences from the real condition (Figure 3). Besides, when we calculate the B^* value, we use the mean value of the shear modulus G, it may be different from the real G value (Table 1). Inevitably, there must be some differences between the water level change calculated from the poro-elastic theory $\Delta h^* = -\frac{B^* \cdot \Delta\sigma_{kk}}{\rho g} = -\frac{B^* \cdot (E^* \cdot \Delta\varepsilon_{kk})}{\rho g}$ and the observed water level change Δh, thus the mean value $\bar{q} = 7.90$ is relatively acceptable).

Secondly, the static strain values of well 10-27 are obviously smaller than that of well 1-9 (except well 7) (Table 2), the seismic energy density in the relatively far field (D> 430 km) may be too small to initiate un-drained consolidation and dilatation, a distinct mechanism is required to explain the water level changes at such distances.

From the analysis above, we may just get 9 wells which can fit for the poro-elastic theory in the intermediate field (well: 1, 2, 3, 4, 5, 6, 8, 9, 14) (Table 2). Since that the B values of those 9 wells are valid, meanwhile we use "/" to indicate the invalid B values (Table 2).

Mechanism analysis

Among those 9 wells to which the poro-elastic theory can be applied, only 6 can form groups with similar epicentral distances (well: 1, 2, 3, 4, 5, 6). We find that large pre-earthquake *B* values correspond to large magnitude of co-seismic water level changes, this phenomenon exists in those 6 wells (Figure 6). We use poro-elastic theory to analyze the mechanism of this phenomenon.

From equation (3) we can see the water level change $\Delta h = \Delta p / \rho g$ is related to the static stress change $\Delta \sigma_{kk} / 3$ and the Skempton's coefficient *B*. From Table 2 we can see, the amplitude of the co-seismic water level change in each group is not always in accordance with the amount of the static strain change. Set group a as an example, the volume strain change of well 2 is larger than that of well 1, but the amplitude of the co-seismic water level change of well1 is larger than that of well 2. This phenomenon widely exists in group a, b, and c. This obviously shows that, *B* governs the magnitude of water level change induced by the applied stress.

Large *B*-values come with large changes in water level. This phenomenon is in accordance with the poro-elastic theory. When *B*-values are high, the stiff rock matrix supports the load with high coupling to the fluid (Nur and Byerlee, 1971) , the applied stress is mostly transferred into changing pore pressure, which leads to relatively large changes in water level. When *B*-values are low, the stiff rock matrix supports the load with low coupling to the fluid (Nur and Byerlee, 1971) , the applied stress can be easily transferred outside the aquifer system without increasing the pore pressure resulting in small water level changes (Sil, 2006). This can be used to explain: why two wells stay close with each other, but the amplitudes of their co-seismic water level changes are different.

In the other 3 groups (group: d, e, g), the water level changes in 3 wells (well 8, 9, 14) can be explained by the poro-elastic theory, while the other 3 can not (well 7, 10, 13) (Table 2). As show in Table 2, in group e and g, the water level change of the well to which the poro-elastic theory can be applied is smaller than the other one. Therefore, we can imply that the water level change in the 2 wells (well 10, 13) may be induced by the transfer of the seismic waves. As has been reported, earthquakes can produce sustained water level changes in certain distant wells that are often orders of magnitude larger than can be explained by static stress changes (Bower and Heaton, 1978). The shape of water level change in well 7 is sustained, although the amplitude is not large, we may assume that it is also caused by the transition of the seismic waves.

Discussion

Water level changes in regions to which poro-elastic theory can be applied are consistent with the volume strain changes. That means, when the volume strain change is positive (dilatational) the water level decrease, and when the volume strain change is negative (compressional) the water level increase (Table 2). Among those 27 wells the water level change of 8 wells are not consistent with the volume strain change (well: 13, 15,16,17,19, 21, 22, 25), and those wells are distributed in different areas in the Chinese Mainland (Figure3). As we calculated the q values of those wells, the result is in accordance with the above result. In those wells (well: 13,15,16,17, 19, 21, 22, 25) the q values are obviously much larger than the others, it means that the poro-elastic theory can not be applied to those wells, and the water level change in those wells are definitely not caused by the static volume strain change.

For intermediate distance earthquakes, several authors previously obtained similar empirical equations (shown below) relating water level change, epicentral distance, and magnitude of the earthquakes (Roeloffs, 1998; Matsumoto et al., 2003; Yang et al., 2005; Sil and Freymueller, 2006). And this empirical equation is based on the mechanism of shaking induced water level change. They attribute the magnitude of the water level change to two major impact factors: earthquake magnitude and epicentral distance. The empirical relation found by them can be written as:

$$\log_{10}\Delta h_i = w_1 M + w_2 \log_{10} D + w_3 . \qquad (6)$$

In this equation w_1, w_2 , and w_3 are constants, Δh_i is the size of the water level change in centimeters, M is the earthquake magnitude, and D is the well- hypocenter distance in kilometers (Roeloffs, 1998). The importance of equation 6 is that, for intermediate distances, it can explain earthquake induced water level changes, where poro-elastic theory generally is not applicable.

However, it is hard to explain water level changes in group f and group h to l with equation 6. According to equation 6, with the same magnitude and similar epicentral distances, water level changes in the same group (in group f and group h to l) are expected to be similar, but as shown in Table 2 the amplitude of those water level changes are different.

From our analysis, we think the unfitness of equation 6 lies in the shape conformity of those co-seismic water level changes and the geology condition of those wells. As studied by Roeloffs (1998), and Yang et al. (2005), they just use one specific well with different earthquakes to obtain equation 6, all those co-seismic water level changes observed by them are ascending. Matsumoto et al. (2003) use the co-seismic water level changes induced by different earthquakes in the Haibara well to obtain equation 6, and all those co-seismic water level changes are descending. Sil and Freymueller (2006), use the co-seismic water level changes induced by the Sumatra-Andaman earthquake in wells with the same geology condition in one specific area to obtain this empirical equation, those co-seismic water level changes are all rising. Thus, same geology condition and shape conformity of those co-seismic water level changes (all rising or all descending) may be the premise of empirical equation 6.

As discussed above, equation 6 can be used to explain water level changes in well 15 and 17 of group h. The amplitude of the water level changes in the two wells are similar (Table 2). Well

15 and 17 lie in the similar area (Figure 3) with the same geology condition (Table 2), and both of them are increasing. For well 24 and 25, though lie in the same area with similar geology, one of the water level decrease while the other increase，which may lead to the unfitness of equation 6, thus the amplitude of water level changes in well 24 and 25 are different. On the other hand, the estimated co-seismic water level changes calculated from equation 6 are not always in conformity with the observed water level changes (Roeloffs, 1998; Matsumoto et al., 2003; Yang et al., 2005; Sil and Freymueller, 2006), sometimes there will be a little difference, which may lead to the different amplitude of water level changes in the same group.

Besides, the obscurity may also be caused by the large distances between those wells and the epicenter, and there are lots of faults, so the medium is not uniform. The Okada's dislocation model is based on the assumption that the whole land is isotropic and homogeneous. Therefore, there may be huge differences between the calculated volume strain change and the real value in those wells (well 19~27, except for well 24, 25), thus it is possible that their q values are not accurate. Since that, it is hard for us to study the mechanism of the water level changes in those wells based on the q values, and we should research those water level changes in further studies.

As discussed earlier, the shear modulus G will change with the change of the stress, and it is found to be the function of the Skempton's coefficient B (Berryman, 2004). We can hardly get the in suit value of the shear modulus of those wells by experiment. Thus there may be ranges of uncertainty in B values getting from mean G values, and this needs to be further studied.

We couldn't find data from near field (0< D <100 kilometers) wells with the similar epicentral distance during the Wenchuan earthquake.

Magnitude of the Wenchuan earthquake is relatively large (Ms 8). Therefore, even without computing, we can expect that the static strain field from the earthquake will affect a relatively large area (The area is about 430 kilometers away from the epicenter according to our study in this paper). Thus we assume that our observation is not contradicting any existing theory of earthquake induced water level changes. For the relatively far field, shaking induced by the transition of the seismic waves may be the major mechanism of the co-seismic water level changes.

Conclusions

In this paper we discussed the mechanism of the co-seismic water level changes of different amplitude in two (or several) wells with similar epicentral distances.

As has discussed above, we can conclude: (1) the poro-elastic theory can mainly be applied to the area with epicentral distance < ~1.5 fault rupture length, during this area the coseismic water level changes are mostly induced by the undrained dilatation and consolidation of the sediments, and the difference of the water level changes in wells with similar epicentral distances is mostly related to the difference of the Skempton's coefficient *B* of those wells (group a, b, c). (2) When the poro-elastic theory can only be applied to one of the wells with similar epicentral distances, the water level change of the other well is usually much larger and more gradual, and we may infer the water level change of the other well is induced by the earthquake shaking, which is caused by the transition of the seismic waves (group d, e, g). (3) When the area with epicentral distance > ~1.5 fault rupture length, we may assume those water level changes may be caused by the transition of the seismic waves (group f and group h to l).

Besides, there may be some other mechanisms of the water level change, such as: mobilization of gas bubbles, (Roeloffs, 1998), fracture of an impermeable fault (King et al., 1999), fracture clearing (Brodsky et al., 2003).That needs to be clarified in our further study.

Data and Resources

Data used in this paper were collected using a classified network of the China Earthquake Networks Center and cannot be released to the public.

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Figure Captions:

Table 1. Dynamic deformation parameters of rocks. The range of the dynamic elastic modulus and dynamic Poisson's ratio are referred to Rock Mass Mechanism (Liu and Tang, 1998). From those parameters we calculate the range of the Dynamic shear modulus according to the formula $G = \frac{E}{2(1+\sigma)}$, and estimate the rough value of the dynamic shear modulus. Approximately, we choose the mean value.

Table 2. Epicentral Distances, Water Level Changes, Volume Strain Changes, Lithology and Shear Modulus for the stations separated into 12 groups (group a to group l). The difference of the epicentral distances of wells in each group is less than 16.68 kilometers (0.15 degrees). The volume strain change is calculated according to Okada's dislocation model (Huang, 2008). "-" means water level decrease in the water level change column and means compression in the volume strain change column. B^* is the value of the Skempton's coefficient of those well rocks, as we supposing the poro-elastic theory can be applied to the aquifer of those 27 wells. *B* is the real Skempton's coefficient as we have judged the poro-elastic theory can be applied to those wells. We use "/" to indicate those water level changes which can not be explained with the poro-elastic theory. We define $q = \frac{\Delta h}{\Delta h^*}$, it

represents ratio of the observed water level change Δh and the water level change Δh^* calculated from the poro-elastic theory.

Figure 1. Correlation coefficient of water level with solid tide, barometric pressure and volume strain for Changping station from January 1, 2008 to May 11, 2008 in the frequency-domain (Lai et al, 2009).

Figure 2. Raw hourly water level data and tidal strain data (a); Water level and the tidal strain after removing linear trend (b); Frequency domain analysis of the water level and the tidal strain (c); Distilled frequency of M2 wave from the water level and the tidal strain (d) (Zhang et. al, 2009).

Figure 3. Those 27 wells which can form groups that have the similar epicentral distance in mainland China. The serial number is in accordance with the number listed in table 2. The base map comes from the Mapseis software programmed by Shengle Li and the fault is plotted by Qidong Deng.

Figure 4. The spatial distribution of the static volume strain change of Wenchuan earthquake, which is calculated according to elastic half-space dislocation model (Okada, 1992). The solid line indicates inflation, while the dashed line represents compression. The pentagram is the epicenter of the Wenchuan earthquake, and the triangles represent the distributed 27 stations. Parameters of the focal mechanism: trend, 229°; angle of inclination, 35°; angle of slide,138°; depth, 15km; rupture length, 300km; width, 40km; slide range, 447cm.

Figure 5. Original water level changes of those wells (well: 1-27). It is the same with the description of Huang (2008): the sequential number of y-coordinate depends on the type of the well, "sequential

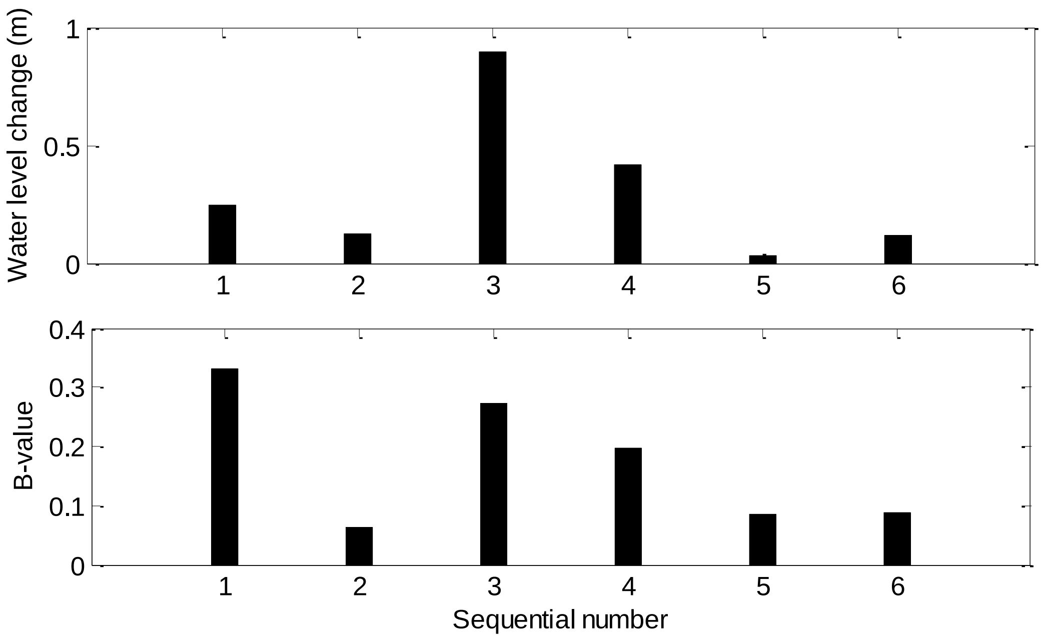
number increase from low to high" indicates an artesian well, and the free water surface is higher than the artesian discharge point or the ground, the coordinate value means the height from the free water surface to the artesian discharge point or to the ground. "Sequential number decrease from low to high" indicates a non-artesian well, and the coordinate value means the depth from the free water surface to the ground. All the ascendant patterns in the picture indicate water level ascending, while all those descendent patterns in the picture indicate water level descending.

Figure 6. Water level changes and *B* values plotted according to the serial number of those 6 wells. In each group (group a, b, c), large pre-earthquake *B* values come with large co-seismic water level changes.

Table 2. Epicentral Distances, Water Level Changes, Volume Strain Changes, Lithology and Shear Modulus for the stations separated into 12 groups (group a to group l). The difference of the epicentral distances of wells in each group is less than 16.68 kilometers (0.15 degrees). The volume strain change is calculated according to Okada's dislocation model (Huang, 2008). "-" means water level decrease in the water level change column and means compression in the volume strain change column. B^* is the value of the Skempton's coefficient of those well rocks, as we supposing the poro-elastic theory can be applied to the aquifer of those 27 wells. B is the real Skempton's coefficient as we have judged the poro-elastic theory can be applied to those wells. We use "/" to indicate those water level changes which can not be explained with the poro-elastic theory. We define $q = \frac{\Delta h}{\Delta h^*}$, it represents ratio of the observed water level change Δh and the water level change Δh^* calculated from the poro-elastic theory.

Serial Number	**Group**	**Station**	**Epicentral Distance D (km)**	**Water Level Change (m)**	**Volume Strain Change /10-9**	**Lithology**	**Shear Modulus G (Gpa)**	**B***	**q**	**B**
1	a	Dazu	185.4687	-0.25	100.4	Sandstone	8	0.331	8.01399	0.331
2	a	Rongchang	186.4838	-0.127	135.5	Sandstone	8	0.062	2.80831	0.062
3	b	Beibei	209.4532	-0.9	54.06	Sandstone	8	0.273	29.0513	0.273
4	b	Nanxi	217.7074	-0.42	163.6	Sandstone	8	0.197	10.5581	0.197
5	c	Xichang03	342.2935	0.03	-32.35	Graniton	36	0.084	1.21427	0.084
6	c	Xichangtail	350.68	0.119	-27.9	Graniton	36	0.087	5.39227	0.087
7	d	Shangrao	379.473	-0.015	0.3169	Quartzite	20	0.0275	**3093.33**	/
8	d	Luguhu	384.256	0.022	-27.28	Limestone	20	0.1862	0.84551	0.186
9	e	Qingshuiwe	425.681	0.02	-19.62	Sandstone	8	0.087	5.31599	0.087
10	e	Jinyangkou	430.448	0.835	-9.153	Limestone	20	0.1856	95.955	/
11	f	Xiaxian	465.8363	0.106	-3.503	Gneiss	40	0.0339	85.3015	/
12	f	Luonan	473.9955	0.07	-6.082	Limestone	20	0.0296	75.9071	/
13	g	Linxia	521.5619	-0.153	-0.7463	Psephite	4	0.4116	**-503.22**	/
14	g	Panzhihua	527.4969	0.068	-9.513	Diorite	30	0.0412	22.8225	0.041
15	h	Haiyuan	606.2586	-0.036	-6.952	Sandstone	8	0.1117	**-21.034**	/
16	h	Jiujiang	623.3212	0.072	0.3121	Sandstone	8	0.1193	**-877.35**	/
17	h	Guyuanzhei	638.6394	-0.026	-6.383	Sandstone	8	0.0073	**-252.82**	/
18	h	Kunming	650.7373	0.012	-1.245	Limestone	20	0.0992	113.808	/
19	h	Lasa	661.047	0.005	0.3116	Granite	28	0.0074	**-297.21**	/
20	i	Baoshan	793.4069	0.0410	-4.915	Sandstone	8	0.018	210.262	/
21	i	Kaiyuan	799.662	-0.155	-0.0835	Limestone	20	0.1977	**-1833.9**	/
22	j	Huangmeid	848.861	0.124	0.2208	Sandstone	8	0.0748	**-3406.4**	/
23	j	Lingwudaqi	856.022	0.053	-2.723	Sandstone	8	0.0605	145.964	/
24	k	Guigangdoi	899.981	-0.014	1.943	Sandstone	8	0.0722	45.2783	/
25	k	Guiping	900.8791	0.575	2.068	Sandstone	8	0.1768	**-713.52**	/
26	l	Jining	1131.181	0.012	-0.8496	Whinstone	48	0.0087	147.384	/
27	l	Qixian	1146.9055	0.831	-1.944	Limestone	20	0.2462	338.953	/

Figure 6. Water level changes and *B* values plotted according to the serial number of those 6 wells. In each group (group a, b, c), large pre-earthquake *B* values come with large co-seismic water level changes.



Mechanism of Different Co-Seismic Water Level Changes in Wells with Similar Epicentral Distances of Intermediate Field

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Abstract

Water level changes at different monitoring stations are observed during the Wenchuan earthquake (Ms8.0) in the Chinese mainland. In the intermediate field, we observed co-seismic water level changes of different amplitude in wells with similar epicentral distances. In order to study about the mechanism of those co-seismic water level changes, we calculated the static strain change with the Okada's dislocation model. Compare the calculated co-seismic water level change based on the poro-elastic theory with the observed water level change, we can judge whether the poro-elastic theory can be applied to the aquifer of the well, from which we find that: When the water level change of those wells can be explained by the poro-elastic theory (from our research we find the poro-elastic theory can be applied to the area with epicentral distance < ~1.5 fault rupture length), the difference of the water level change in wells with similar epicentral distances is mostly related to the difference of the Skempton's coefficient *B*. Otherwise, the water level change

may be induced by the transition of the seismic waves, since it is usually larger than the one induced by the un-drained dilatation and consolidation, and changes more gradual.

Introduction

Several types of earthquake induced groundwater level changes and corresponding mechanisms have been recognized for decades. In the near field (generally, epicentral distance D between 0-100 km), most documented water level shows abrupt (step-like) coseismic changes (Wakita 1975; Quilty and Roeloffs, 1997; Wang et al., 2001, 2004; Chia et al., 2001; Wang and Chia, 2008). Undrained dilatation and consolidation of the sediments may be responsible for the step-like water level changes in the near field, and can often be quantitatively related to the poroelastic response to the earthquake’s static strain. In the intermediate field (epicentral distance D between 100-1000 km), most documented changes are gradual and can persist for days or weeks. These are coined by Roeloffs (1998) as the ‘sustained’ water level changes, and an earthquake-enhanced permeability may be responsible for this intermediate field phenomenon (Wang and Chia, 2008). In the far field (epicentral distance D larger than 1000 km), transient oscillations of the water level have been documented, and sustained water level changes also have been observed. There are several existing models for far-field coseismic pore pressure changes: mobilization of gas bubbles, (Roeloffs, 1998), shaking induced dilatancy (Bower and Heaton, 1978), fracture of an impermeable fault (King et al., 1999), fracture clearing (Brodsky et al., 2003), and shaking induced by surface waves (West et al., 2005; Sil and Freymueller, 2006).

Investigation of coseismic water level changes has been of scientific interest for decades (Wang and Manga, 2010). Groundwater level changes following earthquakes can affect water supply; seismic waves can affect oil well production, and it has been suggested that in some cases the induced seismicity can stimulate oil production (Beresnev and Johnson, 1994). Earthquake-induced fluid pressure changes are hypothesized to control the timing and/or location of the aftershocks and trigger seismicity (Hill et al., 1995; Gomberg, 1996). Finally, these groundwater level changes could also be related to the hydrologic earthquake precursors (Roeloffs, 1998).

In this paper we calculate Skempton's coefficient *B* from the poroelastic relationship between water level changes and tidal strain using data prior to the earthquake. Further analysis of the water level data from the Groundwater Monitoring Network (GMN) (see Data and Resources Section) is done during the Wenchuan earthquake for intermediate field. A relation between the amplitude of the water level , the earthquake magnitude and distance is developed by Roeloffs (1998) for the "sustained" water level changes. To develop this relationship, different intermediate field earthquakes are used. Several authors have obtained similar empirical relations between water level change, epicentral distance, and the earthquake magnitude (Matsumoto et al., 2003; Yang et al., 2005; Sil and Freymueller, 2006).

In addition to the above observation, we find that the size of the water level change at GMN stations in the intermediate field is not only related to the earthquake magnitude and the epicentral distance. Several wells with similar epicentral distances have different amplitude of co-seismic water level changes, and some of those wells even stay close to each other in one fault. We calculated the static strain change with the Okada's dislocation model. Supposing the poro-elastic theory can be applied to all of those wells, based on that we calculated the Skempton's coefficient *B* of all those wells. With the calculated static strain change and the Skrmpton's coefficient *B*, we derived the co-seismic water level changes. Compare the calculated co-seismic water level change based on the poro-elastic theory with the observed water level change, we can judge whether the poro-elastic theory can be applied to the aquifer of the well, so as to study about the mechanism of those co-seismic water level changes.

In this paper, we find that: the poro-elastic theory can mainly be applied to the area with epicentral distance < ~1.5 fault rupture length, during this area the coseismic water level changes are mostly induced by the undrained dilatation and consolidation of the sediments, and the difference of the water level change in wells with similar epicentral distances is mostly related to the difference of the Skempton's coefficient *B* of those wells. Large *B*-values come with large changes in water level. This phenomenon is in accordance with the poro-elastic theory. When the poro-elastic theory can only be applied to one of the wells with similar epicentral distances, usually the water level change of the other well is more gradual and with much larger amplitude, we may infer it is induced by the earthquake shaking, which is caused by the transition of the seismic waves.

Theory

Skempton's coefficient *B* is a significant pore-fluid parameter in poroelastic theory. A poroelastic material consists of an elastic matrix containing interconnected fluid saturated pores. Fluid saturated crust behaves as a poroelastic material to a good degree of approximation.

Rice and Cleary (1976) summarized the following equations for a linearly elastic isotropic porous medium, which are the building blocks of the poroelastic theory:

$$2G\varepsilon_{ij} = \sigma_{ij} - \frac{\nu}{1+\nu}\sigma_{kk}\delta_{ij} + \frac{3(\nu_u - \nu)}{B(1+\nu)(1+\nu_u)} p\delta_{ij}, \qquad (1)$$

$$m - m_0 = \frac{3\rho(\nu_u - \nu)(\sigma_{kk} + 3p/B)}{2GB(1+\nu)(1+\nu_u)}. \qquad (2)$$

Here m-m_o is the change of the fluid mass, ε_{ij} is the strain tensor, σ_{ij} is the stress tensor, δ_{ij} is the Kronecker delta function, G is the shear modulus, ρ is the density of the fluid, *B* is the Skempton's coefficient, p is the pore pressure, ν is the Poisson's ratio, and ν_u is the "undrained" Poisson's ratio. Rice and Cleary (1976) describe equation 1 as a stress balance equation and equation 2 as a mass balance equation.

For the undrained condition, the poroelastic effect on the crust can be obtained by putting m-m_o =0 in equation 2, and therefore we obtain:

$P = -B\sigma_{kk} / 3$ or $\Delta p = -B\Delta\sigma_{kk} / 3$. (3)

Equation 3 says under "undrained" condition, the change in fluid pressure (Δp) is proportional to the change in mean stress ($\Delta\sigma_{kk}$/3). This is the mechanism of water level changes for poroelastic material. (p=ρg h, where h is the water column height, g is the acceleration due to gravity and ρ is the density of water).

According to equation 3, Skempton's coefficient *B* can be qualitatively defined: In the "undrained" condition, *B* is the ratio of the induced pore pressure divided by the change in mean stress (Wang, 2000). *B* governs the magnitude of water level changes due to an applied stress since pore pressure is directly proportional to water level. The value of *B* is always between 0 and 1. When *B* is 1, the applied stress is completely transferred into changing pore pressure. *B* equals 0 indicates no change in pore pressure after applying the stress. Thus a low value of *B* indicates the stiff rock matrix that supports the load with low coupling to the fluid (Nur and Byerlee, 1971). Laboratory studies indicate the value of *B* depends upon the fluid saturated pore volume of the sample (Wang, 2000).

Equation 3 can be expressed in terms of tidal strain as well (Roeloffs, 1996):

$$\Delta h = -\frac{2GB(1+\nu_u)}{3\rho g(1-2\nu_u)}\Delta\varepsilon_t \quad (4)$$

Equation 4 shows that water level changes proportionally in a poroelastic material under the influence of tidal strain (ε_t). Here Δh is the change in height of water level, and $\Delta\varepsilon_t$ is the corresponding tidal strain change (Sil, 2006).

From equation 4 we obtain:

$$B = -\frac{3\rho g(1-2\nu_u)}{2G(1+\nu_u)}\frac{\Delta h}{\Delta\varepsilon_t} \quad (5)$$

With equation (5) we can get the value of *B* with water level and tidal strain. However, the calculation must be on the strict premise of the undrained condition, the good correlation ship between the water level and the tidal strain and should not be influenced by the other factors.

For the effect of the solid tide on the crust, when the wavelength of the tidal strain is much larger than the size of the aquifer, we can suppose the aquifer system is undrained (Huang, 2008). The wavelength of the M2 wave is about 2 406 329 km ($\lambda=\omega\times r\times T$, $\omega=1.4\times10{-}4$/s is the angular frequency of M2 wave, r=384 400 km is the distance from the earth to the moon, T=745.236 min is the period of the M2 wave), which is much larger than the size of the radius of the Earth, and is definitely much larger than the thickness of the aquifer systems of those wells. Thus, the effect of the M2 wave in the crust can meet with the undrained condition (Zhang et. al, 2009). Besides, those wells can record clear tidal strains and as we calculate the phase lags between the water levels and the tidal strains are small, thus the wells can meet with the undrained condition well. In the M2 wave frequency domain the water level and the tidal strain have a good correlationship, we just set the Changping station as an example to see the correlationship clearly (Figure 1). We can see in the M2 wave frequency domain the correlationship between the tidal strain and the water level approaches 1, which means a good correlationship between them. Besides, the M2 wave is hardly influenced by atmospheric pressure. Since that, we distill the frequency domain of the M2 wave from the water level and the tidal strain by using band-pass filter (the frequency of the M2 wave is 0.0805114 h^{-1}) to calculate the Skempton's coefficient *B* (Figure 2). Disposing the obtained frequency domain of the M2 wave by IFFT (inverse fast Fourier transform) and adjusting their phase, through the least square fit and putting the results into equation (5), we can finally derive *B*. More details of the method are explained by the paper "Research on Skempton's coefficient *B* based on the observation of groundwater of Changping station" (Zhang et. al, 2009). All the Water level observations come from the sensor of water level, while tidal strain data are calculated via Mapsis software, which is programmed by Shengle Li.

Methods

Water level changes at different monitoring stations are observed during the Wenchuan earthquake (Ms8.0) in the Chinese mainland. We aim at exploring the mechanism of those co-seismic water level changes of different amplitude in wells with similar epicentral distances.

We only find 27 wells which can form groups that have similar epicentral distance (within a range of less than 0.15 degrees or 16.68 km) in the intermediate field of mainland China (Figure 3). One well (Weinanshuangwang) has been deleted since we can not confirm the range of the shear modulus of its lithology (Sand clay). We divided those 27 wells into twelve groups (group a to group l), each group has a specific range of epicentral distance (Table 2). As show in Figure 3, wells in group a (well 1, 2), b (well 3, 4), c (well 5, 6) , h (well 15, 17) and k (well 24, 25) stay close with each other.

First of all, we suppose the poro-elastic theory can be applied to all of those 27 wells. We apply the method of *B*-value calculation to those 27 wells. Pre-earthquake analysis is carried out using data from May 2, 2008 to May 10, 2008 to obtain the $B*$ values (Table 2). Calculation is performed using $\rho = 1000 kg/m^3$, $g = 9.8 m/s^2$, and $\nu_u = 0.29$. Since the shear modulus will change with the change of the stress, we can hardly get the in suit value of the shear modulus of those wells by experiment, which is as hard as getting the in suit Skempton's coefficient *B*. We have investigated the geology of each well and referred to the Rock Mass Mechanism (Liu and Tang, 1998), using the dynamic elastic modulus and dynamic Poisson's ratio to estimate the range of the shear modulus of those rocks, and approximately choose the mean value (Table 1).

Then, we must check if the prediction of the poro-elastic theory is consistent with the observed water level changes, so as to check whether the poro-elastic theory can be applied to the aquifer of the 27 wells. Since that, we show the co-seismic volume strain changes in Table 2, which is already calculated by Fuqiong Huang in her PhD Dissertation with Okada's dislocation model (Huang, 2008). We have plotted those wells with the spatial distribution of the static volume strain change of Wenchuan earthquake (Figure 4), and also plotted the original water level change of those 27 wells in Figure 5. From equation (3) we can obtain $\Delta h^* = -\frac{B^* \cdot \Delta\sigma_{kk}}{\rho g} = -\frac{B^* \cdot (E^* \cdot \Delta\varepsilon_{kk})}{\rho g}$, we calculated the water level change from B^* and the static stress change $E^* \cdot \Delta\varepsilon_{kk}$. We can judge whether the well aquifer can fit for the poro-elastic theory just by comparing the observed water level change Δh and the water level change Δh^* calculated from the poro-elastic theory. Define $q = \frac{\Delta h}{\Delta h^*}$, we calculated *q* values of those 27 wells (Table 2). As show in Table 2, when the value of *q* is too large (it means there are huge differences between the theory value Δh^* and the real value Δh) or $q < 0$ (it implies the sign of the water level change is not consistent with the direction of the volume strain change, and is not caused by the un-drained consolidation or dilatation) , the well aquifer may not fit for the poro-elastic theory, and we should not use the poro-elastic theory to explain the mechanism of water level change. Since that, the B^* value which is calculated based on the poro-elastic theory will be invalid.

Generally, according to the *q* and static strain change values in Table 2 and also take the patterns of those co-seismic water level changes in Figure 5 into consideration, we take *q*=30 as the threshold value, when q<30 we suppose the poro-elastic theory can be applied to the well aquifer, otherwise if q>30 the mechanism of the water level change may not be the static strain change, thus the poro-elastic theory may not be applied to the well aquifer.

Firstly, as show in Table 2, except for well 7, q values of well 1 to well 9 are all smaller than 30, they are much smaller than those q values of well 10 to well 27 (except well 14). The mean q value of those wells (well 1 to 9, discarding well 7) is $\bar{q} = 7.90$, which is relatively acceptable (The Okada's dislocation model is based on the assumption that the whole land is isotropic and homogeneous, and does not consider about the geology conditions. However, there are several faults between those regions and the epicenter, so the medium is not uniform, and the volume strain change $\Delta\varepsilon_{kk}$ got from this model will definitely have some differences from the real condition (Figure 3). Besides, when we calculate the B^* value, we use the mean value of the shear modulus G, it may be different from the real G value (Table 1). Inevitably, there must be some differences between the water level change calculated from the poro-elastic theory $\Delta h^* = -\frac{B^* \cdot \Delta\sigma_{kk}}{\rho g} = -\frac{B^* \cdot (E^* \cdot \Delta\varepsilon_{kk})}{\rho g}$ and the observed water level change Δh, thus the mean value $\bar{q} = 7.90$ is relatively acceptable).

Secondly, the static strain values of well 10-27 are obviously smaller than that of well 1-9 (except well 7) (Table 2), the seismic energy density in the relatively far field (D> 430 km) may be too small to initiate un-drained consolidation and dilatation, a distinct mechanism is required to explain the water level changes at such distances.

From the analysis above, we may just get 9 wells which can fit for the poro-elastic theory in the intermediate field (well: 1, 2, 3, 4, 5, 6, 8, 9, 14) (Table 2). Since that the B values of those 9 wells are valid, meanwhile we use "/" to indicate the invalid B values (Table 2).

Mechanism analysis

Among those 9 wells to which the poro-elastic theory can be applied, only 6 can form groups with similar epicentral distances (well: 1, 2, 3, 4, 5, 6). We find that large pre-earthquake *B* values correspond to large magnitude of co-seismic water level changes, this phenomenon exists in those 6 wells (Figure 6). We use poro-elastic theory to analyze the mechanism of this phenomenon.

From equation (3) we can see the water level change $\Delta h = \Delta p / \rho g$ is related to the static stress change $\Delta \sigma_{kk} / 3$ and the Skempton's coefficient *B*. From Table 2 we can see, the amplitude of the co-seismic water level change in each group is not always in accordance with the amount of the static strain change. Set group a as an example, the volume strain change of well 2 is larger than that of well 1, but the amplitude of the co-seismic water level change of well1 is larger than that of well 2. This phenomenon widely exists in group a, b, and c. This obviously shows that, *B* governs the magnitude of water level change induced by the applied stress.

Large *B*-values come with large changes in water level. This phenomenon is in accordance with the poro-elastic theory. When *B*-values are high, the stiff rock matrix supports the load with high coupling to the fluid (Nur and Byerlee, 1971) , the applied stress is mostly transferred into changing pore pressure, which leads to relatively large changes in water level. When *B*-values are low, the stiff rock matrix supports the load with low coupling to the fluid (Nur and Byerlee, 1971) , the applied stress can be easily transferred outside the aquifer system without increasing the pore pressure resulting in small water level changes (Sil, 2006). This can be used to explain: why two wells stay close with each other, but the amplitudes of their co-seismic water level changes are different.

In the other 3 groups (group: d, e, g), the water level changes in 3 wells (well 8, 9, 14) can be explained by the poro-elastic theory, while the other 3 can not (well 7, 10, 13) (Table 2). As show in Table 2, in group e and g, the water level change of the well to which the poro-elastic theory can be applied is smaller than the other one. Therefore, we can imply that the water level change in the 2 wells (well 10, 13) may be induced by the transfer of the seismic waves. As has been reported, earthquakes can produce sustained water level changes in certain distant wells that are often orders of magnitude larger than can be explained by static stress changes (Bower and Heaton, 1978). The shape of water level change in well 7 is sustained, although the amplitude is not large, we may assume that it is also caused by the transition of the seismic waves.

Discussion

Water level changes in regions to which poro-elastic theory can be applied are consistent with the volume strain changes. That means, when the volume strain change is positive (dilatational) the water level decrease, and when the volume strain change is negative (compressional) the water level increase (Table 2). Among those 27 wells the water level change of 8 wells are not consistent with the volume strain change (well: 13, 15,16,17,19, 21, 22, 25), and those wells are distributed in different areas in the Chinese Mainland (Figure3). As we calculated the q values of those wells, the result is in accordance with the above result. In those wells (well: 13,15,16,17, 19, 21, 22, 25) the q values are obviously much larger than the others, it means that the poro-elastic theory can not be applied to those wells, and the water level change in those wells are definitely not caused by the static volume strain change.

For intermediate distance earthquakes, several authors previously obtained similar empirical equations (shown below) relating water level change, epicentral distance, and magnitude of the earthquakes (Roeloffs, 1998; Matsumoto et al., 2003; Yang et al., 2005; Sil and Freymueller, 2006). And this empirical equation is based on the mechanism of shaking induced water level change. They attribute the magnitude of the water level change to two major impact factors: earthquake magnitude and epicentral distance. The empirical relation found by them can be written as:

$$\log_{10} \Delta h_i = w_1 M + w_2 \log_{10} D + w_3. \quad (6)$$

In this equation w_1, w_2 , and w_3 are constants, Δh_i is the size of the water level change in centimeters, M is the earthquake magnitude, and D is the well- hypocenter distance in kilometers (Roeloffs, 1998). The importance of equation 6 is that, for intermediate distances, it can explain earthquake induced water level changes, where poro-elastic theory generally is not applicable.

However, it is hard to explain water level changes in group f and group h to l with equation 6. According to equation 6, with the same magnitude and similar epicentral distances, water level changes in the same group (in group f and group h to l) are expected to be similar, but as shown in Table 2 the amplitude of those water level changes are different.

From our analysis, we think the unfitness of equation 6 lies in the shape conformity of those co-seismic water level changes and the geology condition of those wells. As studied by Roeloffs (1998), and Yang et al. (2005), they just use one specific well with different earthquakes to obtain equation 6, all those co-seismic water level changes observed by them are ascending. Matsumoto et al. (2003) use the co-seismic water level changes induced by different earthquakes in the Haibara well to obtain equation 6, and all those co-seismic water level changes are descending. Sil and Freymueller (2006), use the co-seismic water level changes induced by the Sumatra-Andaman earthquake in wells with the same geology condition in one specific area to obtain this empirical equation, those co-seismic water level changes are all rising. Thus, same geology condition and shape conformity of those co-seismic water level changes (all rising or all descending) may be the premise of empirical equation 6.

As discussed above, equation 6 can be used to explain water level changes in well 15 and 17 of group h. The amplitude of the water level changes in the two wells are similar (Table 2). Well

15 and 17 lie in the similar area (Figure 3) with the same geology condition (Table 2), and both of them are increasing. For well 24 and 25, though lie in the same area with similar geology, one of the water level decrease while the other increase，which may lead to the unfitness of equation 6, thus the amplitude of water level changes in well 24 and 25 are different. On the other hand, the estimated co-seismic water level changes calculated from equation 6 are not always in conformity with the observed water level changes (Roeloffs, 1998; Matsumoto et al., 2003; Yang et al., 2005; Sil and Freymueller, 2006), sometimes there will be a little difference, which may lead to the different amplitude of water level changes in the same group.

Besides, the obscurity may also be caused by the large distances between those wells and the epicenter, and there are lots of faults, so the medium is not uniform. The Okada's dislocation model is based on the assumption that the whole land is isotropic and homogeneous. Therefore, there may be huge differences between the calculated volume strain change and the real value in those wells (well 19~27, except for well 24, 25), thus it is possible that their q values are not accurate. Since that, it is hard for us to study the mechanism of the water level changes in those wells based on the q values, and we should research those water level changes in further studies.

As discussed earlier, the shear modulus G will change with the change of the stress, and it is found to be the function of the Skempton's coefficient B (Berryman, 2004). We can hardly get the in suit value of the shear modulus of those wells by experiment. Thus there may be ranges of uncertainty in B values getting from mean G values, and this needs to be further studied.

We couldn't find data from near field (0< D <100 kilometers) wells with the similar epicentral distance during the Wenchuan earthquake.

Magnitude of the Wenchuan earthquake is relatively large (Ms 8). Therefore, even without computing, we can expect that the static strain field from the earthquake will affect a relatively large area (The area is about 430 kilometers away from the epicenter according to our study in this paper). Thus we assume that our observation is not contradicting any existing theory of earthquake induced water level changes. For the relatively far field, shaking induced by the transition of the seismic waves may be the major mechanism of the co-seismic water level changes.

Conclusions

In this paper we discussed the mechanism of the co-seismic water level changes of different amplitude in two (or several) wells with similar epicentral distances.

As has discussed above, we can conclude: (1) the poro-elastic theory can mainly be applied to the area with epicentral distance $< \sim 1.5$ fault rupture length, during this area the coseismic water level changes are mostly induced by the undrained dilatation and consolidation of the sediments, and the difference of the water level changes in wells with similar epicentral distances is mostly related to the difference of the Skempton's coefficient B of those wells (group a, b, c). (2) When the poro-elastic theory can only be applied to one of the wells with similar epicentral distances, the water level change of the other well is usually much larger and more gradual, and we may infer the water level change of the other well is induced by the earthquake shaking, which is caused by the transition of the seismic waves (group d, e, g). (3) When the area with epicentral distance $> \sim 1.5$ fault rupture length, we may assume those water level changes may be caused by the transition of the seismic waves (group f and group h to l).

Besides, there may be some other mechanisms of the water level change, such as: mobilization of gas bubbles, (Roeloffs, 1998), fracture of an impermeable fault (King et al., 1999), fracture clearing (Brodsky et al., 2003).That needs to be clarified in our further study.

Data and Resources

Data used in this paper were collected using a classified network of the China Earthquake Networks Center and cannot be released to the public.

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Figure Captions:

Table 1. Dynamic deformation parameters of rocks. The range of the dynamic elastic modulus and dynamic Poisson's ratio are referred to Rock Mass Mechanism (Liu and Tang, 1998). From those parameters we calculate the range of the Dynamic shear modulus according to the formula $G = \frac{E}{2(1+\sigma)}$, and estimate the rough value of the dynamic shear modulus. Approximately, we choose the mean value.

Table 2. Epicentral Distances, Water Level Changes, Volume Strain Changes, Lithology and Shear Modulus for the stations separated into 12 groups (group a to group l). The difference of the epicentral distances of wells in each group is less than 16.68 kilometers (0.15 degrees). The volume strain change is calculated according to Okada's dislocation model (Huang, 2008). "-" means water level decrease in the water level change column and means compression in the volume strain change column. B^* is the value of the Skempton's coefficient of those well rocks, as we supposing the poro-elastic theory can be applied to the aquifer of those 27 wells. *B* is the real Skempton's coefficient as we have judged the poro-elastic theory can be applied to those wells. We use "/" to indicate those water level changes which can not be explained with the poro-elastic theory. We define $q = \frac{\Delta h}{\Delta h^*}$, it

represents ratio of the observed water level change Δh and the water level change Δh^* calculated from the poro-elastic theory.

Figure 1. Correlation coefficient of water level with solid tide, barometric pressure and volume strain for Changping station from January 1, 2008 to May 11, 2008 in the frequency-domain (Lai et al, 2009).

Figure 2. Raw hourly water level data and tidal strain data (a); Water level and the tidal strain after removing linear trend (b); Frequency domain analysis of the water level and the tidal strain (c); Distilled frequency of M2 wave from the water level and the tidal strain (d) (Zhang et. al, 2009).

Figure 3. Those 27 wells which can form groups that have the similar epicentral distance in mainland China. The serial number is in accordance with the number listed in table 2. The base map comes from the Mapseis software programmed by Shengle Li and the fault is plotted by Qidong Deng.

Figure 4. The spatial distribution of the static volume strain change of Wenchuan earthquake, which is calculated according to elastic half-space dislocation model (Okada, 1992). The solid line indicates inflation, while the dashed line represents compression. The pentagram is the epicenter of the Wenchuan earthquake, and the triangles represent the distributed 27 stations. Parameters of the focal mechanism: trend, 229°; angle of inclination, 35°; angle of slide,138°; depth, 15km; rupture length, 300km; width, 40km; slide range, 447cm.

Figure 5. Original water level changes of those wells (well: 1-27). It is the same with the description of Huang (2008): the sequential number of y-coordinate depends on the type of the well, “sequential

number increase from low to high" indicates an artesian well, and the free water surface is higher than the artesian discharge point or the ground, the coordinate value means the height from the free water surface to the artesian discharge point or to the ground. "Sequential number decrease from low to high" indicates a non-artesian well, and the coordinate value means the depth from the free water surface to the ground. All the ascendant patterns in the picture indicate water level ascending, while all those descendent patterns in the picture indicate water level descending.

Figure 6. Water level changes and *B* values plotted according to the serial number of those 6 wells. In each group (group a, b, c), large pre-earthquake *B* values come with large co-seismic water level changes.

Table 1. Dynamic deformation parameters of rocks. The range of the dynamic elastic modulus and dynamic Poisson's ratio are referred to Rock Mass Mechanism (Liu and Tang, 1998). From those parameters we calculate the range of the Dynamic shear modulus according to the formula $G = \frac{E}{2(1+\sigma)}$, and estimate the rough value of the dynamic shear modulus. Approximately, we choose the mean value.

Rock	**Dynamic Elastic Modulus (Gpa)** E*	**Dynamic Poisson's Ratio** σ*	**Dynamic Shear Modulus (Gpa) G**	**Rough value of dynamic Shear Modulus (Gpa)**
Sandstone	5.3 ~ 37.9	0.20 ~ 0.22	2.17 ~ 15.79	8
Graniton	63.4 ~ 114.8	0.20 ~ 0.21	26.20 ~ 47.83	36
Quartzite	20.4 ~ 76.3	0.23 ~ 0.26	8.10 ~ 31.02	20
Limestone	12.1 ~ 88.3	0.24 ~ 0.25	4.84 ~ 35.60	20
Gneiss	76.0 ~ 129.1	0.22 ~ 0.24	30.65 ~ 52.91	40
Granite	37.0 ~ 106.0	0.24 ~ 0.31	14.12 ~ 42.74	28
Whinstone	53.1 ~ 162.8	0.10 ~ 0.22	21.76 ~ 74.00	48
Diorite	52.8 ~ 96.2	0.23 ~ 0.34	19.7 ~ 39.11	30
Psephite	3.4 ~ 16	0.19 ~ 0.22	1.39 ~ 6.723	4

*see Liu, Y. R., and H. M. Tang (1998). Rock Mass Mechanics, Press of China University of Geosciences, Beijing, 112.

Table2

Table 2. Epicentral Distances, Water Level Changes, Volume Strain Changes, Lithology and Shear Modulus for the stations separated into 12 groups (group a to group l). The difference of the epicentral distances of wells in each group is less than 16.68 kilometers (0.15 degrees). The volume strain change is calculated according to Okada's dislocation model (Huang, 2008). "-" means water level decrease in the water level change column and means compression in the volume strain change column. B^* is the value of the Skempton's coefficient of those well rocks, as we supposing the poro-elastic theory can be applied to the aquifer of those 27 wells. B is the real Skempton's coefficient as we have judged the poro-elastic theory can be applied to those wells. We use "/" to indicate those water level changes which can not be explained with the poro-elastic theory. We define $q = \frac{\Delta h}{\Delta h^*}$, it represents ratio of the observed water level change Δh and the water level change Δh^* calculated from the poro-elastic theory.

Serial Number	**Group**	**Station**	**Epicentral Distance D (km)**	**Water Level Change (m)**	**Volume Strain Change /10-9**	**Lithology**	**Shear Modulus G (Gpa)**	**B^***	**q**	**B**
1	a	Dazu	185.4687	-0.25	100.4	Sandstone	8	0.331	8.01399	0.331
2	a	Rongchang	186.4838	-0.127	135.5	Sandstone	8	0.062	2.80831	0.062
3	b	Beibei	209.4532	-0.9	54.06	Sandstone	8	0.273	29.0513	0.273
4	b	Nanxi	217.7074	-0.42	163.6	Sandstone	8	0.197	10.5581	0.197
5	c	Xichang03	342.2935	0.03	-32.35	Graniton	36	0.084	1.21427	0.084
6	c	Xichangtail	350.68	0.119	-27.9	Graniton	36	0.087	5.39227	0.087
7	d	Shangrao	379.473	-0.015	0.3169	Quartzite	20	0.0275	**3093.33**	/
8	d	Luguhu	384.256	0.022	-27.28	Limestone	20	0.1862	0.84551	0.186
9	e	Qingshuiwe	425.681	0.02	-19.62	Sandstone	8	0.087	5.31599	0.087
10	e	Jinyangkou	430.448	0.835	-9.153	Limestone	20	0.1856	95.955	/
11	f	Xiaxian	465.8363	0.106	-3.503	Gneiss	40	0.0339	85.3015	/
12	f	Luonan	473.9955	0.07	-6.082	Limestone	20	0.0296	75.9071	/
13	g	Linxia	521.5619	-0.153	-0.7463	Psephite	4	0.4116	**-503.22**	/
14	g	Panzhihua	527.4969	0.068	-9.513	Diorite	30	0.0412	22.8225	0.041
15	h	Haiyuan	606.2586	-0.036	-6.952	Sandstone	8	0.1117	**-21.034**	/
16	h	Jiujiang	623.3212	0.072	0.3121	Sandstone	8	0.1193	**-877.35**	/
17	h	Guyuanzhen	638.6394	-0.026	-6.383	Sandstone	8	0.0073	**-252.82**	/
18	h	Kunming	650.7373	0.012	-1.245	Limestone	20	0.0992	113.808	/
19	h	Lasa	661.047	0.005	0.3116	Granite	28	0.0074	**-297.21**	/
20	i	Baoshan	793.4069	0.0410	-4.915	Sandstone	8	0.018	210.262	/
21	i	Kaiyuan	799.662	-0.155	-0.0835	Limestone	20	0.1977	**-1833.9**	/
22	j	Huangmeid	848.861	0.124	0.2208	Sandstone	8	0.0748	**-3406.4**	/
23	j	Lingwudaq	856.022	0.053	-2.723	Sandstone	8	0.0605	145.964	/
24	k	Guigangdon	899.981	-0.014	1.943	Sandstone	8	0.0722	45.2783	/
25	k	Guiping	900.8791	0.575	2.068	Sandstone	8	0.1768	**-713.52**	/
26	l	Jining	1131.181	0.012	-0.8496	Whinstone	48	0.0087	147.384	/
27	l	Qixian	1146.9055	0.831	-1.944	Limestone	20	0.2462	338.953	/

Figure1

Figure 1. Correlation coefficient of water level with solid tide, barometric pressure and volume strain for Changping station from January 1, 2008 to May 11, 2008 in the frequency-domain (Lai et al, 2009).

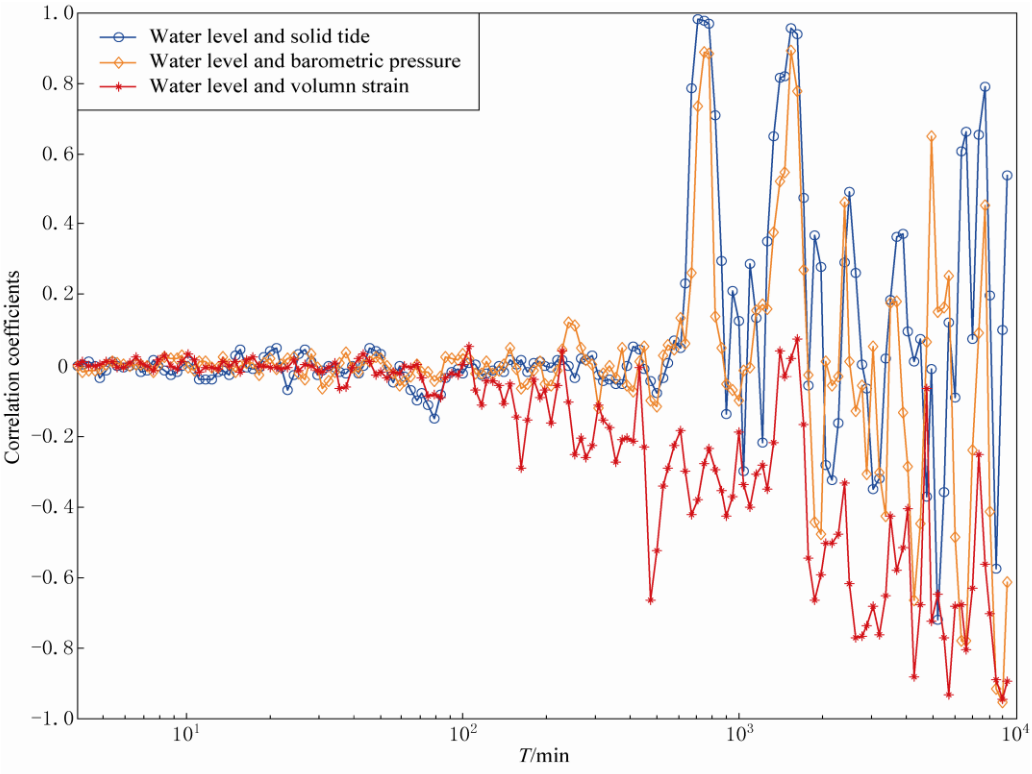


Figure2

Figure 2. Raw hourly water level data and tidal strain data (a); Water level and the tidal strain after removing linear trend (b); Frequency domain analysis of the water level and the tidal strain (c); Distilled frequency of M2 wave from the water level and the tidal strain (d) (Zhang et. al, 2009).

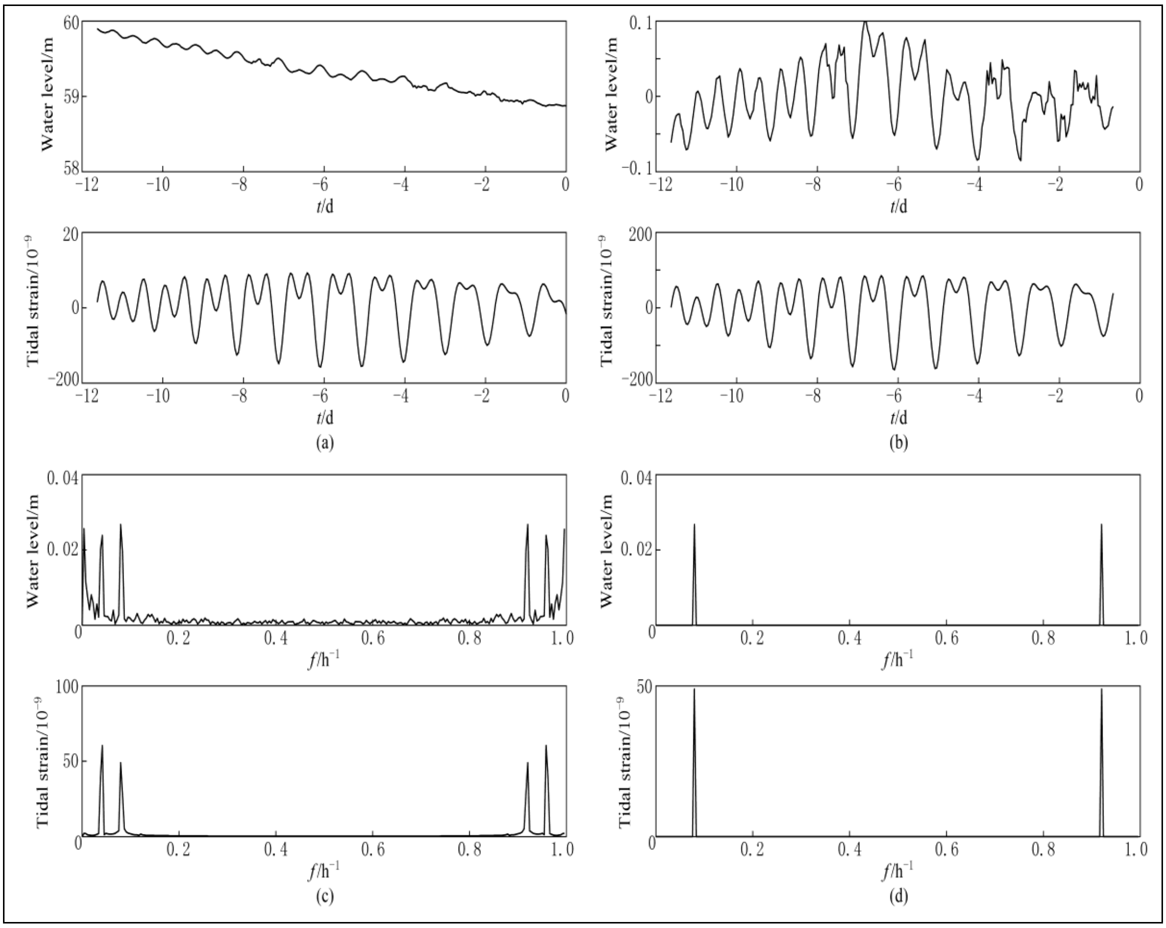


Figure3

Figure 3. Those 27 wells which can form groups that have the similar epicentral distance in mainland China. The serial number is in accordance with the number listed in table 2. The base map comes from the Mapseis software programmed by Shengle Li and the fault is plotted by Qidong Deng.

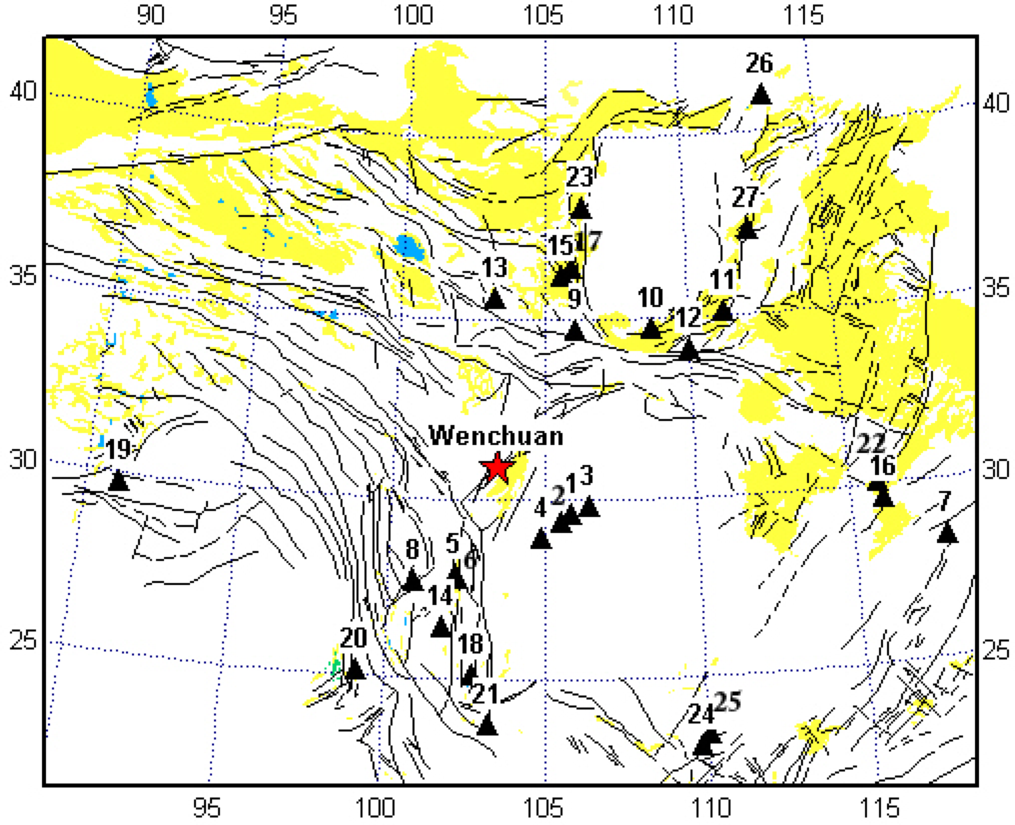


Figure4

Figure 4. The spatial distribution of the static volume strain change of Wenchuan earthquake, which is calculated according to elastic half-space dislocation model (Okada, 1992). The solid line indicates inflation, while the dashed line represents compression. The pentagram is the epicenter of the Wenchuan earthquake, and the triangles represent the distributed 27 stations. Parameters of the focal mechanism: strike(trend), 229°; angle of inclination, 35°; angle of slide,138°; depth, 15km; rupture length, 300km; width, 40km; slide range, 447cm.

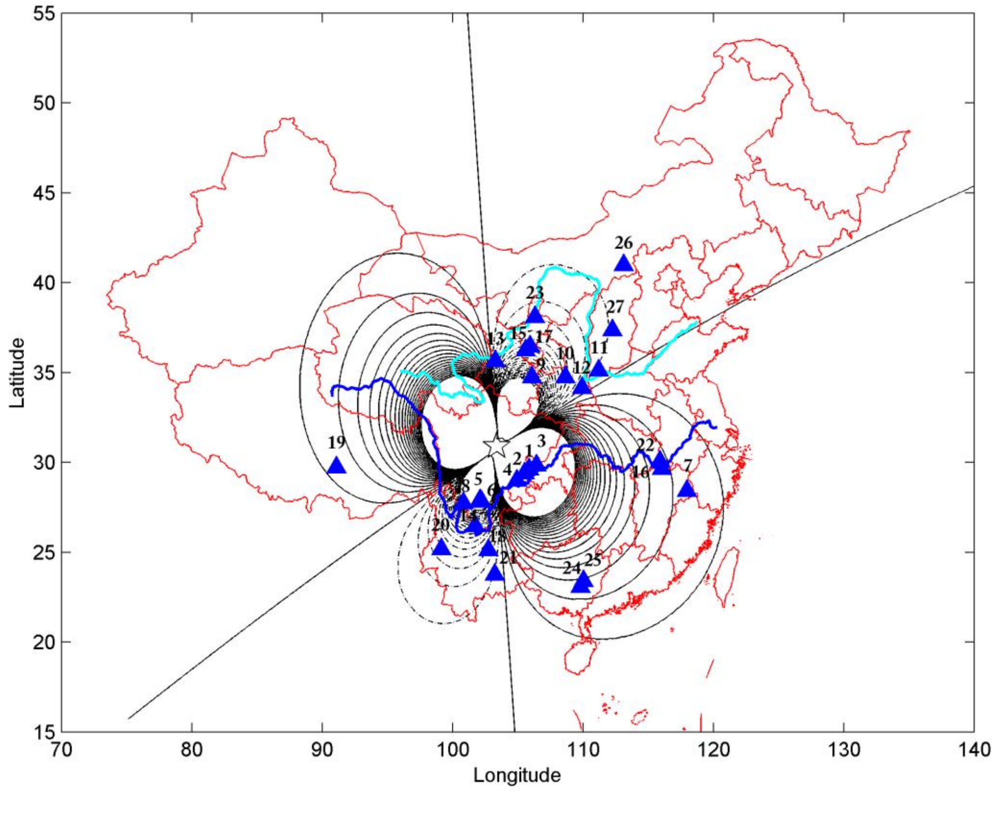
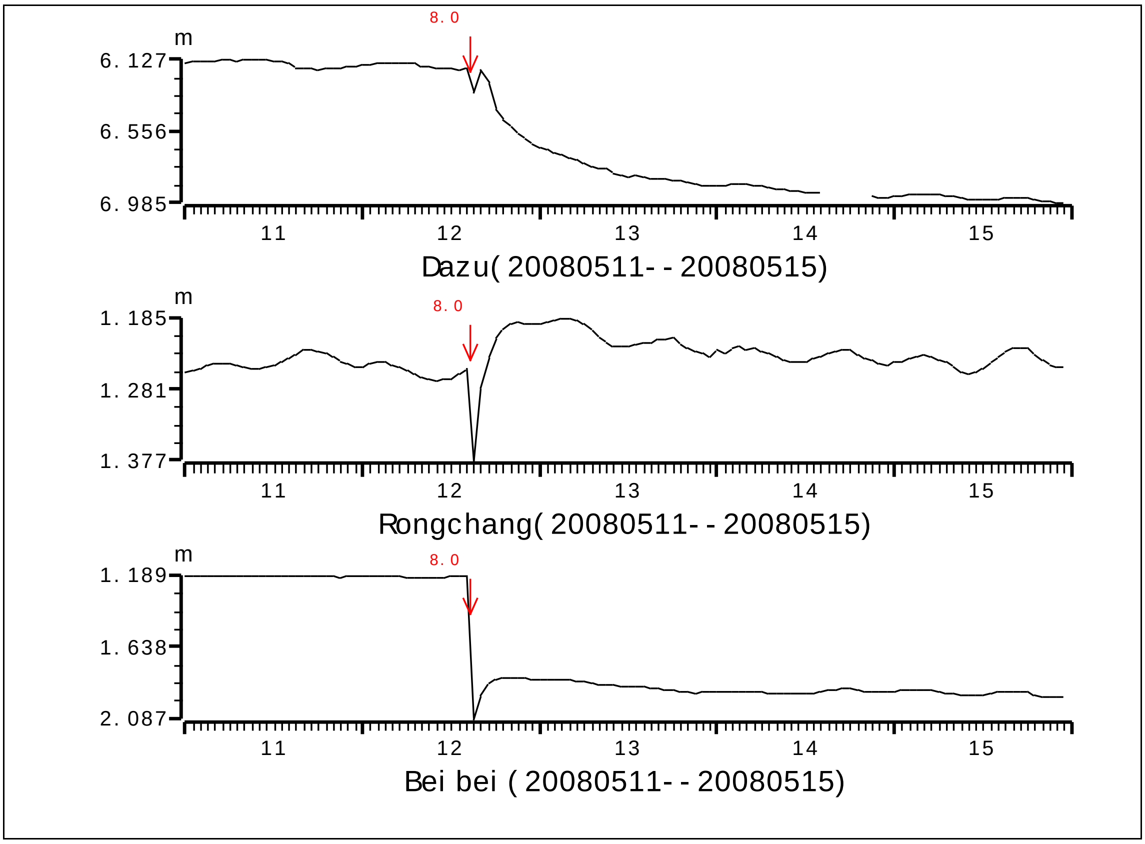
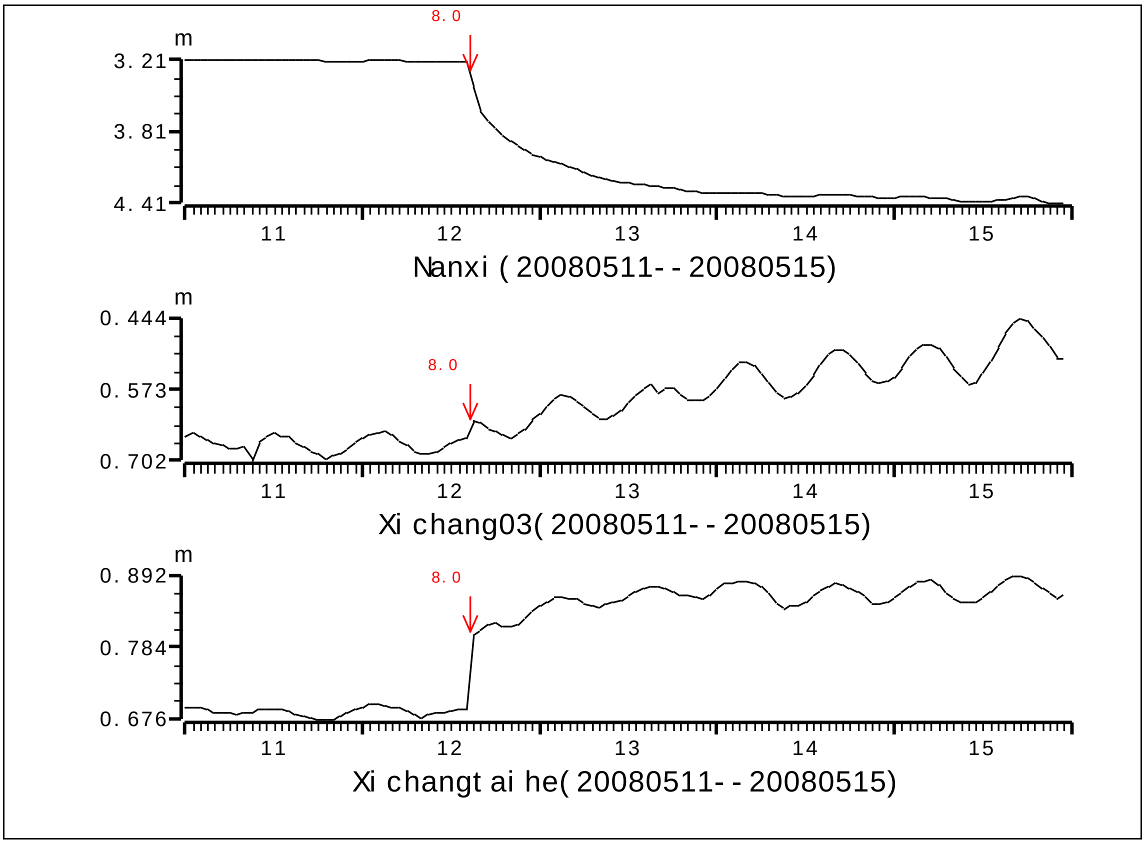
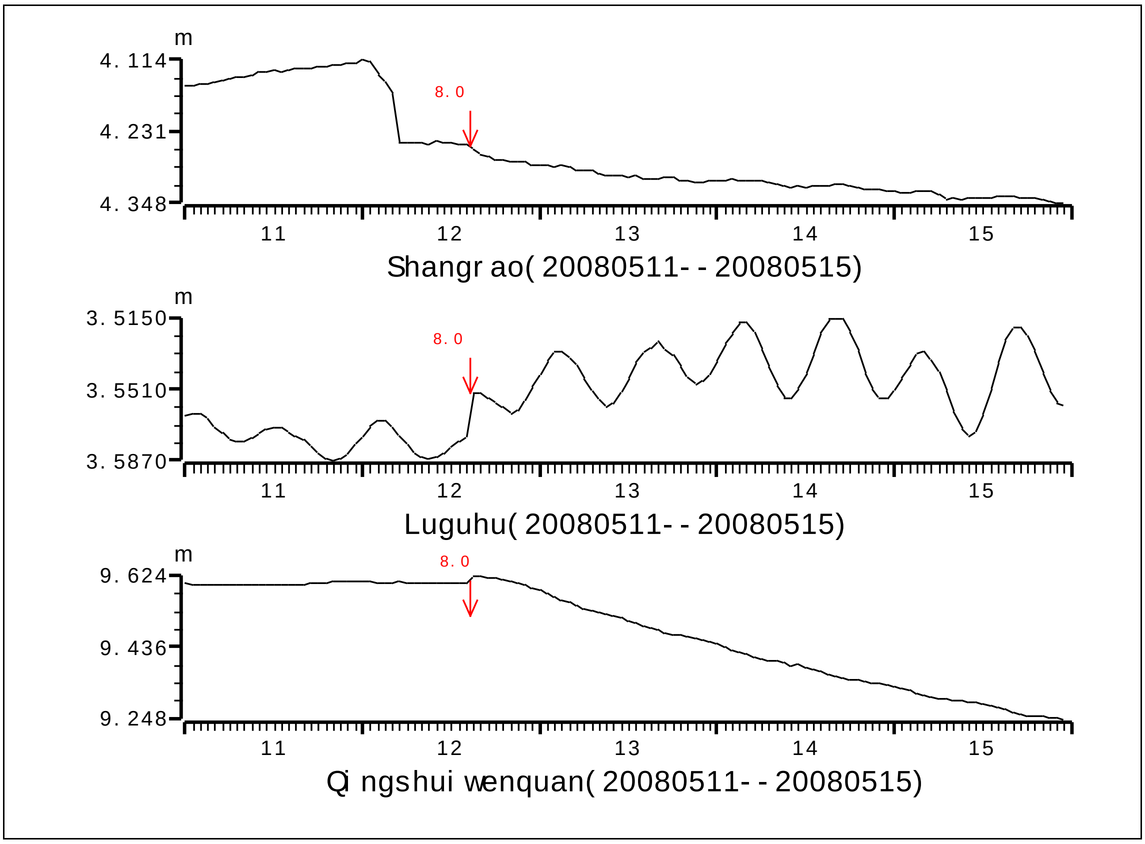


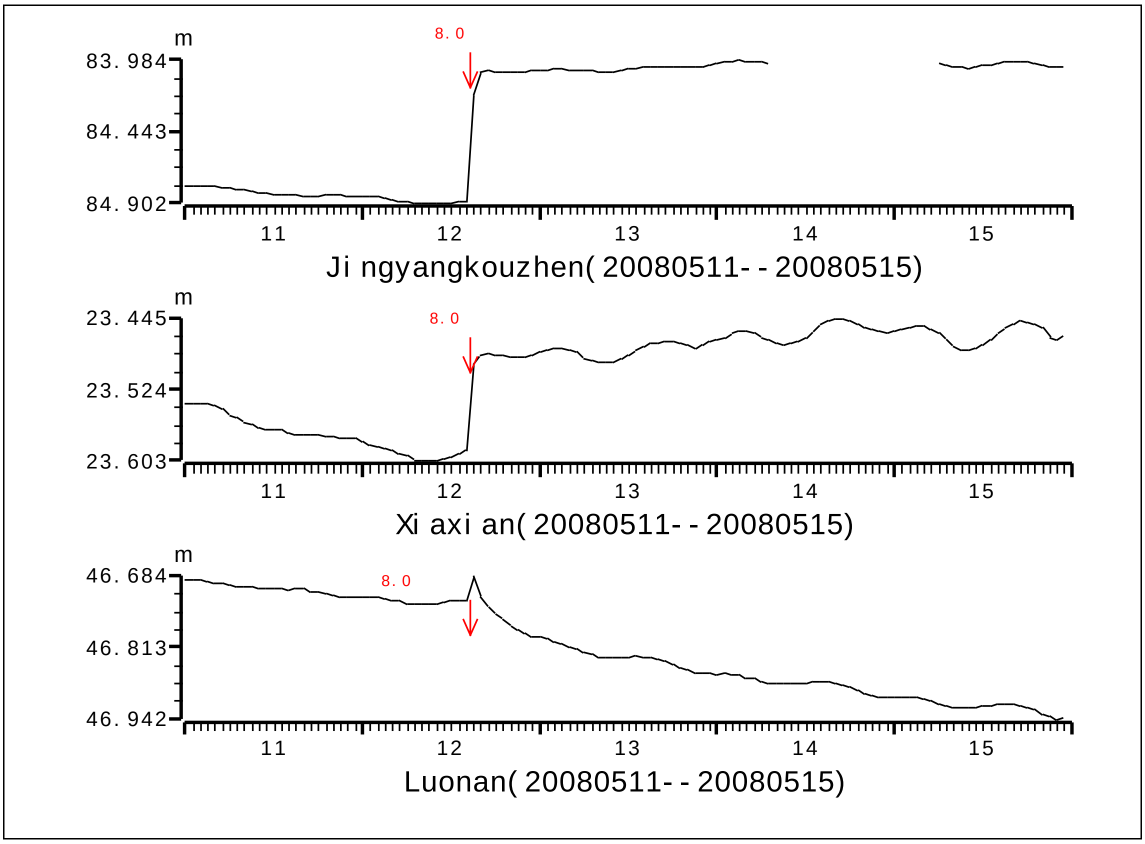
Figure5

Figure 5. Original water level changes of those wells (well: 1-27), which are plotted with hourly data. It is the same with the description of Huang (2008): the sequential number of y-coordinate depends on the type of the well, "sequential number increase from low to high" indicates an artesian well, and the free water surface is higher than the artesian discharge point or the ground, the coordinate value means the height from the free water surface to the artesian discharge point or to the ground. "Sequential number decrease from low to high" indicates a non-artesian well, and the coordinate value means the depth from the free water surface to the ground. All the ascendant patterns in the picture indicate water level ascending, while all those descendent patterns in the picture indicate water level descending.

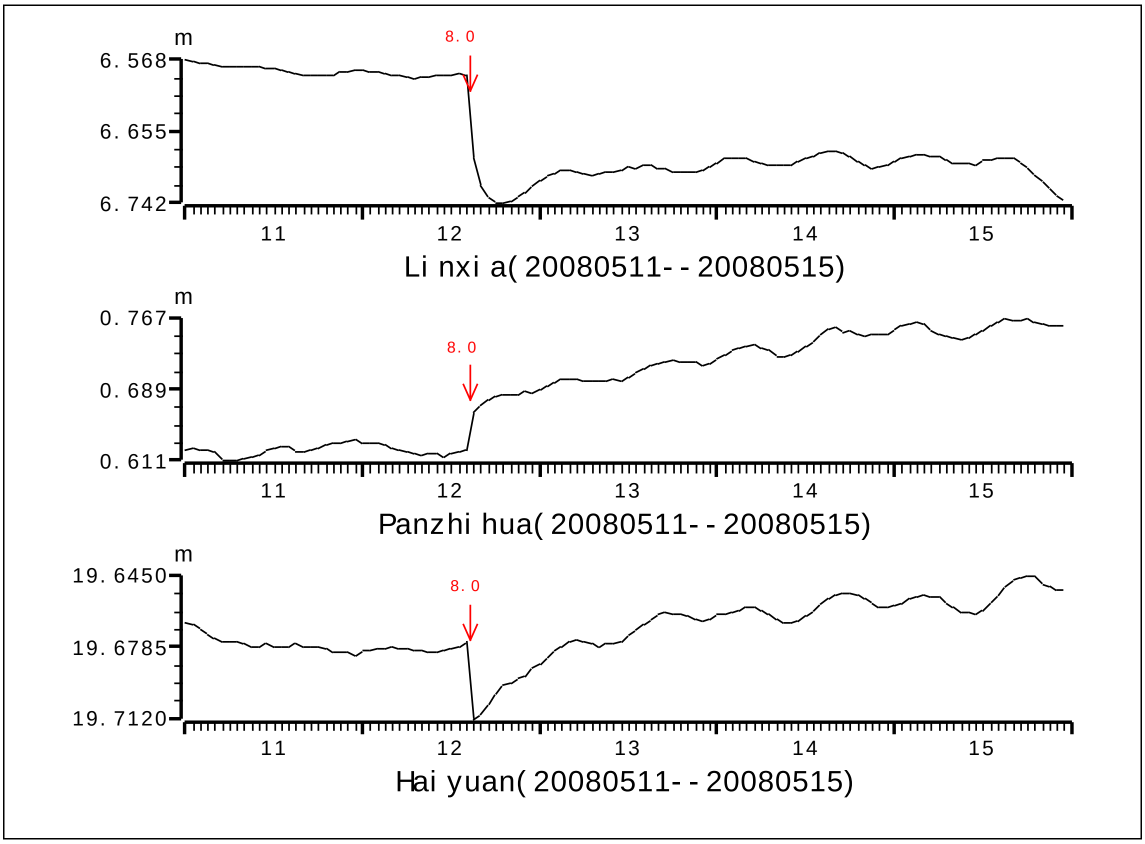



m
3. 21
3. 81
4. 41
8. 0
11
12
13
14
15
Nanxi (20080511- - 20080515)
0. 444
0. 573
0. 702
Xi chang03(20080511- - 20080515)
0. 892
0. 784
0. 676
Xi changt ai he(20080511- - 20080515)

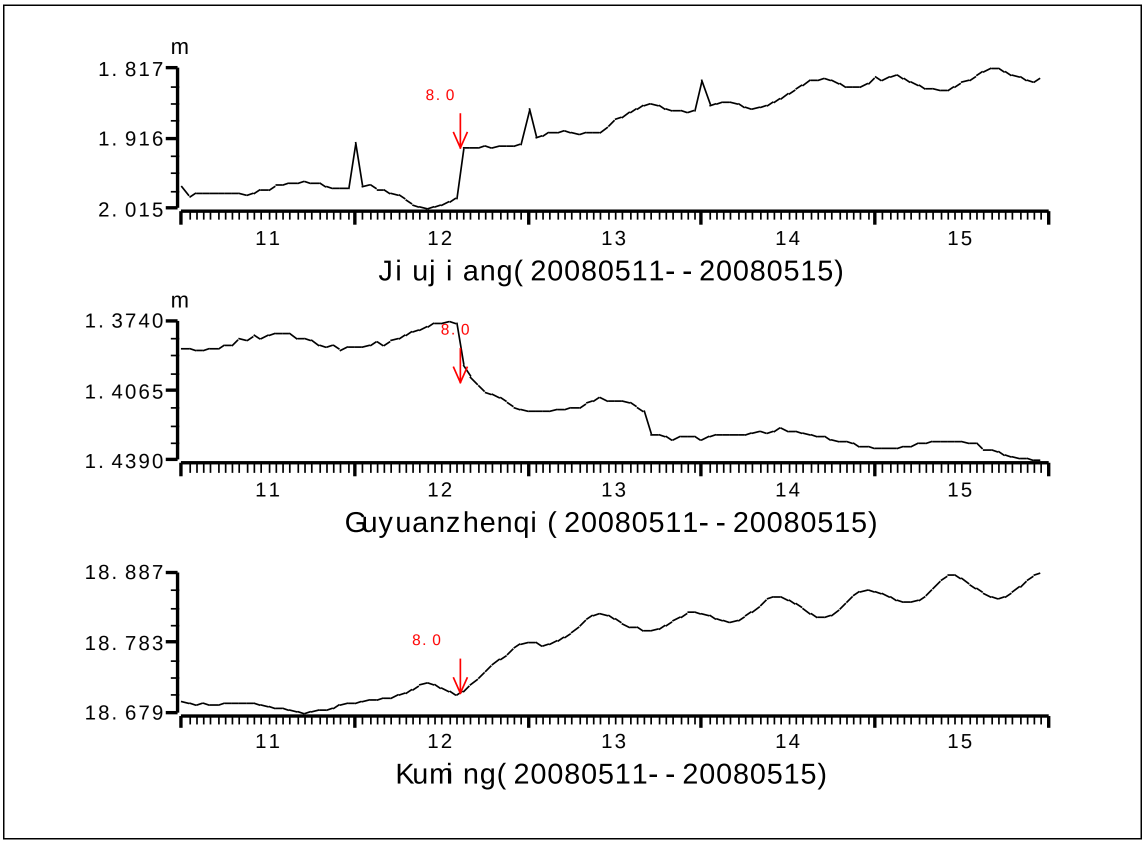

m
4. 114
4. 231
4. 348
8. 0
11
12
13
14
15
Shangr ao(20080511- - 20080515)
3. 5150
3. 5510
3. 5870
Luguhu(20080511- - 20080515)
9. 624
9. 436
9. 248
Qi ngshui wenquan(20080511- - 20080515)

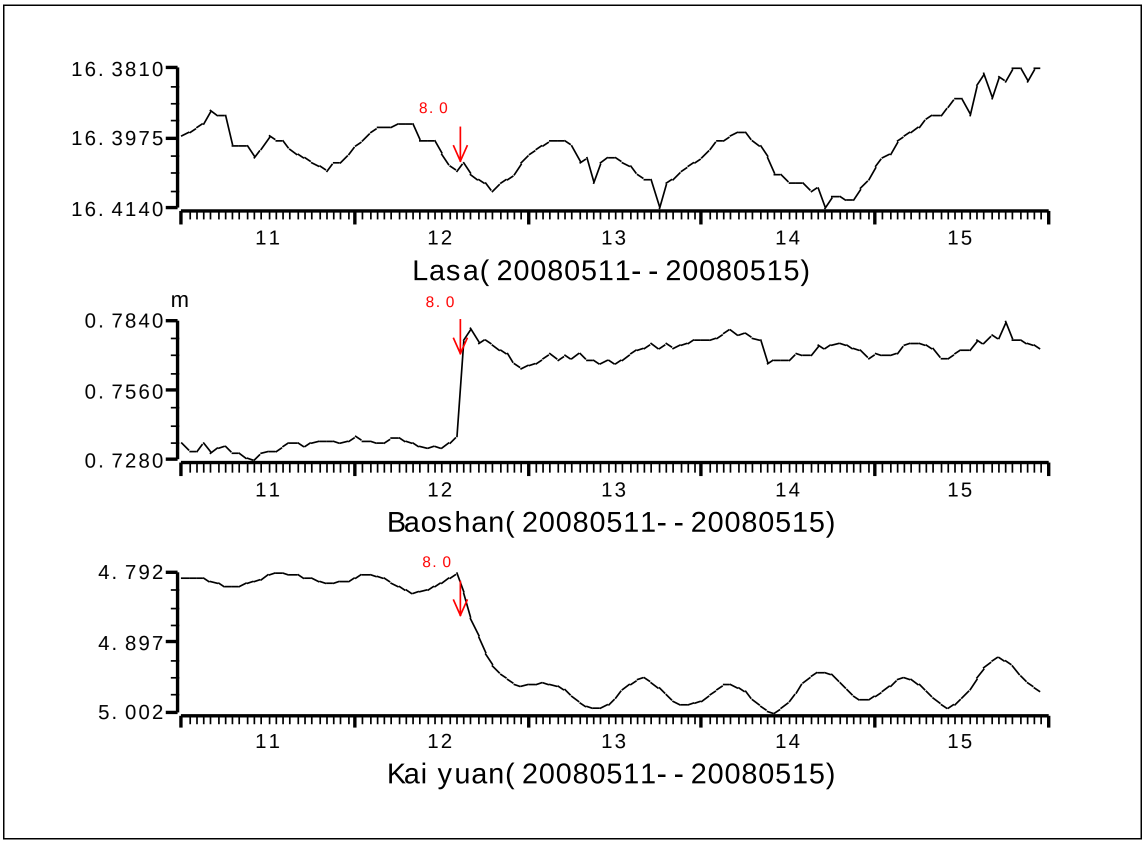


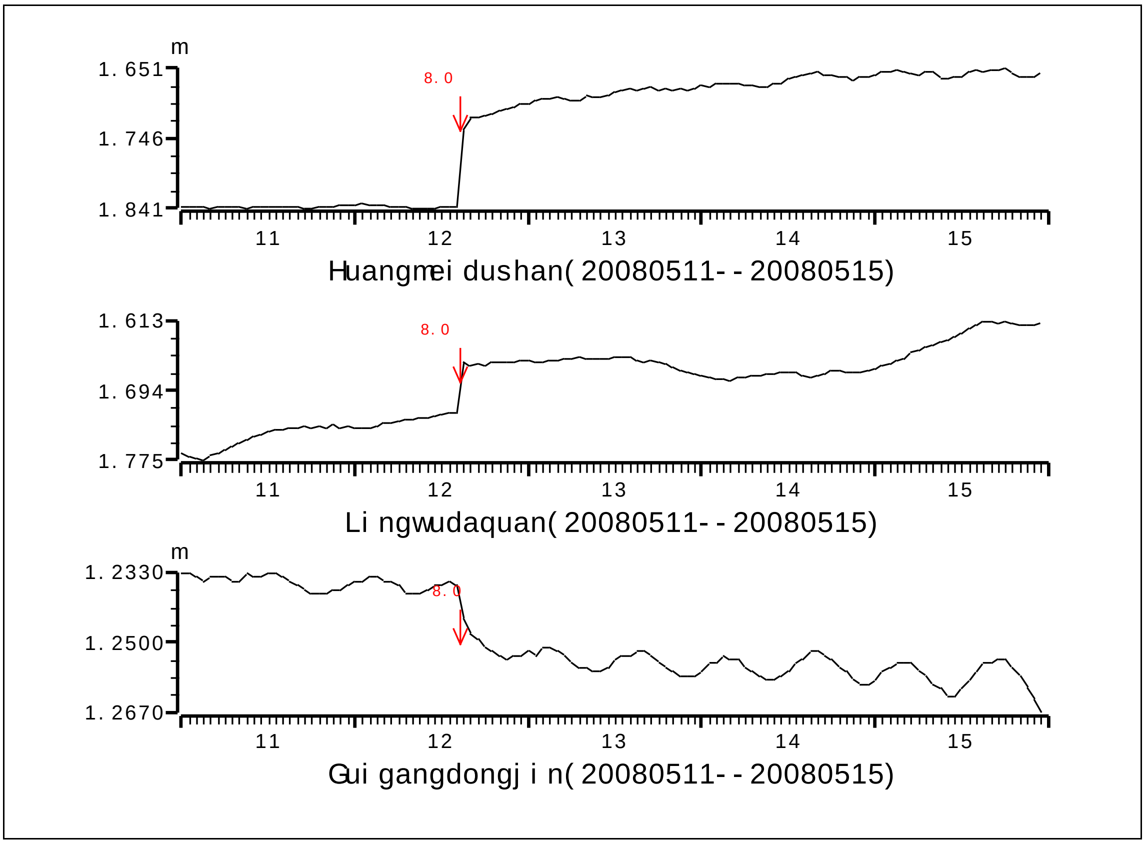
m
83.984
84.443
84.902
8.0
11
12
13
14
15
Jingyangkouzhen(20080511--20080515)
m
23.445
23.524
23.603
8.0
11
12
13
14
15
Xiaxian(20080511--20080515)
m
46.684
46.813
46.942
8.0
11
12
13
14
15
Luonan(20080511--20080515)



m
6.568
6.655
6.742
8.0
11
12
13
14
15
Linxia(20080511--20080515)
m
0.767
0.689
0.611
8.0
11
12
13
14
15
Panzhihua(20080511--20080515)
m
19.6450
19.6785
19.7120
8.0
11
12
13
14
15
Haiyuan(20080511--20080515)


m
1.817
1.916
2.015
8.0
11
12
13
14
15
Jiujiang(20080511--20080515)
m
1.3740
1.4065
1.4390
8.0
11
12
13
14
15
Guyuanzhenqi (20080511--20080515)
18.887
18.783
18.679
8.0
11
12
13
14
15
Kuming(20080511--20080515)


16.3810
16.3975
16.4140
8.0
11
12
13
14
15
Lasa(20080511--20080515)
m
0.7840
0.7560
0.7280
8.0
11
12
13
14
15
Baoshan(20080511--20080515)
4.792
4.897
5.002
8.0
11
12
13
14
15
Kaiyuan(20080511--20080515)


m
1.651
1.746
1.841
8.0
11
12
13
14
15
Huangmei dushan(20080511--20080515)
1.613
1.694
1.775
8.0
Lingwudaquan(20080511--20080515)
m
1.2330
1.2500
1.2670
8.0
Guigangdongjin(20080511--20080515)

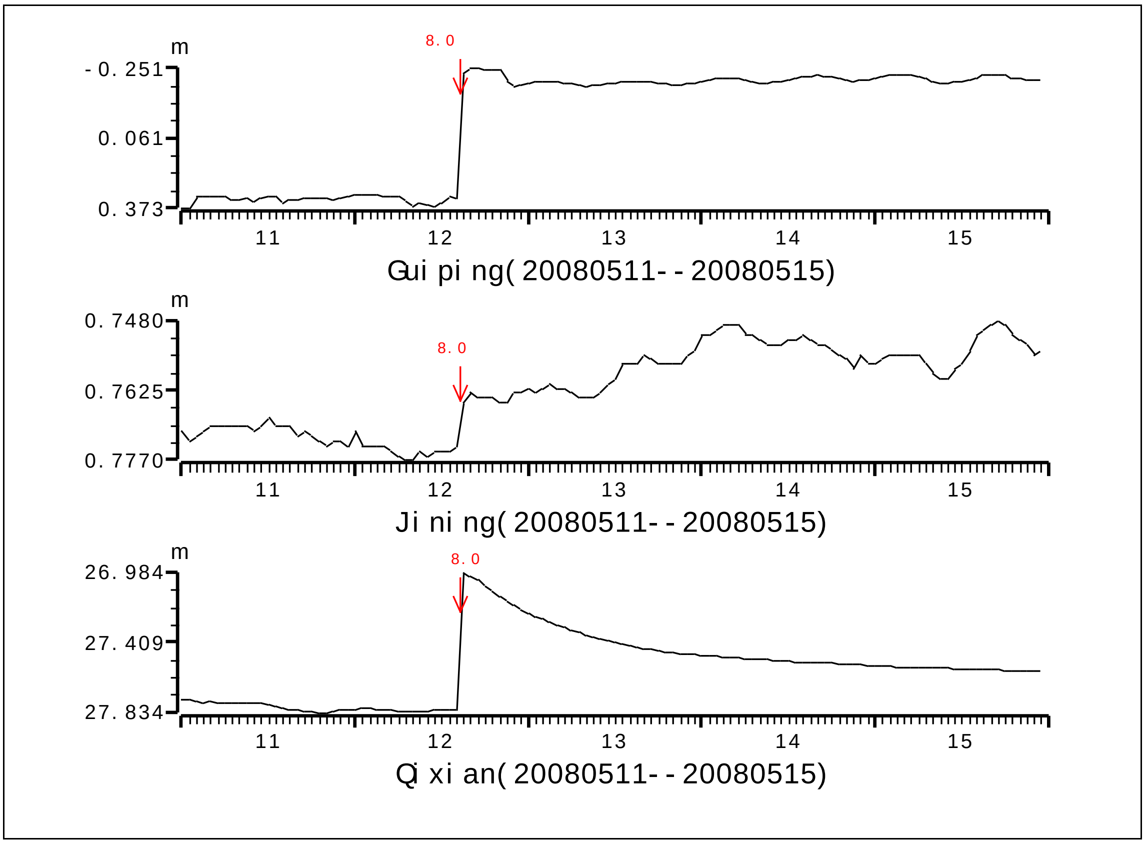
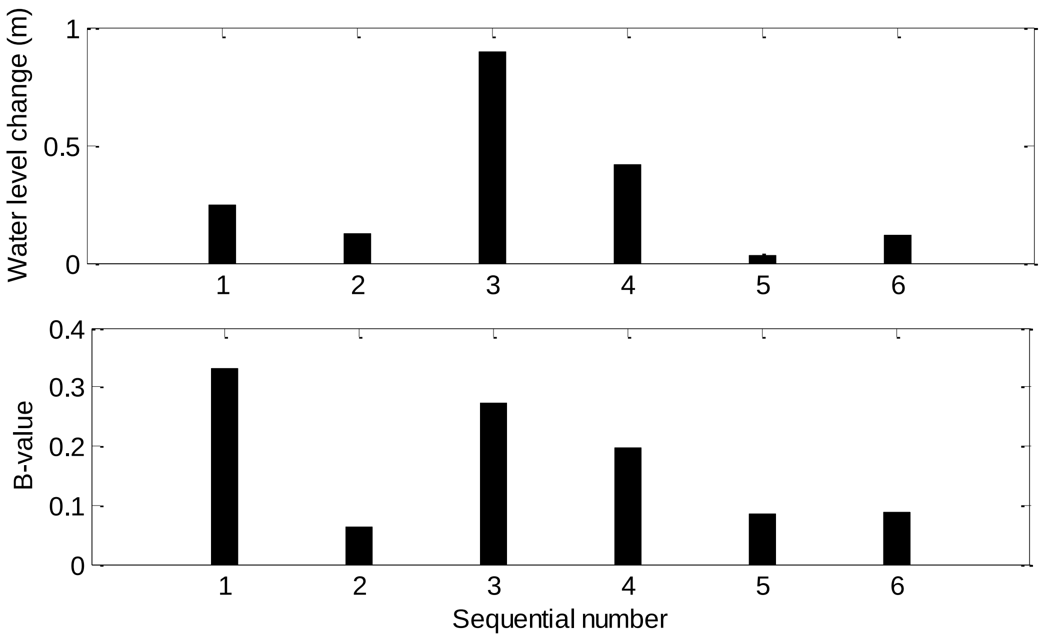

m
-0.251
0.061
0.373
8.0
11
12
13
14
15
Guiping(20080511--20080515)
m
0.7480
0.7625
0.7770
8.0
Jining(20080511--20080515)
m
26.984
27.409
27.834
8.0
Qixian(20080511--20080515)

Figure6

Figure 6. Water level changes and *B* values plotted according to the serial number of those 6 wells. In each group (group a, b, c), large pre-earthquake *B* values come with large co-seismic water level changes.



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