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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 (those co-seismic water level changes are induced by the volumetric changes invoked by un-drained dilatation and consolidation), 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 upload files containing this information during a later step

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Reply:

We have changed a lot in this paper ("Title","Abstract", "Introduction", "Methods", "Mechanism analysis", "discussion" ,"Data and resources" "part of Table 2", "Figure 5" and "Figure 6"), and use the highlighted yellow color to show those changes in the "Response to reviews".

Reviewers' comments:

Reviewer #2: In the submitted manuscript the author made two main conclusions: 1. The waterlevel responses to the 2008 Wenchuan earthquake, as documented in 27 Chinese wells, are consistent with the predictions of the static poroelastic theory. 2. The variation in the magnitude of water-level change at the same epicentral distance is due to the variations in Skempton's coefficient of the wall rocks of the monitory well. The authors' effort to include rock physics data in explaining coseismic water-level changes should be encouraged. Unfortunately, their conclusions appear to be contradicted by the data (Table 2).

1. The waterlevel responses to the 2008 Wenchuan earthquake, as documented by the authors, are inconsistent with the predictions of the static poroelastic theory, as explained below:

In the re-submitted manuscript, the authors compiled rock-physics data for the wall rocks of the monitory wells (Table 1), from which they made 'rough' estimates of the shear modulus (G) for these rocks (Table 2). Using the 'rough' estimates of G and the poroelastic equation (5) they calculate Skempton's coefficient B from the tidal response of the wells (Table 2).

Given the values of Skempton's coefficient for the wall rocks around the wells, it is straightforward to check if the prediction of the poroelastic theory is consistent with the observed water-level changes. Since Skempton's coefficient is the ratio between the undrained change in pore pressure and the applied mean stress, the mean stress that produced the water-level change through the poroelastic effect may be estimated from the quotient between the tabulated water-level changes and values for Skempton's coefficient, which the authors plotted in their Figure 6. The figure shows that this quotient lies between -5 and +5 m, equivalent to an applied mean-stress between -5x104 to +5x104 Pa.

Now this prediction may be compared with the elastic stress produced by the earthquake. Although the authors calculated the elastic stresses using the Okada formula, they did not show the mean stress magnitude. To get an order-of-magnitude estimate, we estimate this from the listed volumetric strain (Table 2) which, for about half of the listed wells, is ~10-9 or less. These volumetric strains correspond to a mean stress of the order of 10 Pa or less, or ~three orders of magnitude smaller than that required by the poroelastic theory to produce the observed water-level changes. Thus, instead of supporting the poroelastic theory in predicting the water-level changes, the authors' result rules out the poroelastic theory as a mechanism for the observed water-level changes.

Reply: this is really a big problem, and we have do large changes in the paper (see the content of “Abstract”, “Introduction”, “Methods”, “Mechanism analysis”, “discussion” , “Table 2”, “Figure 5” and “Figure 6”)

We have changed Figure 5 and Figure 6. In the new Figure 5, we plotted the original water level changes with the hourly data (the figure plotted with the minute data will be obscured in visual sense, especially in the PDF vision), and the amplitude of co-seismic water level changes measured from the hourly figure will be a little different from the minute data. We have modified the amplitude value of co-seismic water level changes in several wells in Table 2 (well: 1, 2, 3, 4, 7, 18), according to the accurate curves. however this little modification will not cause any impact to the result: “large pre-earthquake B value leads to large co-seismic water level changes”

Generally, 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 calculated the co-seismic water level changes based on the poro-elastic theory. 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) through 3 ways: 1) Compare the calculated co-seismic water level change based on the poro-elastic theory with the observed water level change. 2) According to the value of static volume strain changes. 3) According to the pattern of the co-seismic water level changes.

From the analysis above, we can get 13 wells which may fit for the poro-elastic theory in the intermediate field (well: 1, 2, 3, 4, 5, 6, 8, 9, 10, 11, 12, 14, 24) (Table 2). Since that the B values of those 13 wells are valid, meanwhile we use “/” to indicate the invalid B values (Table 2). Among those 10 wells to which the poro-elastic theory can be applied, only 10 can form groups with similar epicentral distances (well: 1, 2, 3, 4, 5, 6, 9, 10, 11, 12). We find that large pre-earthquake B values correspond to large magnitude of co-seismic water level changes, this phenomenon exists in those wells. And then, we use Poro-elastic theory to analyze the mechanism of this phenomenon.

Finally, we concluded: (1)When the water level change of those wells can be explained by the poro-elastic theory, 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, e, f). (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, g, k). (3) When none of those wells with similar epicentral distances can be explained by the poro-elastic theory, and the water level changes are similar in those wells, then we may assume those water level changes may be caused by the transition of the seismic waves (group h : well 15, 16, 17, 18).

Please look at the content of “Abstract”, “Introduction”, “Methods” and “Mechanism analysis” for the details. We have changed a lot in the new version.

2. The claim that Skempton's coefficient is a controlling factor for water-level change is weak, as explained below:

The claim by the author that Skempton's coefficient is a controlling factor for water-level change is based on Figure 5 where the water-level changes at the 27 stations are compared with the calculated Skempton's coefficients. The author's calculated Skempton's coefficient from rock-physics data, however, show large ranges of uncertainty (Table 1). This, in turn, implies that there are large ranges of uncertainty in the estimated Skempton's coefficients. It is likely that, if the large uncertainty were included in Figure 5, the claimed correlation between Skempton's coefficients and water-level change would disappear.

Reply: 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 (Yourong Liu and Huiming 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). There may be ranges of uncertainty in B value getting from the method, but there are no better methods, and scientific workers usually do some reasonable assumption to avoid those complex details in reality, such as Okada’s dislocation model assume the whole land is isotropic and homogeneous, and also use some assumed parameters. Besides, the mean value we use will not cause large degree of inaccuracy to the G value (Table 1).

In addition, a qualitative comparison between the upper and the lower diagrams in Figure 5 is not sufficient to illustrate the existence (or the absence) of a relation between water-level changes and Skempton coefficient. Although the authors stated (p. 9) "Large B values come with large changes in water level. This phenomenon is in accordance with the poro-elastic theory", this is hardly convincing, especially in view that nearly half of the groups (3 out of 7) of stations "do not show any relationship between B-values and water-level changes" (p.10). More quantitative analysis of the data is obviously needed.

Reply: in our last paper the sentence **“nearly half of the groups (3 out of 7) of stations "do not show any relationship between B-values and water-level changes" (p.10)** is used to describe those co-seismic water level changes in the far field, not for the intermediate field, and it is just in the discussion part as a supplement content. So this will not cause any impact on the result of the intermediate field, which we are focused on.

And in our modified version, we just deleted this far field description, so as to avoid any misunderstanding.

There are other relatively minor problems in the manuscript. For example, the authors appeared to be confused about the current models for explaining the earthquake effect on the groundwater level. They tried to explain the observed water-level change with the static poroelastic theory, but, at the same time, argued for their static model using an equation specifically for the dynamic energy associated with seismic waves (p.17).

Reply: We have changed the content in our modified version, and we distinguished the water level changes induced by the poro-elastic theory and those which may be induced by the transition of the seismic waves. Please look at the content of “Mechanism analysis” and “discussion” for details.

Given the above concerns, I cannot recommend the publication of the manuscript in BSSA in its present state.

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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 (those co-seismic water level changes are induced by the volumetric changes invoked by un-drained dilatation and consolidation), 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). At even greater distance (the far field, epicentral distance D larger than 1000 km), only transient oscillations of the water level have been documented. 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 and 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: when the water level change of those wells can be explained by the poro-elastic theory, 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:

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

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=ρ g 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. When an aquifer is not confined, an applied stress can be easily transferred outside the aquifer system without increasing the pore pressure. Thus a low value of *B* indicates a poorly confined aquifer system (Sil, 2006). 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:

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

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) 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 the patterns of those co-seismic water level changes in Figure 5, we take q=100 as the threshold value, when q<100 we suppose the poro-elastic theory can be applied to the well aquifer, otherwise if q>100 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 12 are all smaller than 100, they are much smaller than those q values of well 13 to well 27. The mean q value of those wells (well 1 to 12, discarding well 7) is $\overline{q} = 29.124$, 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 2). 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 $\overline{q} = 29.124$ is relatively acceptable).

Secondly, un-drained dilatation and consolidation of sediments may be responsible for the abrupt water level changes (Wang, 2008). According to Figure 5, the water level in well 1 to 12 (except well 7) show abrupt (step-like) co-seismic changes, which is in accordance with the shape of the co-seismic water level changes caused by the un-drained dilatation and consolidation, and we can use poro-elastic theory to explain those water level changes. While, the co-seismic water level changes in well 13 to 27 are more gradual in general, and that are in conformity with the 'sustained' water level change which is coined by Roeloffs (1998). An earthquake-enhanced permeability may be responsible for the more gradual changes in the intermediate field (Wang, 2008).

Thirdly, the static strain values of 13-27 are obviously smaller than that of well 1-12 (Table 2), the seismic energy density in the relatively far field (D> 500 km) may be too small to initiate undrained consolidation and dilatation, a distinct mechanism is required to explain the sustained water level changes at such distances (Wang, 2008).

From the analysis above, we may just get 13 wells which can fit for the poro-elastic theory in the intermediate field (well: 1, 2, 3, 4, 5, 6, 8, 9, 10, 11, 12, 14, 24) (Table 2). Since that the *B* values of those 13 wells are valid, meanwhile we use "/" to indicate the invalid *B* values (Table 2).

Mechanism analysis

Among those 13 wells to which the poro-elastic theory can be applied, only 10 can form groups with similar epicentral distances (well: 1, 2, 3, 4, 5, 6, 9, 10, 11, 12). We find that large pre-earthquake *B* values correspond to large magnitude of co-seismic water level changes, this phenomenon exists in those 10 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*. Among those 10 wells, wells of group a (1 and 2), group b (3 and 4), group c (5 and 6) stay close with each other, the difference of the epicentral distances is tiny. As show in Figure 3, well 1, 2, 3, 4(group a, b) lie in the same fault, while well 5, 6 (group c) lie in another fault, since that the static stress changes in wells of the same group are similar (As we all know, the static stress changes are not only related to the source parameters, but also related to the parameters (strike, dip, rake) of the receiver fault, when wells are in the same receiver fault, the parameters will be similar, thus the static stress change will be similar). Although well (9, 10) and well (11, 12) not be close to each other in the same group, but the receiver direction and epicentral distance are similar in each group, and the differences of their static strain change is not large as show in Table 2. With the similar static stress changes in wells of group (a, b, c, e, f), the co-seismic water level changes are mainly determined by the

Skempton's coefficient *B*.

What's more, 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 (Table 2). 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, c, e and f. 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 the aquifer is confined (*B*-values are high), the applied stress is mostly transferred into changing pore pressure, which leads to relatively large changes in water level. When an aquifer is unconfined (*B*-values are low), 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 (especially for those wells lie in the same fault with similar static stress changes), but the amplitudes of their co-seismic water level changes are different.

In the other 3 groups (group: d, g, k), the water level changes in 3 wells (well 8, 14, 24) can be explained by the poro-elastic theory, while the other 3 can not (well 7, 13, 25) (Table 2). As show in Table 2, in group (g, k), 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 3, 25) 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 value 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 poroelastic theory generally is not applicable. It can be used to explain those water level changes in group h (well 15, 16, 17, 18), the amplitude of the water level changes in the same group are similar (Table 2), and we can infer those water level changes may be induced by the transfer of seismic waves.

However, it is hard to explain the water level change in the other wells (well 19~27, except for well 24, 25). The obscurity may 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. For well 24 and 25, although the epicentral distances are large, there are just a few faults between the two wells and the epicenter, and the geology condition is more simple than well (19, 20, 21, 22, 23, 25, 26, 27), thus they can fit for the Okada’s premise much better than the others.

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 500 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)When the water level change of those wells can be explained by the poro-elastic theory, 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, e, f). (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, g, k). (3) When none of those wells with similar epicentral distances can be explained by the poro-elastic theory, and the water level changes are similar in those wells, then we may assume those water level changes may be caused by the transition of the seismic waves (group h : well 15, 16, 17, 18).

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). These mechanisms may be useful to explain the water level changes in group (i, j, l), this should 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, 43°; angle of slide,123°; depth, 15km; rupture length, 141km; 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 10 wells. In each group (group a, b, c, e, f), large pre-earthquake *B* values come with large co-seismic water level changes.

modified Table1
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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 $G = \dfrac{E}{2(1+\sigma)}$, 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.

modified Table 2
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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	0.186
11	f	Xiaxian	465.8363	0.106	-3.503	Gneiss	40	0.0339	85.3015	0.034
12	f	Luonan	473.9955	0.07	-6.082	Limestone	20	0.0296	75.9071	0.03
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	Lingwudaqı	856.022	0.053	-2.723	Sandstone	8	0.0605	145.964	/
24	k	Guigangdor	899.981	-0.014	1.943	Sandstone	8	0.0722	45.2783	0.072
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	/

new Figure5

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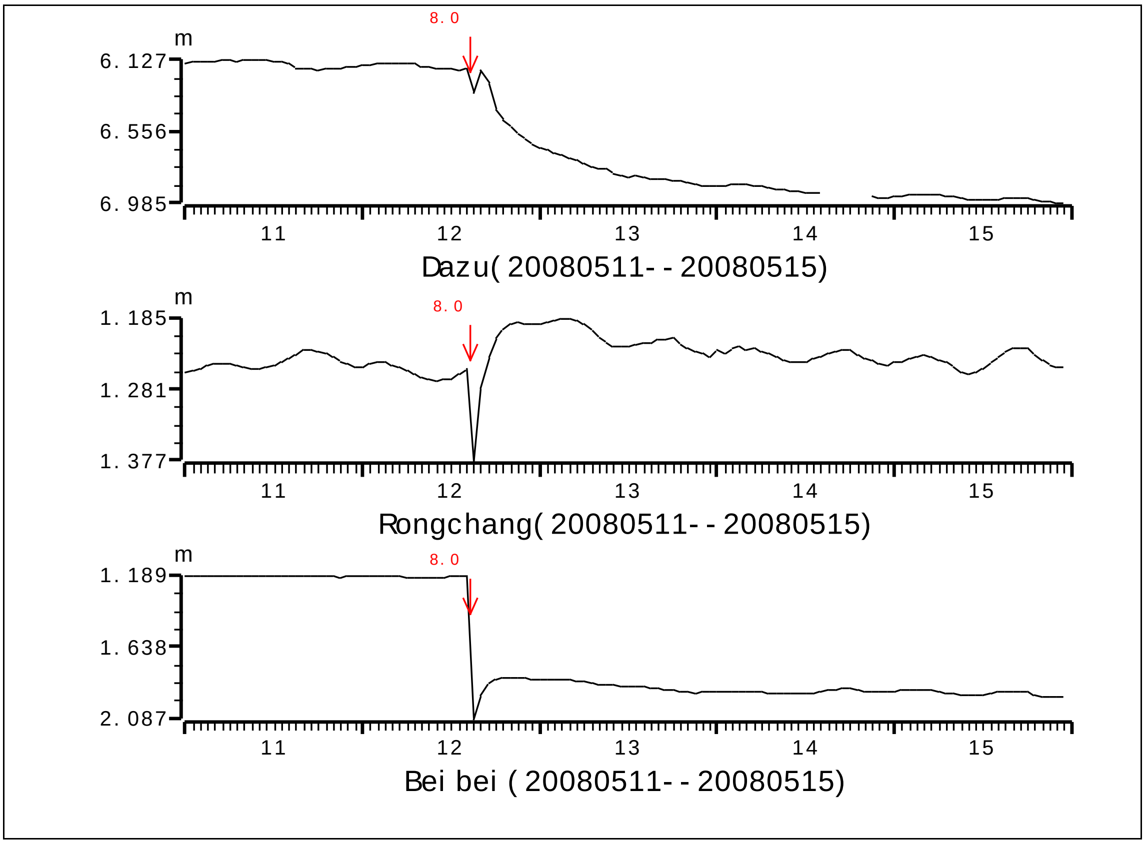
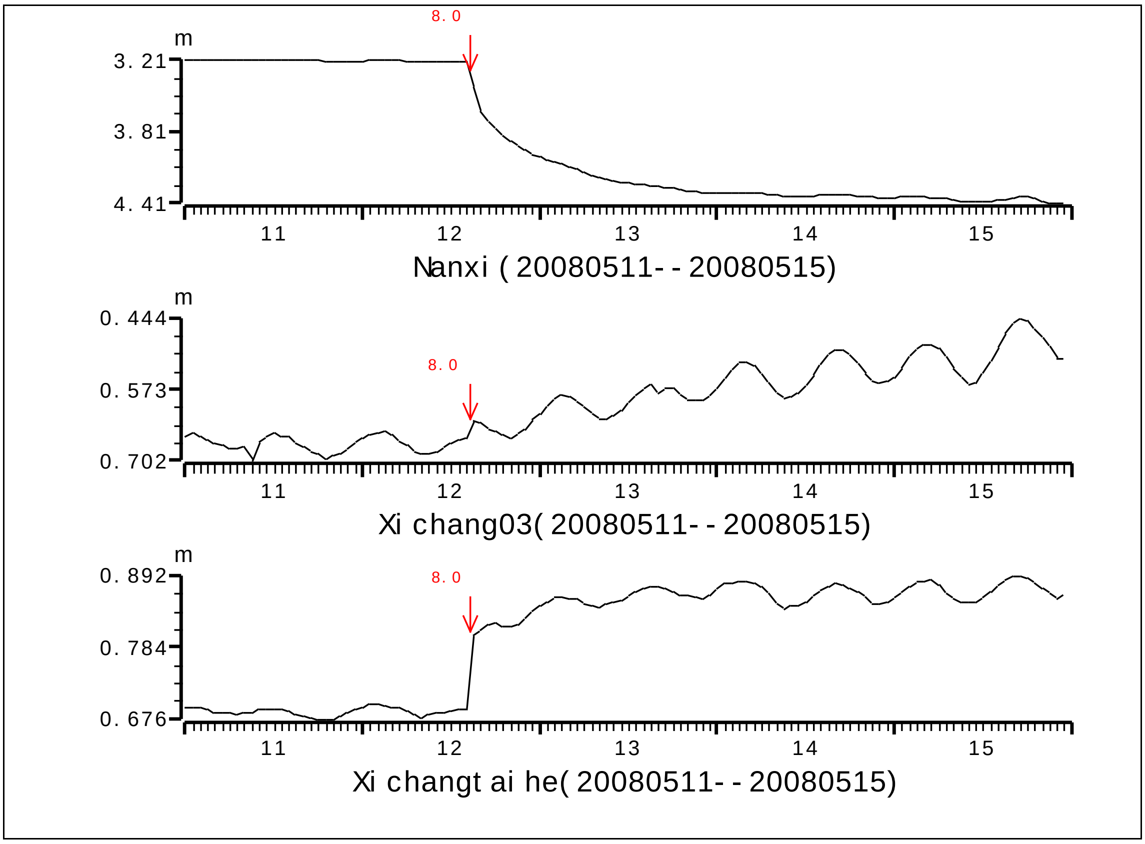
We just added the new content (Original water level changes of those wells**) in Figure 5, show as below:**

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.

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13
14
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0.702
8.0
11
12
13
14
15
Xichang03(20080511--20080515)
m
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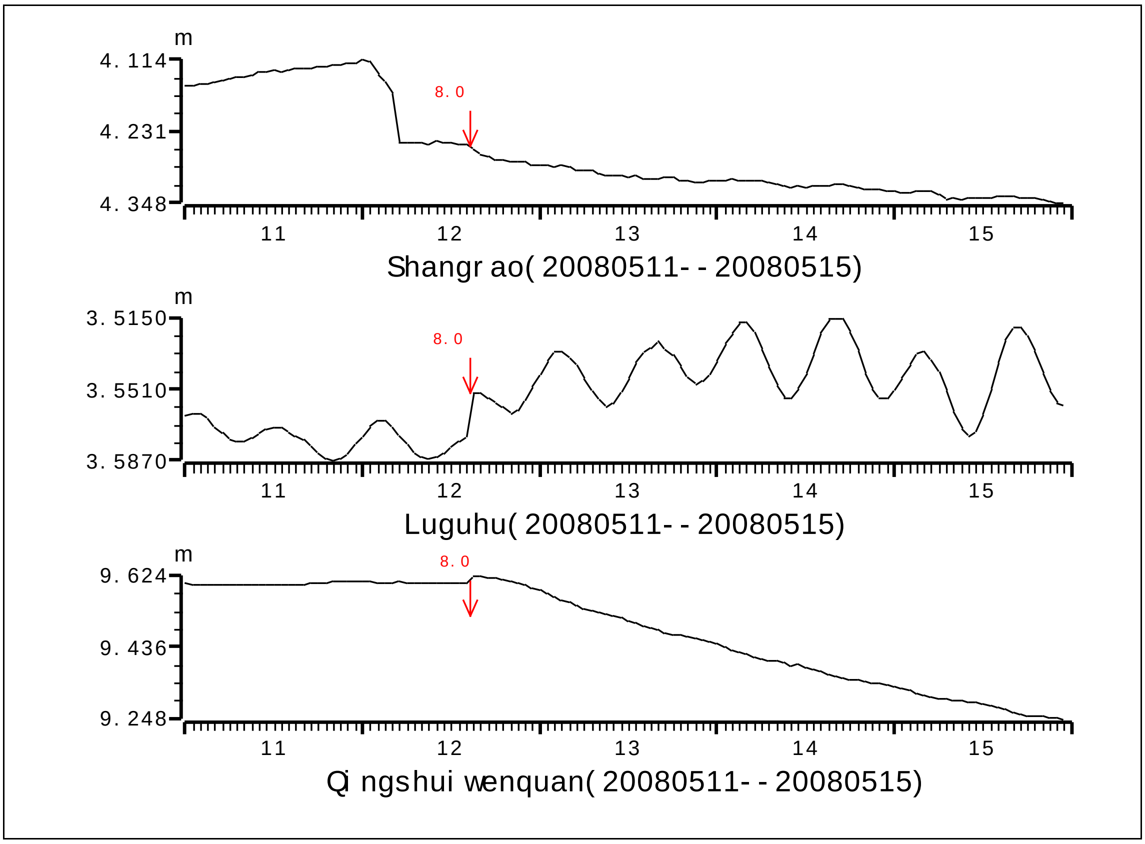
0.784
0.676
8.0
11
12
13
14
15
Xichangtaihe(20080511--20080515)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4.348
8.0
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13
14
15
Shangrao(20080511--20080515)
m
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3.5510
3.5870
8.0
11
12
13
14
15
Luguhu(20080511--20080515)
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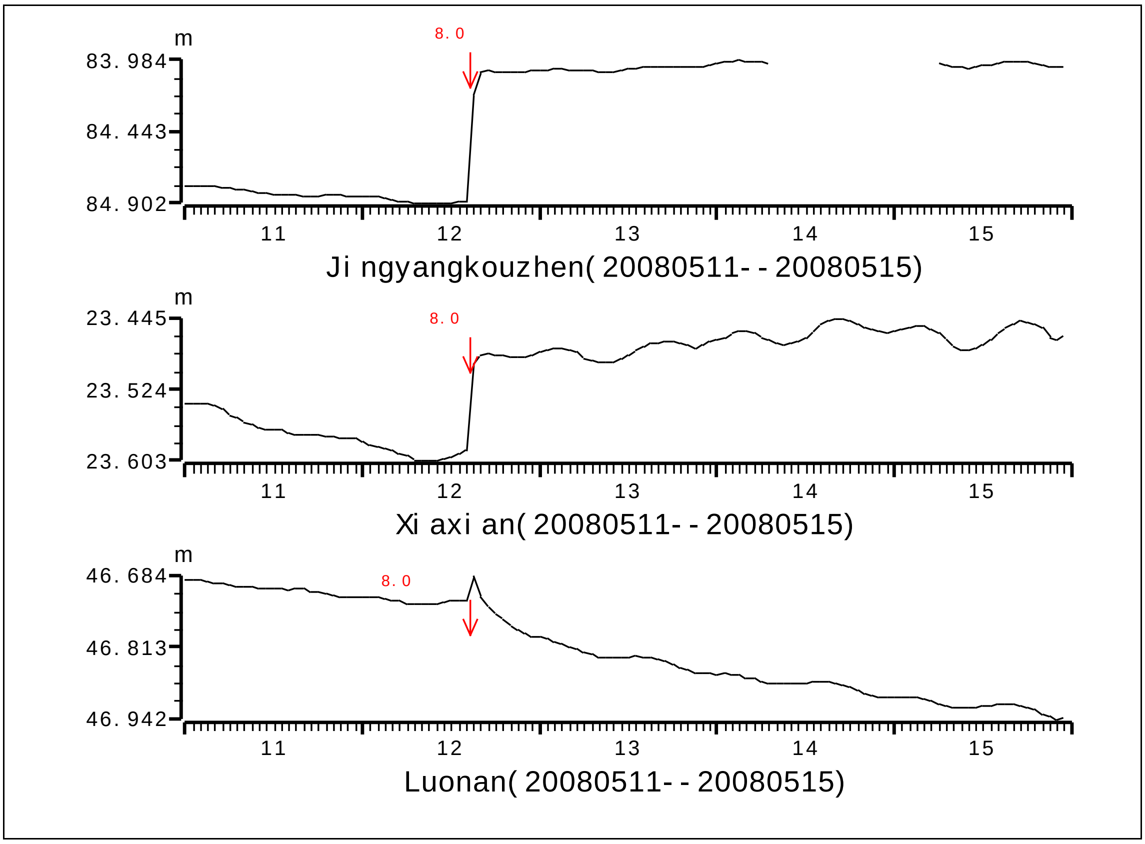
Qingshui wenquan(20080511--20080515)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12
13
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23.524
23.603
8.0
11
12
13
14
15
Xiaxian(20080511--20080515)
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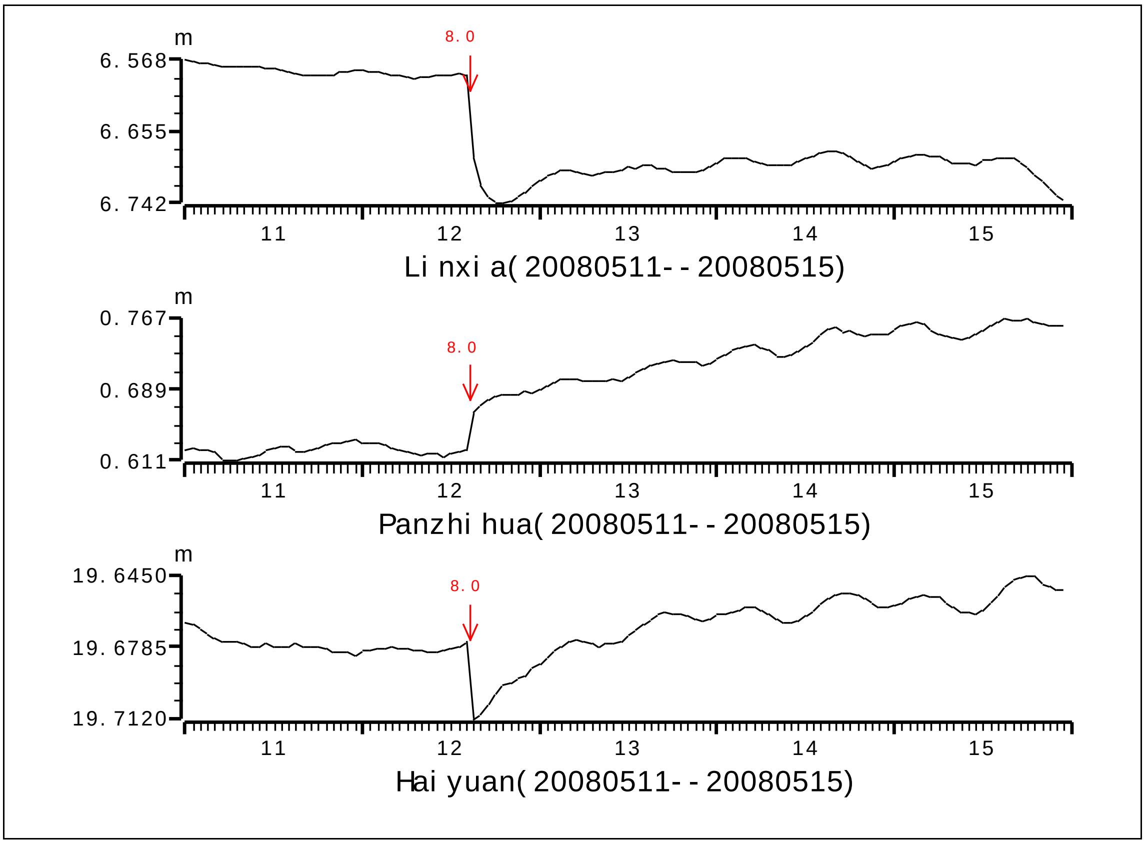
Luonan(20080511--20080515)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13
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0.611
8.0
11
12
13
14
15
Panzhihua(20080511--20080515)
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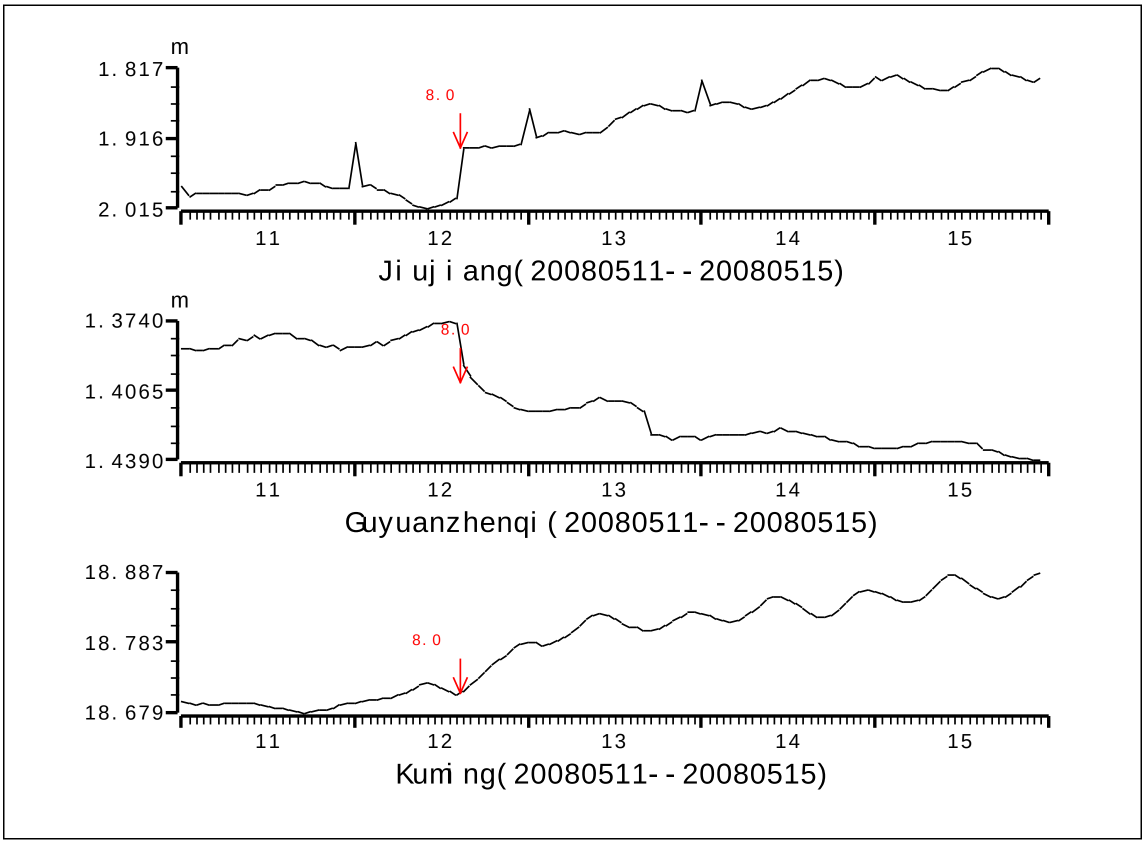
15
Haiyuan(20080511--20080515)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13
14
15
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m
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1.4390
8.0
11
12
13
14
15
Guyuanzhenqi (20080511--20080515)
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12
13
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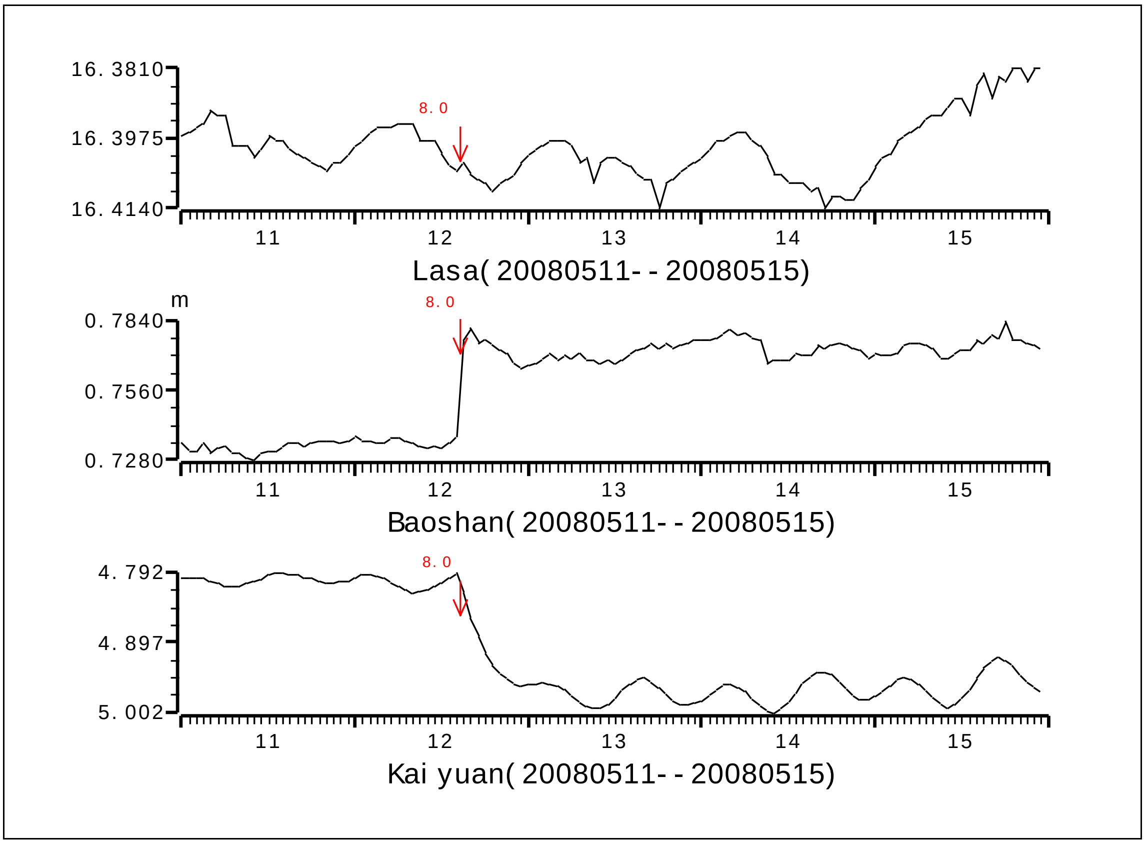
15
Kuming(20080511--20080515)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13
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15
Baoshan(20080511--20080515)
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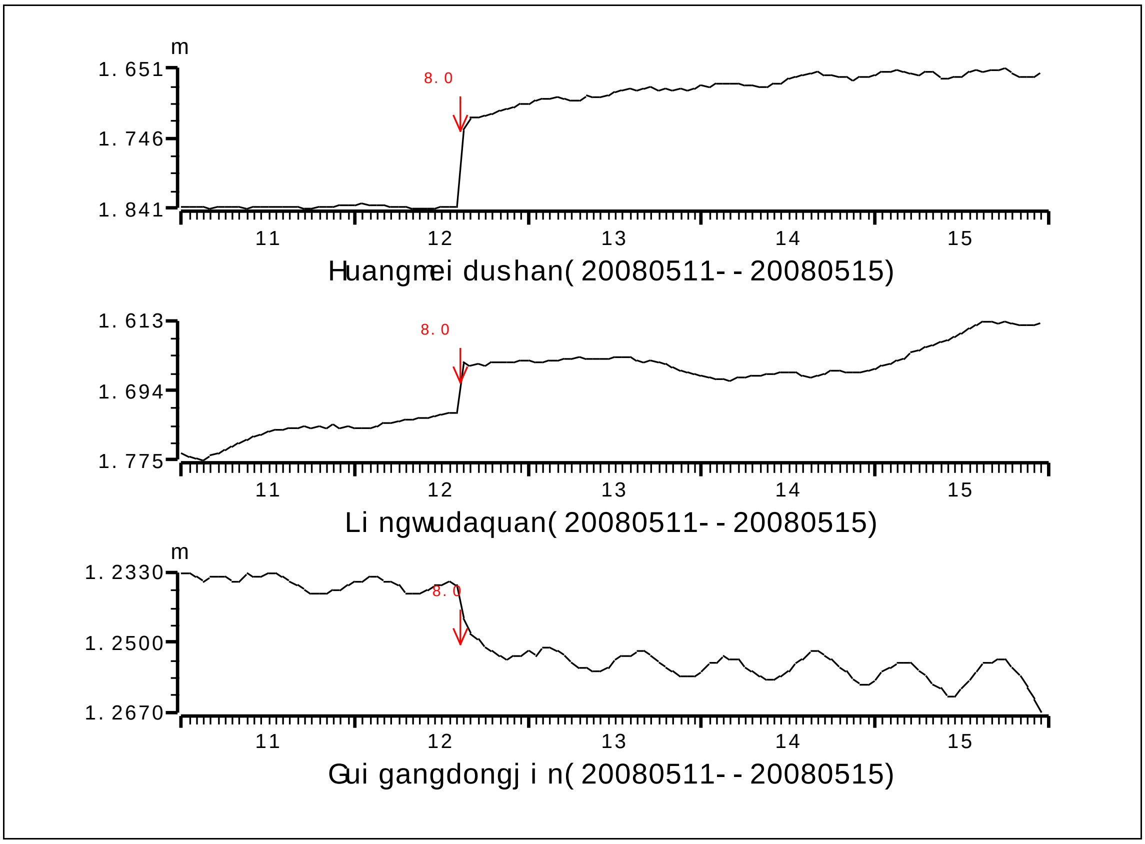
Kaiyuan(20080511--20080515)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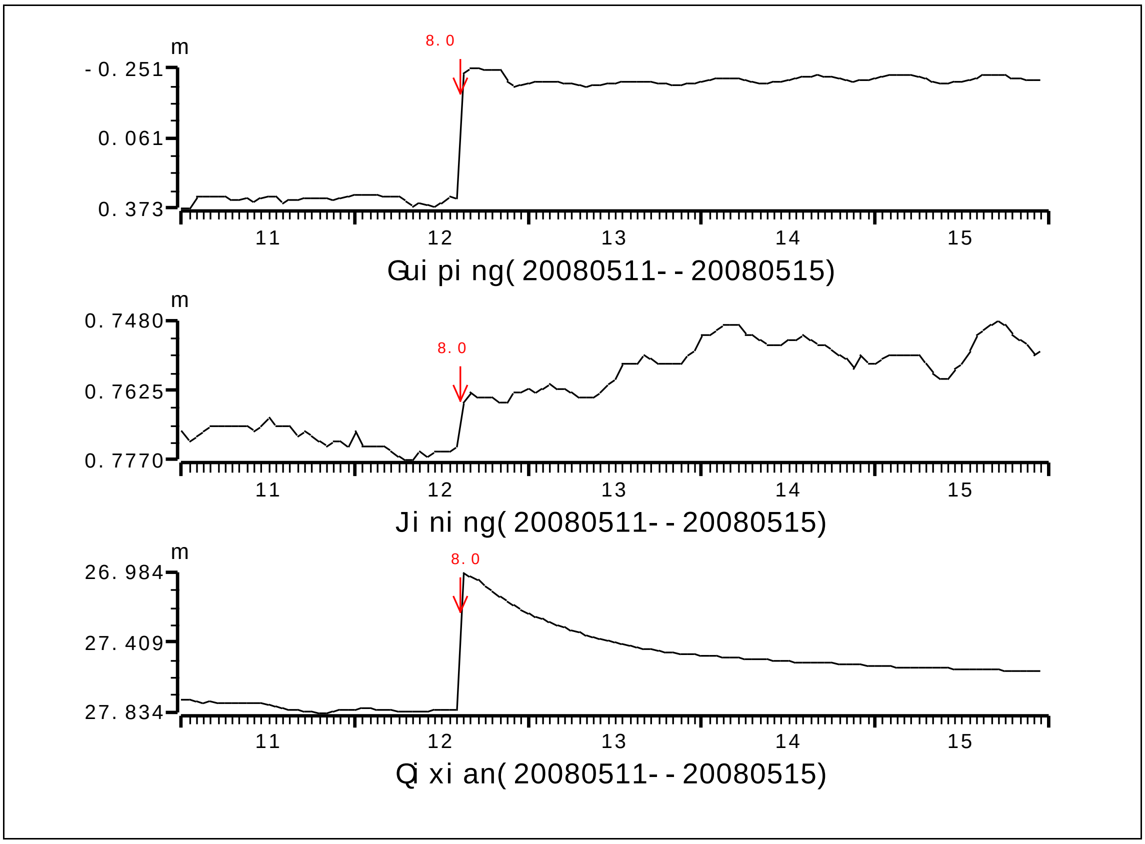
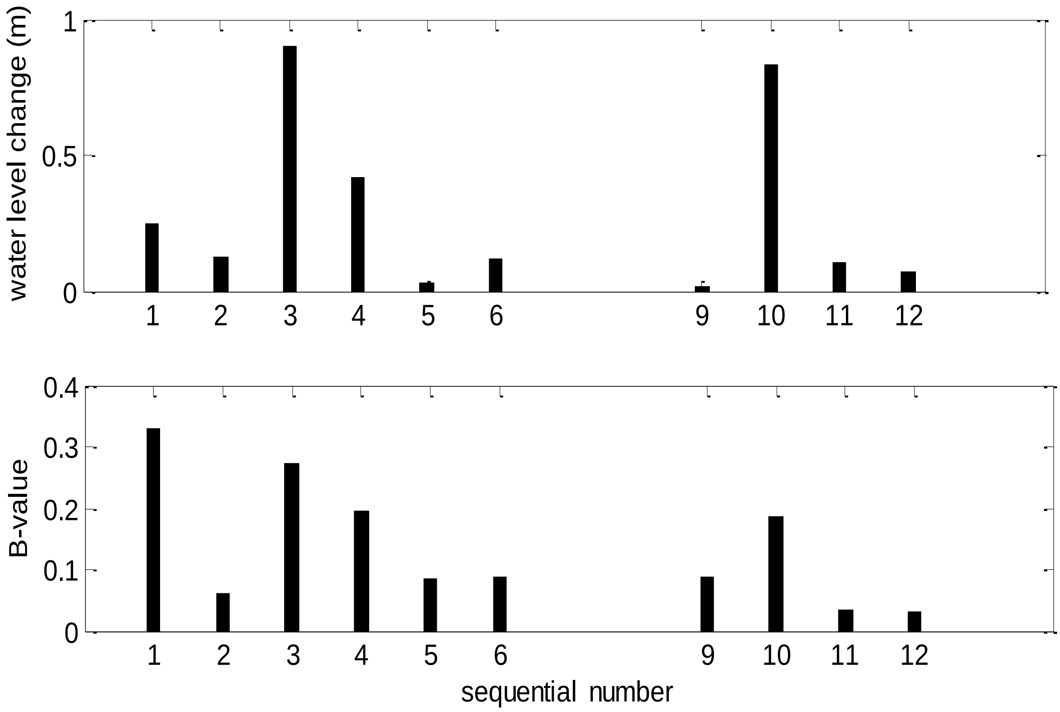
1.775
8.0
11
12
13
14
15
Lingwudaquan(20080511--20080515)
m
1.2330
1.2500
1.2670
8.0
11
12
13
14
15
Guigangdongjin(20080511--20080515)


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
11
12
13
14
15
Jining(20080511--20080515)
m
26.984
27.409
27.834
8.0
11
12
13
14
15
Qixian(20080511--20080515)

We just change the old Figure 5 into Figure 6, show as below:

Figure 6. Water level changes and *B* values plotted according to the serial number of those 10 wells. In each group (group a, b, c, e, f), large pre-earthquake *B* values come with large co-seismic water level changes.



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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 (those co-seismic water level changes are induced by the volumetric changes invoked by un-drained dilatation and consolidation), 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). At even greater distance (the far field, epicentral distance D larger than 1000 km), only transient oscillations of the water level have been documented. 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 and 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: when the water level change of those wells can be explained by the poro-elastic theory, 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=ρ g 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. When an aquifer is not confined, an applied stress can be easily transferred outside the aquifer system without increasing the pore pressure. Thus a low value of *B* indicates a poorly confined aquifer system (Sil, 2006). 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:

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

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) 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 the patterns of those co-seismic water level changes in Figure 5, we take q=100 as the threshold value, when q<100 we suppose the poro-elastic theory can be applied to the well aquifer, otherwise if q>100 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 12 are all smaller than 100, they are much smaller than those q values of well 13 to well 27. The mean q value of those wells (well 1 to 12, discarding well 7) is $\overline{q} = 29.124$, 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 2). 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 $\overline{q} = 29.124$ is relatively acceptable).

Secondly, un-drained dilatation and consolidation of sediments may be responsible for the abrupt water level changes (Wang, 2008). According to Figure 5, the water level in well 1 to 12 (except well 7) show abrupt (step-like) co-seismic changes, which is in accordance with the shape of the co-seismic water level changes caused by the un-drained dilatation and consolidation, and we can use poro-elastic theory to explain those water level changes. While, the co-seismic water level changes in well 13 to 27 are more gradual in general, and that are in conformity with the 'sustained' water level change which is coined by Roeloffs (1998). An earthquake-enhanced permeability may be responsible for the more gradual changes in the intermediate field (Wang, 2008).

Thirdly, the static strain values of 13-27 are obviously smaller than that of well 1-12 (Table 2), the seismic energy density in the relatively far field (D> 500 km) may be too small to initiate undrained consolidation and dilatation, a distinct mechanism is required to explain the sustained water level changes at such distances (Wang, 2008).

From the analysis above, we may just get 13 wells which can fit for the poro-elastic theory in the intermediate field (well: 1, 2, 3, 4, 5, 6, 8, 9, 10, 11, 12, 14, 24) (Table 2). Since that the *B* values of those 13 wells are valid, meanwhile we use "/" to indicate the invalid *B* values (Table 2).

Mechanism analysis

Among those 13 wells to which the poro-elastic theory can be applied, only 10 can form groups with similar epicentral distances (well: 1, 2, 3, 4, 5, 6, 9, 10, 11, 12). We find that large pre-earthquake *B* values correspond to large magnitude of co-seismic water level changes, this phenomenon exists in those 10 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*. Among those 10 wells, wells of group a (1 and 2), group b (3 and 4), group c (5 and 6) stay close with each other, the difference of the epicentral distances is tiny. As show in Figure 3, well 1, 2, 3, 4(group a, b) lie in the same fault, while well 5, 6 (group c) lie in another fault, since that the static stress changes in wells of the same group are similar (As we all know, the static stress changes are not only related to the source parameters, but also related to the parameters (strike, dip, rake) of the receiver fault, when wells are in the same receiver fault, the parameters will be similar, thus the static stress change will be similar). Although well (9, 10) and well (11, 12) not be close to each other in the same group, but the receiver direction and epicentral distance are similar in each group, and the differences of their static strain change is not large as show in Table 2. With the similar static stress changes in wells of group (a, b, c, e, f), the co-seismic water level changes are mainly determined by the

Skempton's coefficient *B*.

What's more, 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 (Table 2). 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, c, e and f. 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 the aquifer is confined (*B*-values are high), the applied stress is mostly transferred into changing pore pressure, which leads to relatively large changes in water level. When an aquifer is unconfined (*B*-values are low), 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 (especially for those wells lie in the same fault with similar static stress changes), but the amplitudes of their co-seismic water level changes are different.

In the other 3 groups (group: d, g, k), the water level changes in 3 wells (well 8, 14, 24) can be explained by the poro-elastic theory, while the other 3 can not (well 7, 13, 25) (Table 2). As show in Table 2, in group (g, k), 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 3, 25) 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 value 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 poroelastic theory generally is not applicable. It can be used to explain those water level changes in group h (well 15, 16, 17, 18), the amplitude of the water level changes in the same group are similar (Table 2), and we can infer those water level changes may be induced by the transfer of seismic waves.

However, it is hard to explain the water level change in the other wells (well 19~27, except for well 24, 25). The obscurity may 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. For well 24 and 25, although the epicentral distances are large, there are just a few faults between the two wells and the epicenter, and the geology condition is more simple than well (19, 20, 21, 22, 23, 25, 26, 27), thus they can fit for the Okada's premise much better than the others.

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 500 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)When the water level change of those wells can be explained by the poro-elastic theory, 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, e, f). (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, g, k). (3) When none of those wells with similar epicentral distances can be explained by the poro-elastic theory, and the water level changes are similar in those wells, then we may assume those water level changes may be caused by the transition of the seismic waves (group h : well 15, 16, 17, 18).

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). These mechanisms may be useful to explain the water level changes in group (i, j, l), this should 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.

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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

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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.

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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	0.186
11	f	Xiaxian	465.8363	0.106	-3.503	Gneiss	40	0.0339	85.3015	0.034
12	f	Luonan	473.9955	0.07	-6.082	Limestone	20	0.0296	75.9071	0.03
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	Lingwudaq	856.022	0.053	-2.723	Sandstone	8	0.0605	145.964	/
24	k	Guigangdor	899.981	-0.014	1.943	Sandstone	8	0.0722	45.2783	0.072
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).

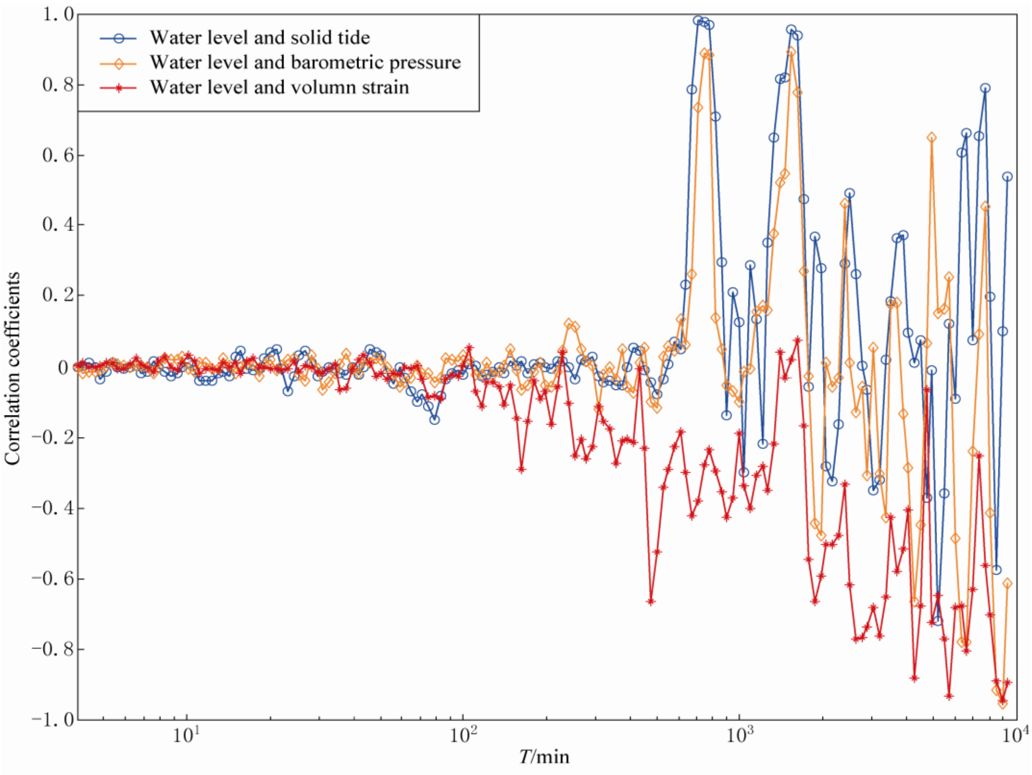

Figure2

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).

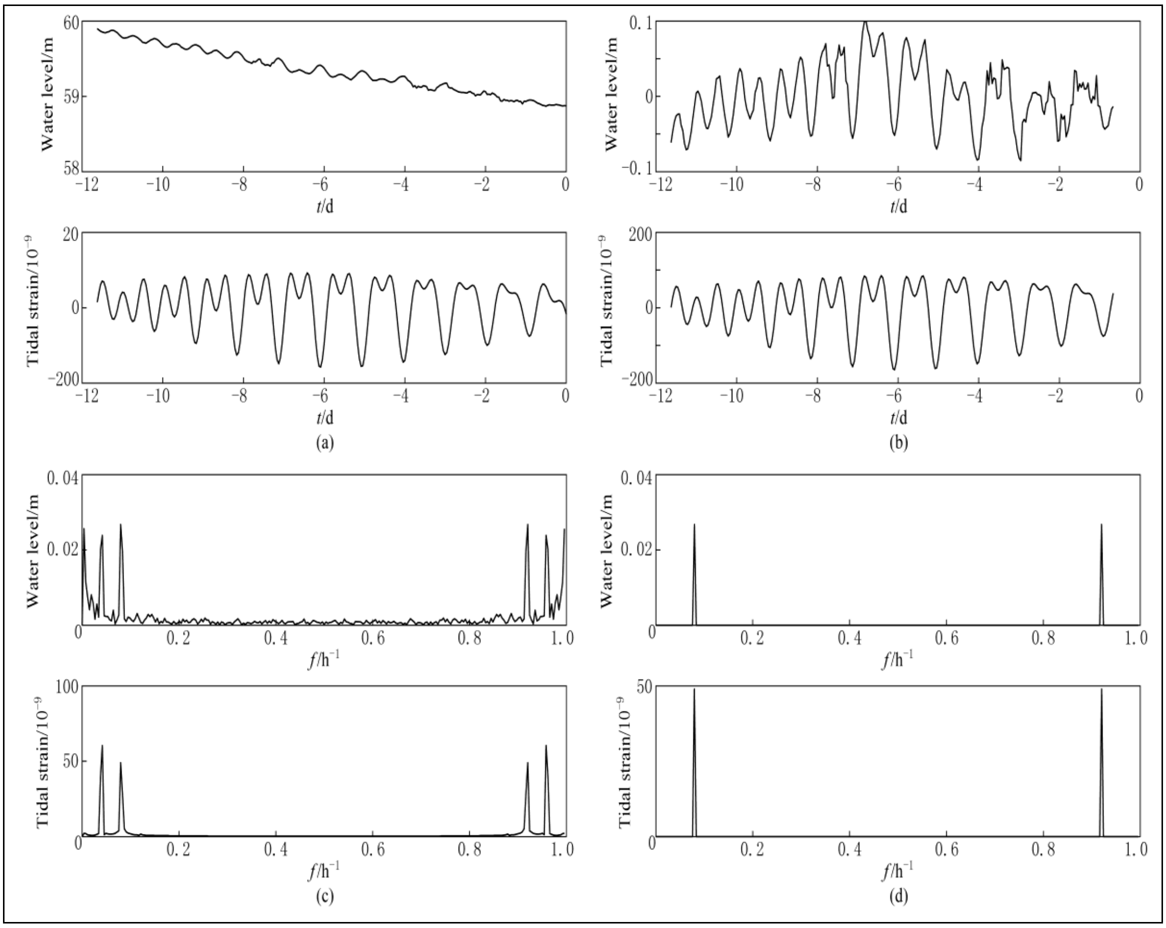

Figure3

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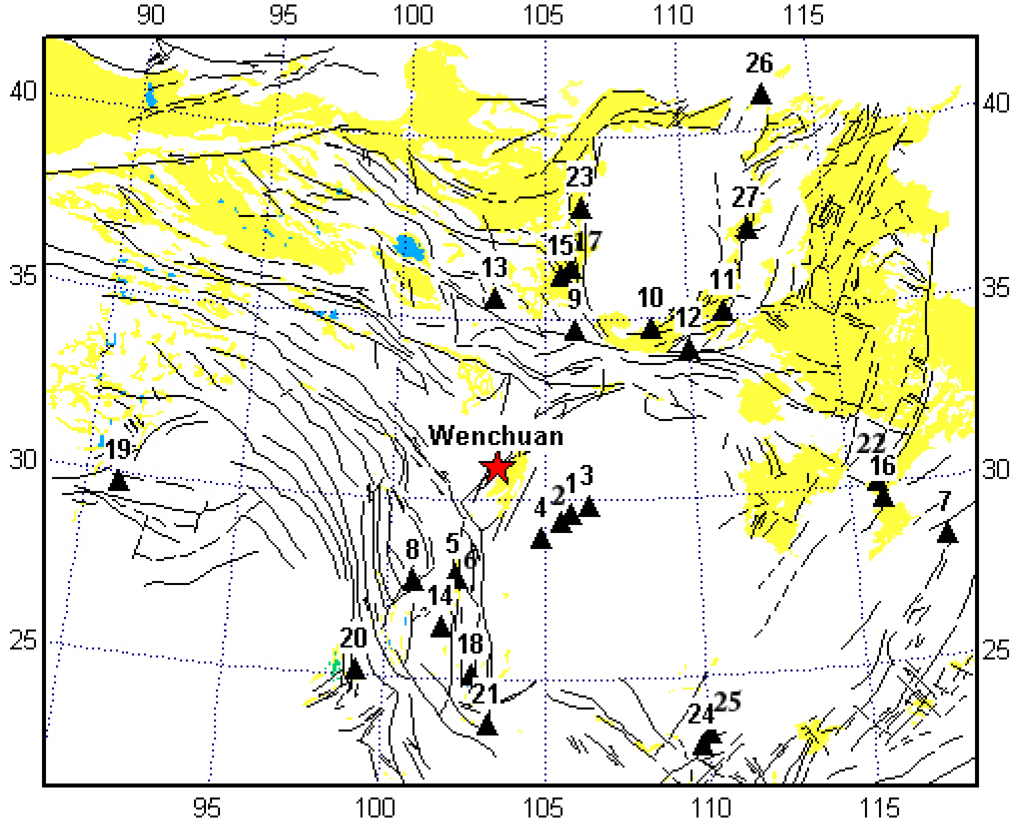

Figure4

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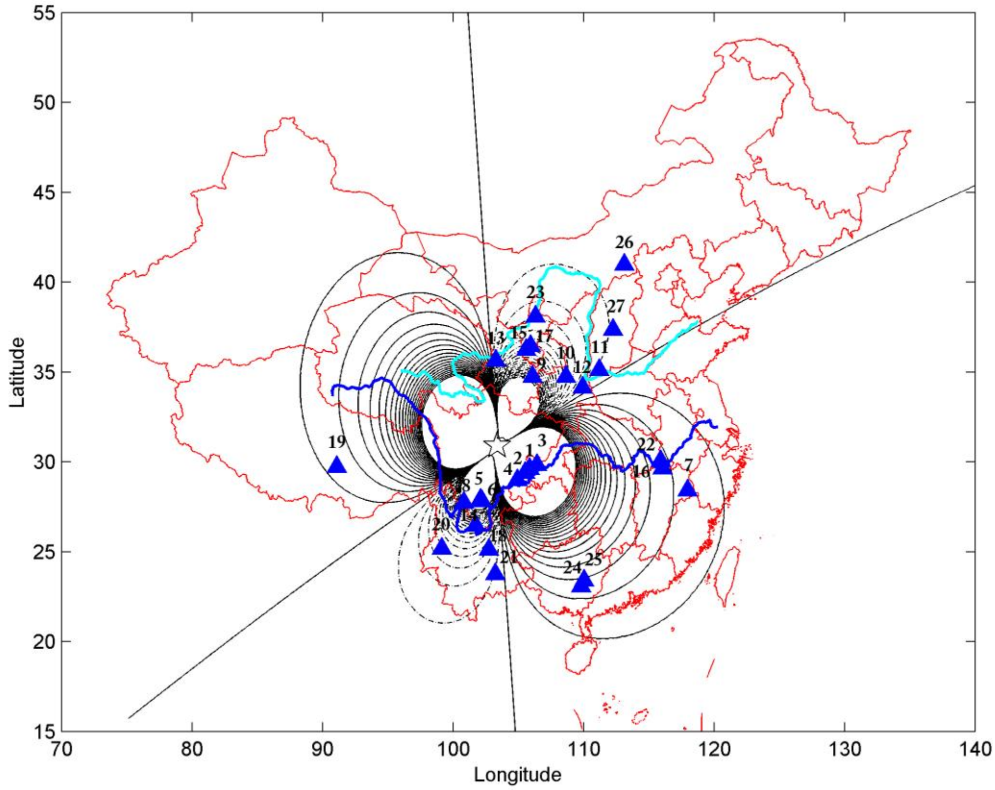
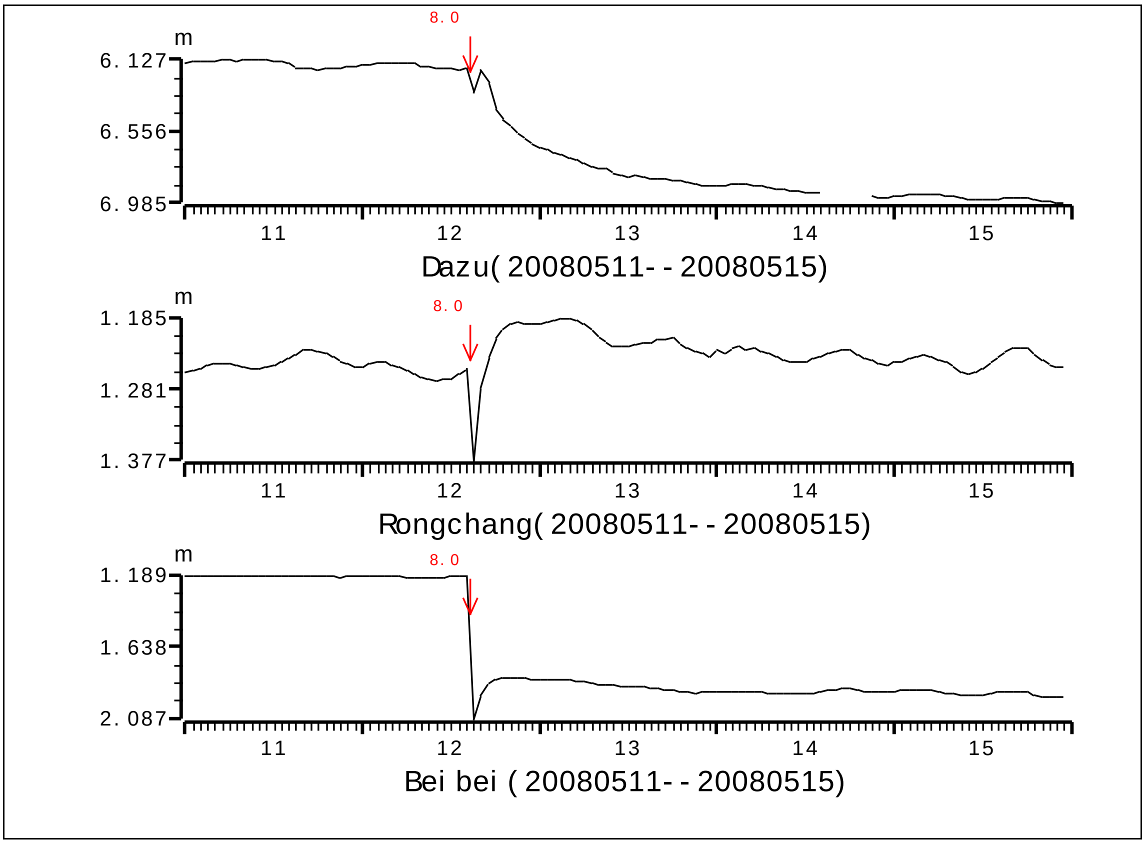
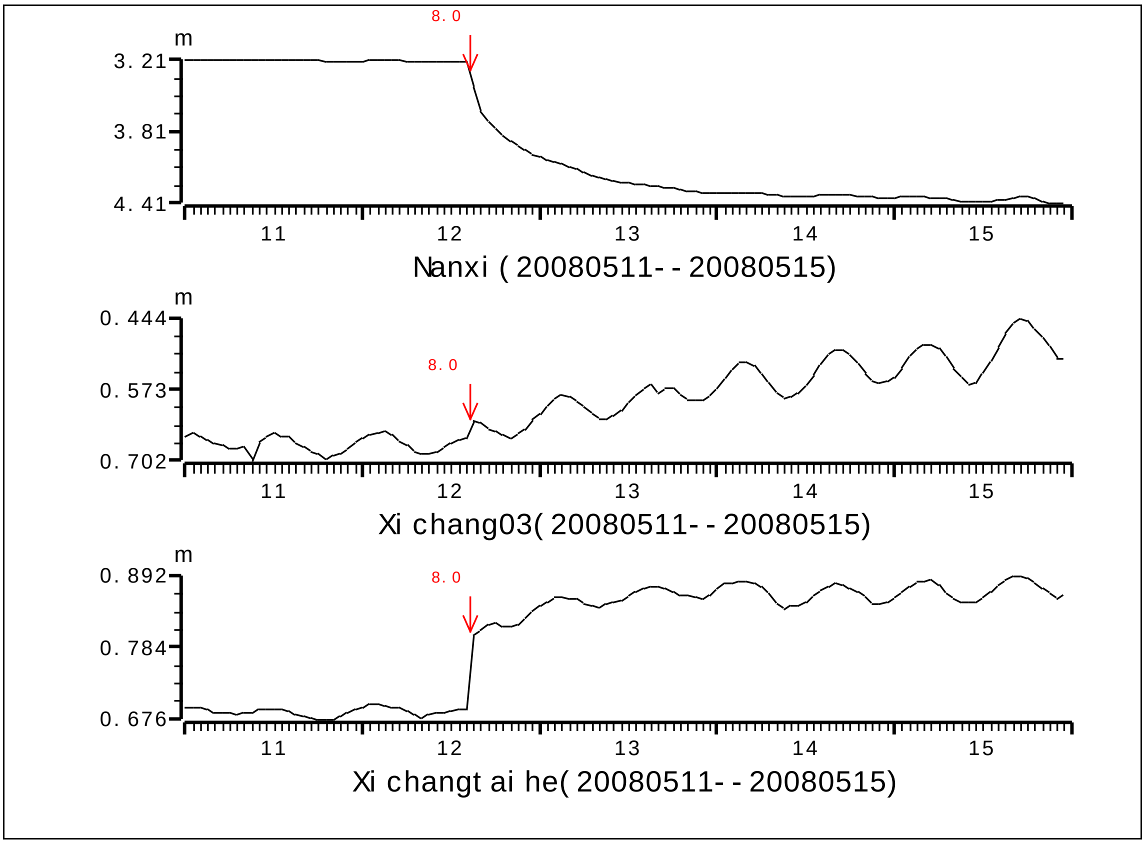


Figure5

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.

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0.702
8.0
11
12
13
14
15
Xichang03(20080511--20080515)
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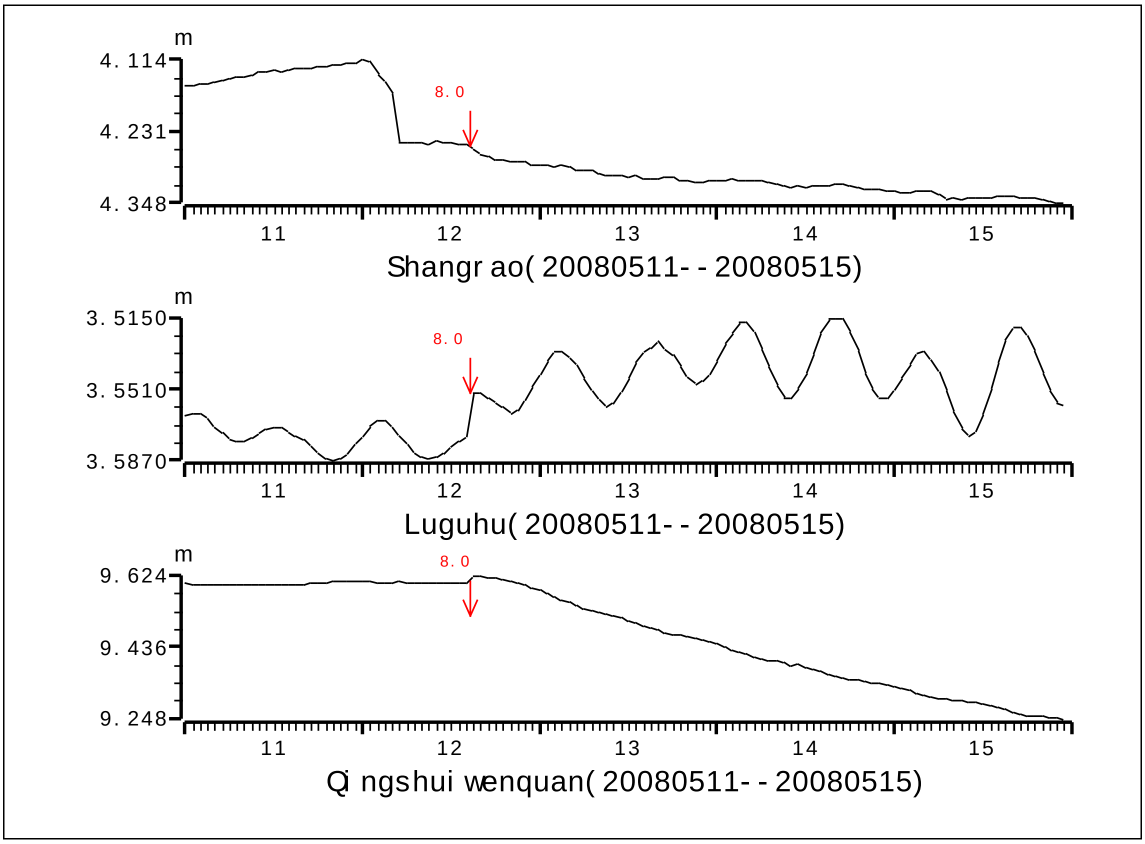
8.0
11
12
13
14
15
Xichangtaihe(20080511--20080515)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3.5510
3.5870
8.0
11
12
13
14
15
Luguhu(20080511--20080515)
m
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9.436
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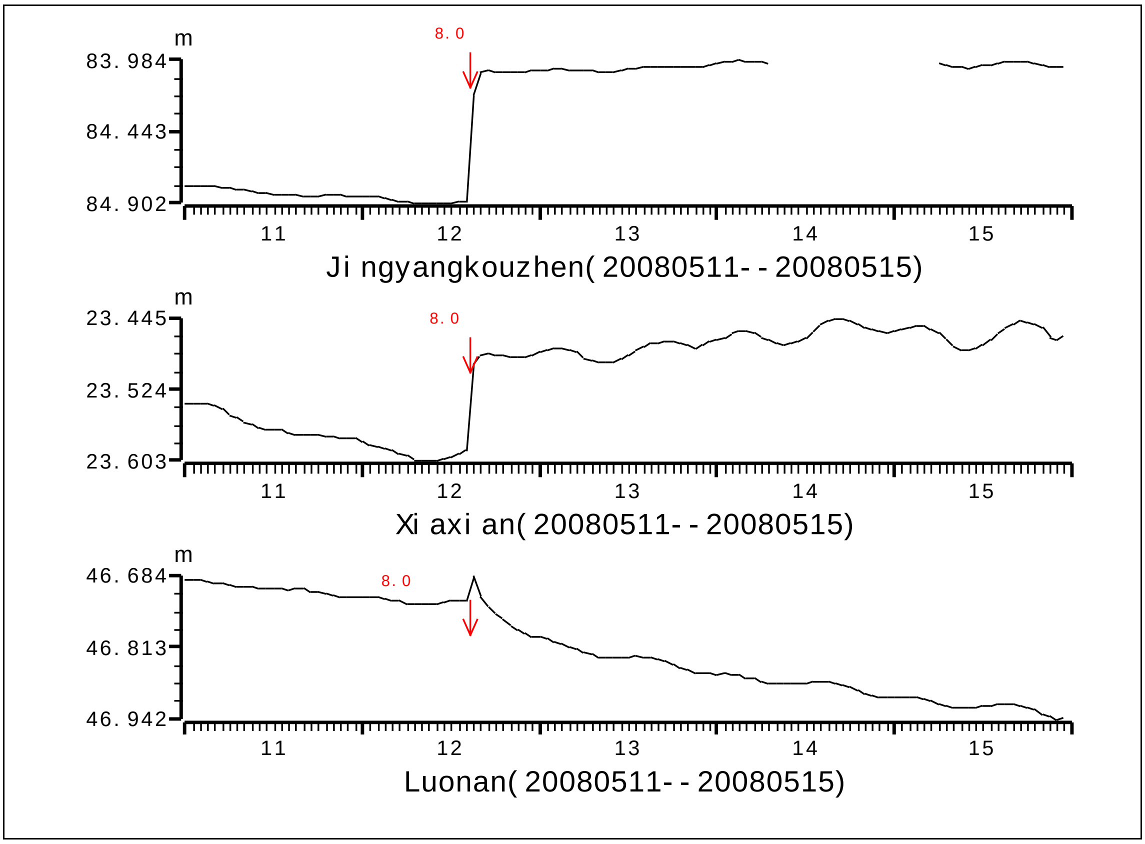
8.0
11
12
13
14
15
Qingshui wenquan(20080511--20080515)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23.524
23.603
8.0
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12
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15
Xiaxian(20080511--20080515)
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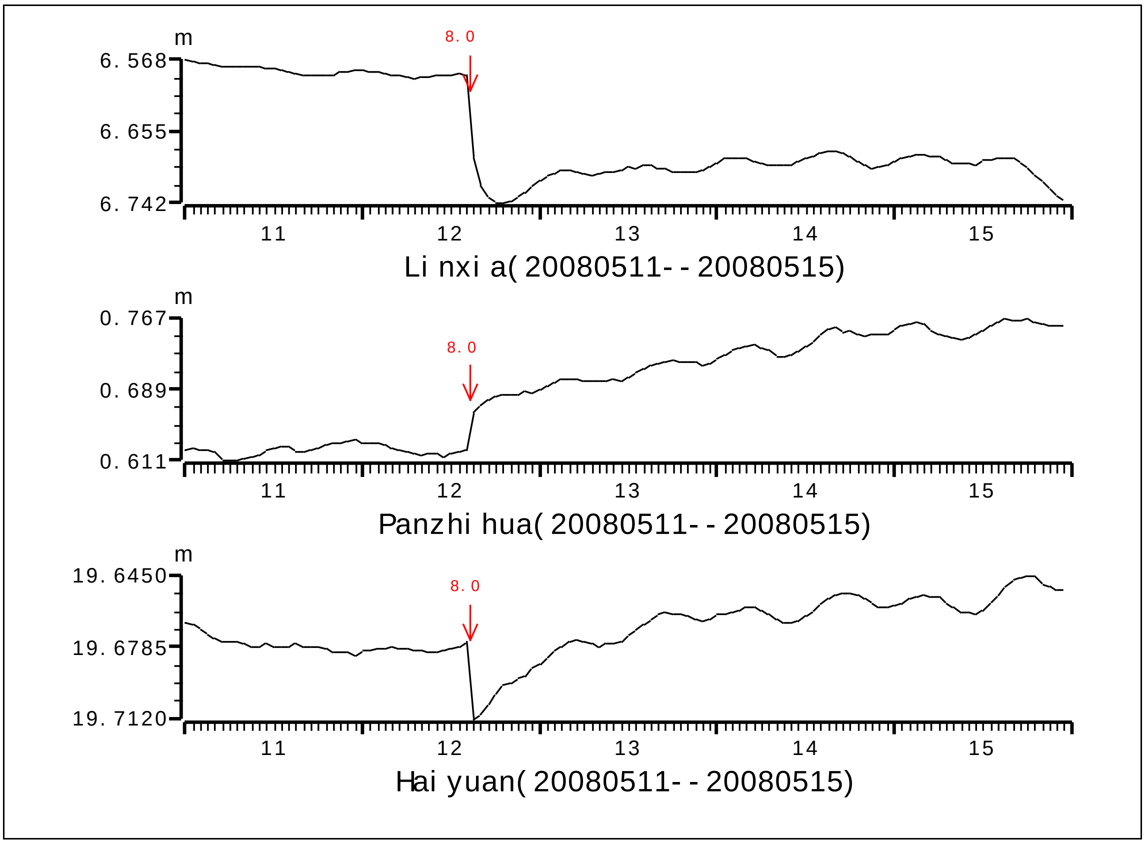
Luonan(20080511--20080515)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8.0
11
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13
14
15
Panzhihua(20080511--20080515)
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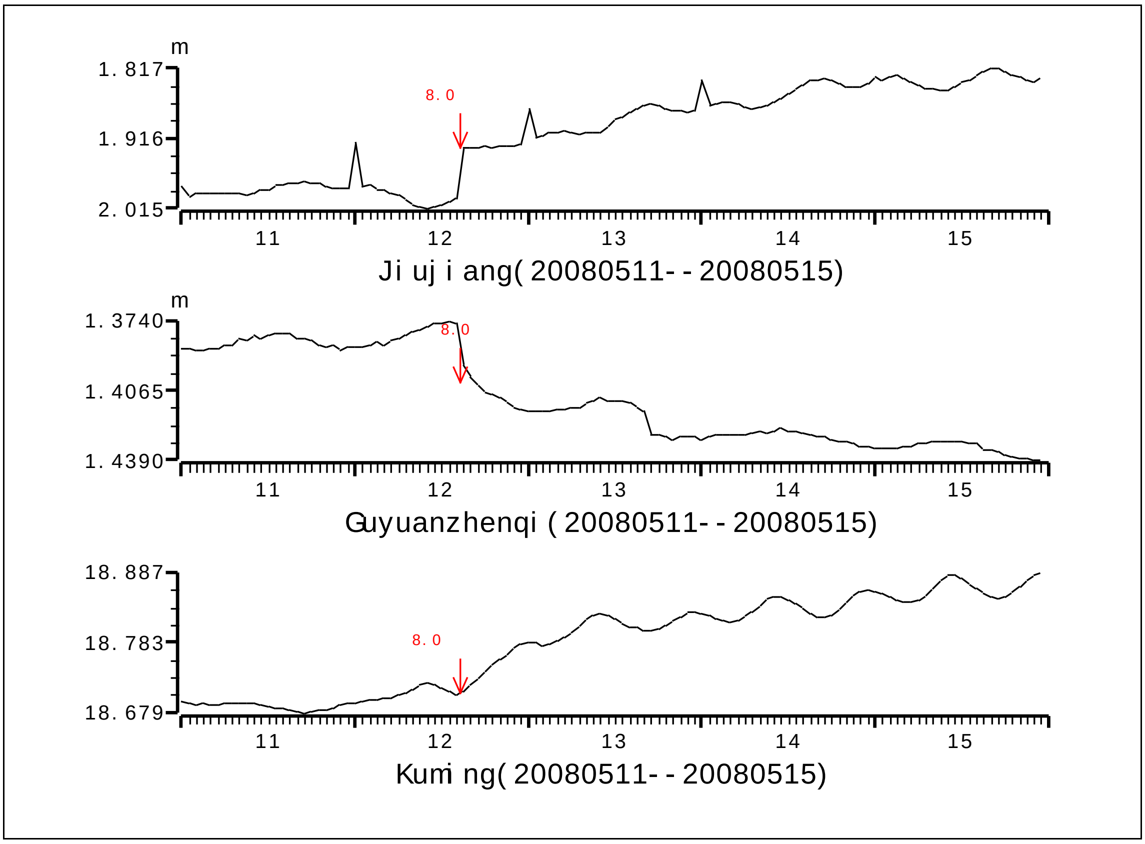
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1.4065
1.4390
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11
12
13
14
15
Guyuanzhenqi (20080511- - 20080515)
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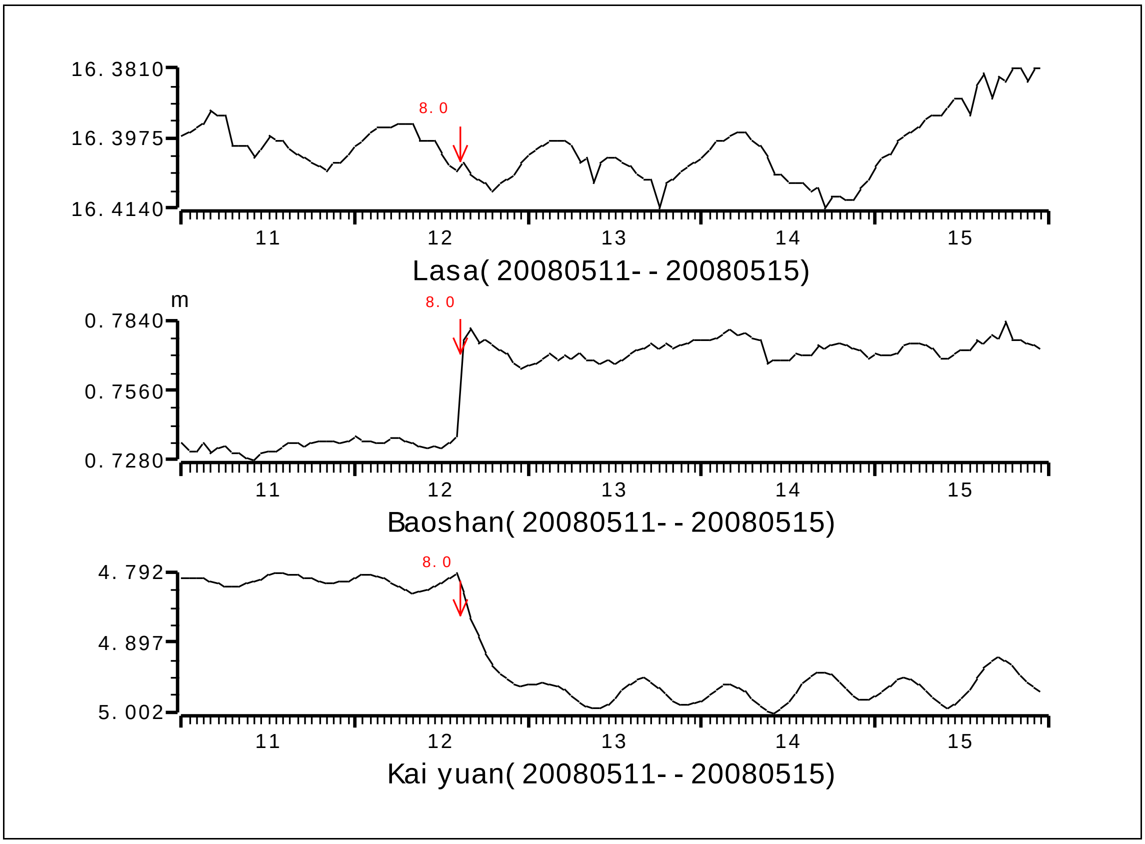
Kumi ng(20080511- - 20080515)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Baoshan(20080511- - 20080515)
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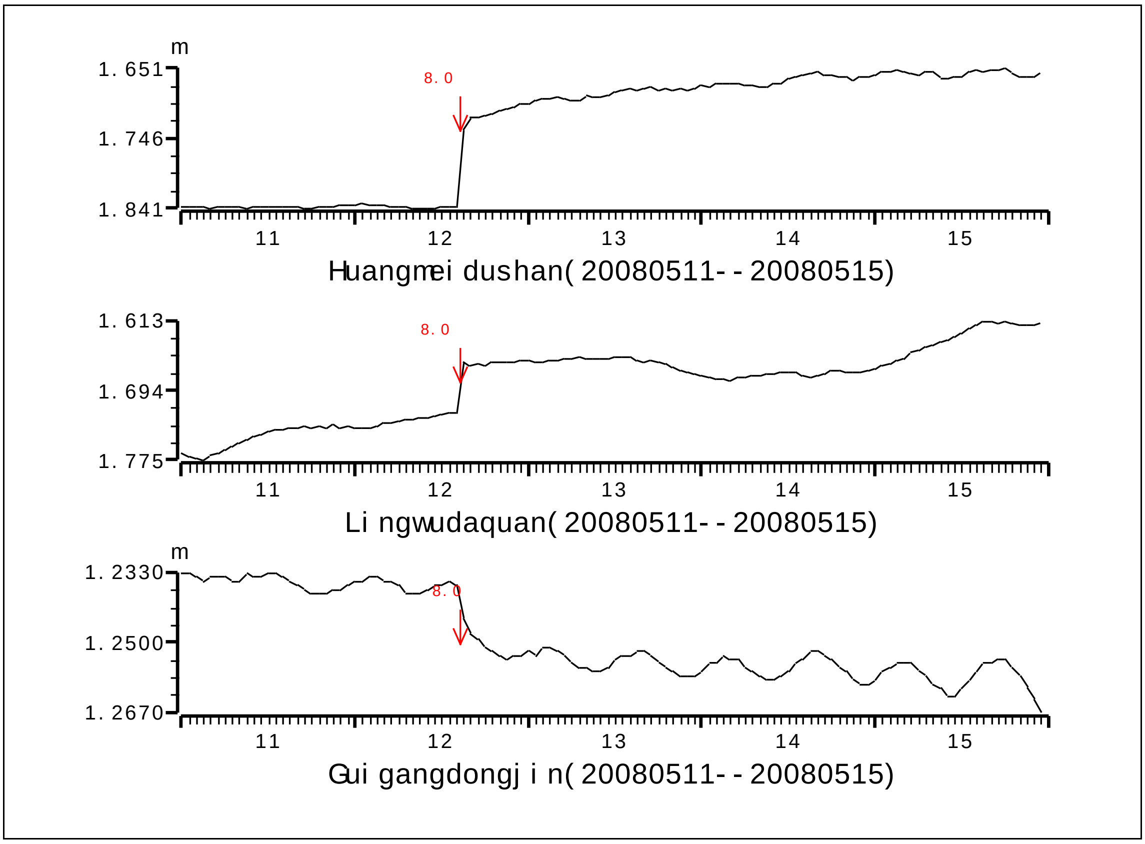
14
15
Kai yuan(20080511- - 20080515)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14
15
Lingwudaquan(20080511--20080515)
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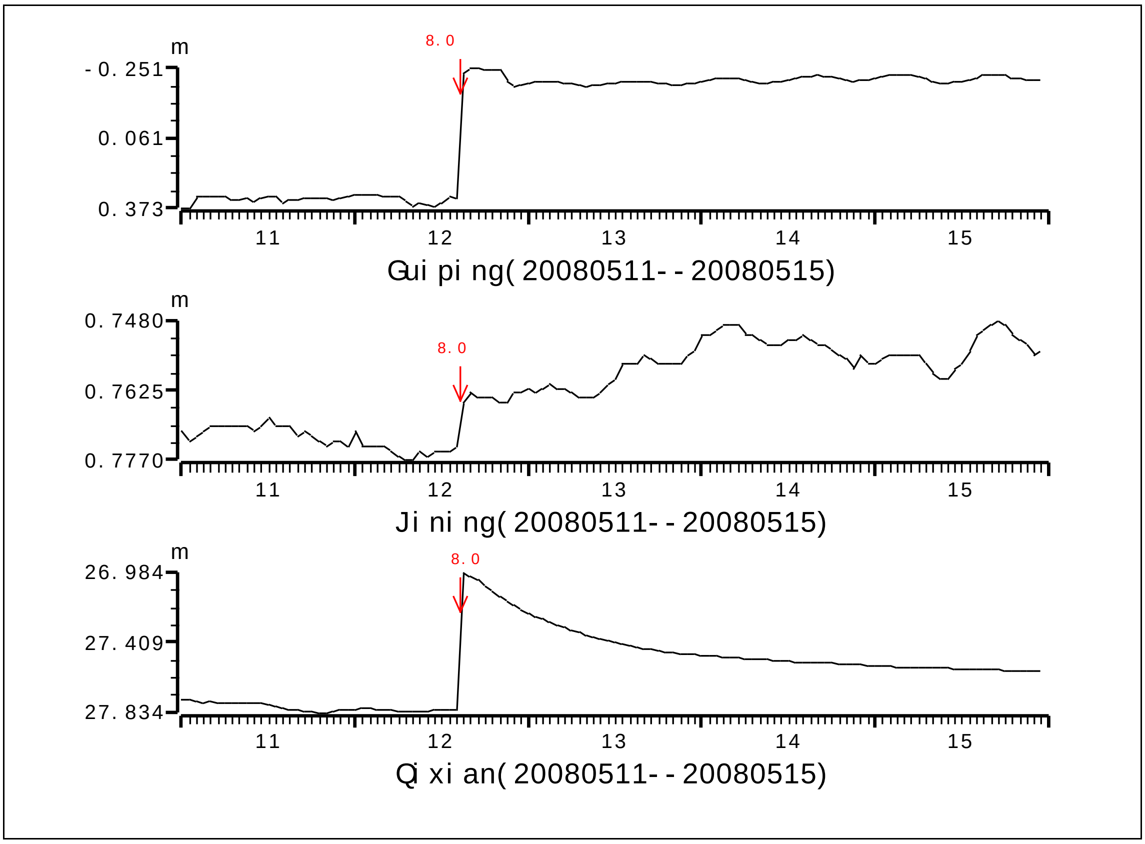
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8.0
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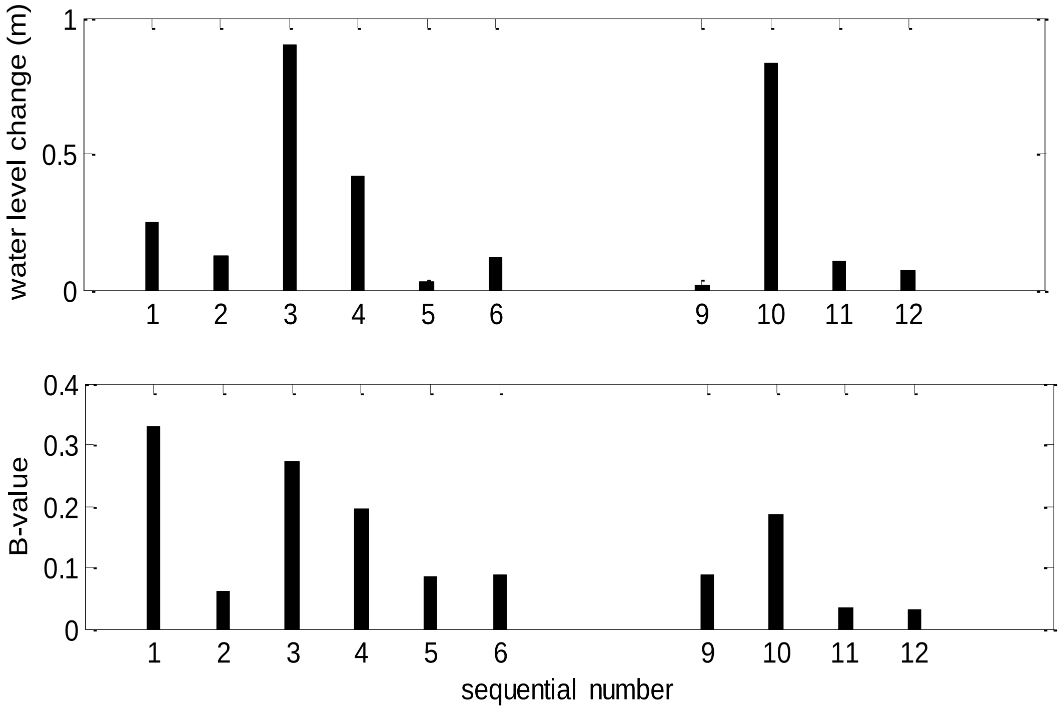
12
13
14
15
Jining(20080511--20080515)
m
26.984
27.409
27.834
8.0
11
12
13
14
15
Qixian(20080511--20080515)

Figure6

Figure 6. Water level changes and *B* values plotted according to the serial number of those 10 wells. In each group (group a, b, c, e, f), large pre-earthquake *B* values come with large co-seismic water level changes.



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