Seismoprogn. Observ. Territ. Azerb., V.11, N.1, 2014, pp.3-13

LOCAL SEISMIC AND HYDRAULIC EFFECTS CAUSED BY TOHOKU (M=9) EARTHQUAKE IN GEORGIA

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ABSTRACT. Presently, there are a lot of observations on the significant impact of such small external forcing's on the seismic regime, namely on the seismicity induced by wave trains of remote strong earthquakes, tides, reservoir exploitation, big explosions, magnetic storms, strong electrical pulses, etc. Many of such results still are subject of intense scientific discussions, but nevertheless are quite logical in the light of undisputable strong nonlinearity of processes underlying seismicity. One of main factors reducing local strength is the pore pressure of fluids, which is the scope of relatively new direction, so called hydroseismology. The stresses imparted by teleseismic wave trains according to assessments are 10^5 times smaller than confining stresses at the depth, where the tremors are generated due to a nonlinear effect of super-sensitivity to a weak impact.

According to publications, the majority of dynamically triggered events were observed in regions of extensional tectonics and high hydrothermal activity. At the same time some evidence of dynamic triggering from great Tohoku (M=9) earthquake has been obtained recently in the West Caucasus, which is a continental collision zone. Besides tremors, clear identical anomalies in water levels at passing S- and Love-Rayleigh teleseismic waves on the large part of territory of Georgia from Borjomi to Kobuleti has been observed. Their relation to seismic tremors is investigated.

Keywords: seismic regime, earthquake, teleseismic, waves, tremor, water level, hydraulic events, triggering, tectonics

1. INTRODUCTION

Presently, there are a lot of observations on the significant impact of such small external forcing's on the seismic regime, namely on the seismicity induced by wave trains of remote strong earthquakes (EQ), tides, reservoir exploitation, big explosions, magnetic storms, strong electrical pulses, etc. Many of such results still are subject of intense scientific discussions, but nevertheless are quite logical in the light of undisputable strong nonlinearity of processes underlying seismicity. One of main factors reducing local strength is the pore pressure of fluids, which is the scope of relatively new direction, so called hydroseismology. The stresses imparted by teleseismic wave trains according to assessments are 10^5 times smaller than confining stresses at the depth, where the tremors are generated [7]. Our laboratory data on stick-slip confirm reality of triggering and synchronization under weak mechanical forcing [4]. According to [2, 13, 14] the dynamically triggered tremors (DTT) can be related to the fluid pore pressure change due to passage of wave trains from remote strong earthquakes; that is why we carried out integrated analysis of seismic and WL data. Good correlation of WL signals with offsets of strongest teleseismic waves *(S, L, R)* should be some validation of hypothesis that perturbations in filtered seismic records of remote earthquakes (EQs) are indeed DTT events.

There are fundamental questions which have to be answered in order to make the domain of dynamically triggered seismicity useful instrument of earth crust physics. It is not clear why dynamic triggering (DT) is not observed everywhere, why it is observed mainly in some specified tectonic zones (extensional, hydrothermal areas), why the same dynamical forcing results in different response in similar tectonic zones, how ubiquitous is the phenomenon, is there a coupling of DT and water level change in boreholes, how DT can be related to the stress state in the depth, where the DT is forming, etc.

2. LOCAL SEISMIC EVENTS TRIGGERED BY TOHOKU EARTHQUAKE IN GEORGIA

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Manuscript received 30 Septembe 2014

The dynamic triggering due to the great Tohoku M 9 earthquake (2011), Japan was observed in local seismicity all around the globe [6, 9]. We presume that Tohoku EQ could also trigger local seismic events in Georgia (Caucasus), which is a continental collision area, separated from Japan by 7800 km. The teleseismic waves' phases onsets at Tbilisi and Oni seismic stations (s/s) for the main shock are as following (UTC/GMT): *p* - 05 57 41, *S* - 06 07 26; Love - 06 18 00, Rayleigh - 06 21 30. As the Caucasus is dominated by compression tectonics and the triggering examples from such areas are rare, presented data are significant for understanding triggering mechanisms.

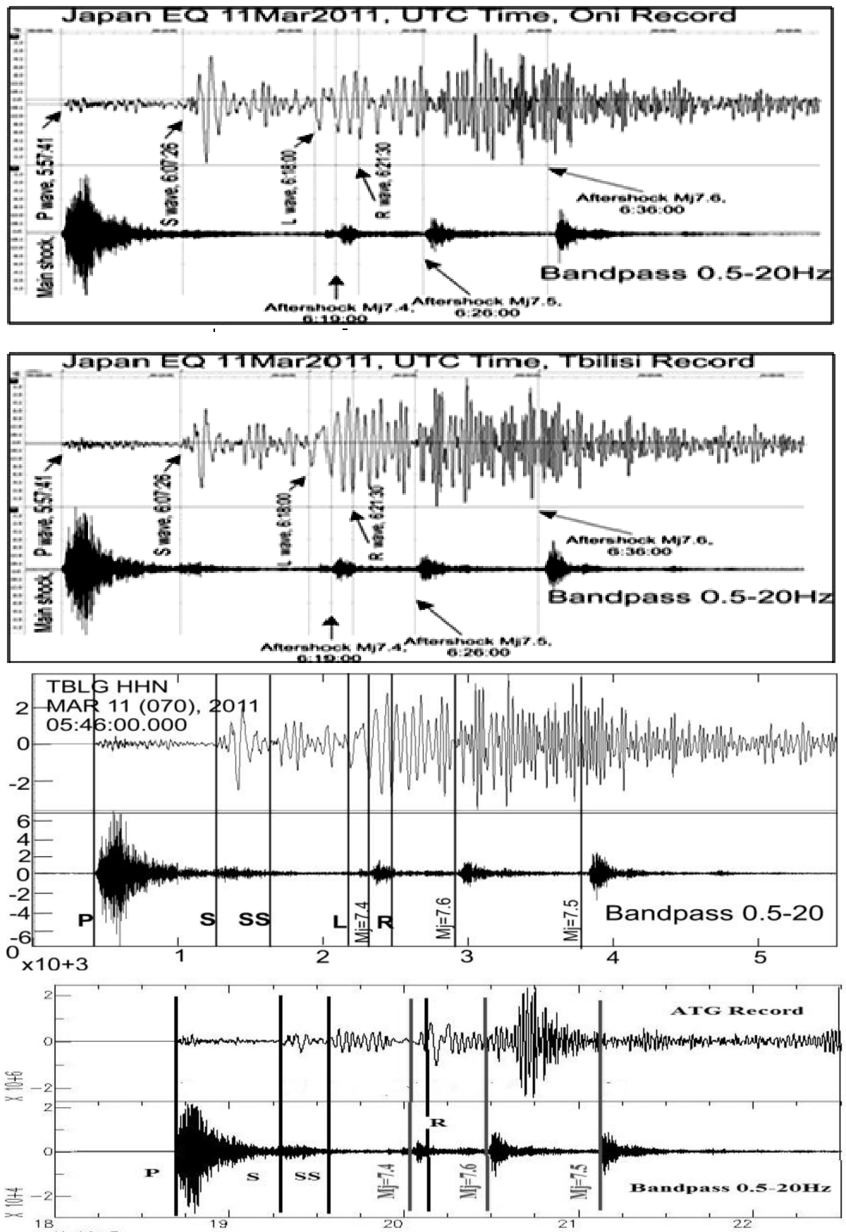


Figure 1. a, b, c, d. Broadband record of M 9 Tohoku EQ, Japan (11.03.2011) wave train NS-component (upper channel) and the same high-pass band (0.5-20 Hz) filtered record (lower channel) processed using the SEISMOTOOL program: (a) at Tbilisi s/s (b) the same for Oni s/s; high-pass band (0.5-20 Hz) filtered records (lower channel) processed by standard filtering procedure [3] (c) at Tbilisi s/s (NS-component); Altiagach s/s, Azerbaijan (40.860N, 48.940E). Seismoprogn. Observ. Territ. Azerb., V.11, N.1, 2014

Band pass (0.5-20 Hz) filtered records at two broadband seismic stations located in Oni (South slope of Greater Caucasus) and Tbilisi (valley of river Kura), separated by the distance 130 km processed using SEISMOTOOL program (Chelidze et al, in print) as well as by standard Procedure [3] are shown in Fig.1.

The sequence of triggered events is quite similar at both stations. This is an argument in favor of interpreting these signals as DTT events. The strongest event in the filtered signal coincides with the arrival of *p*-waves. The source of the strong seismic signal at *p*-wave arrival time in the bandpass 0.5-20 Hz filtered record (Fig. 1) is ambiguous: maybe it is a processing artifact caused by the specific range of filter as the burst practically vanishes at 5-20 Hz bandpass filtering. Thus in the following analysis we ignore this effect (see section 4). Nevertheless, we still prefer to use bandpass filter 0.5-20 Hz as the DTT corresponding to *S*, *L* and *R*- waves can be clearly distinguished in the filtered record.

The number of tremors increased 4-6 times in both Tbilisi and Oni stations during the first several hours after Tohoku EQ (Fig. 2). Of course some of tremors can be false and can be related to strong aftershocks – this should be studies in future.

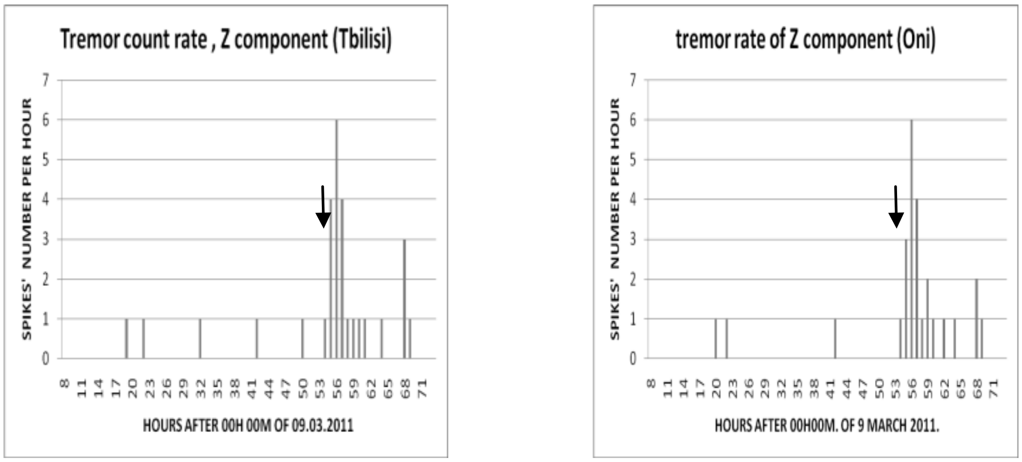


Figure 2. Tremor rate (number of local events per hour) before, during and after Tohoku event. Tohoku earthquake p-wave arrival time is marked by the arrow.

3. SEISMOHYDRAULIC EFFECTS IN GEORGIA RELATED TO TOHOKU EQ

Our next task was to compare the possible tremor signals with anomalies in water levels (WL) in deep wells' network in Georgia (Fig.3), operated by the M. Nodia Institute of Geophysics. Regular monitoring by this network is going on for several decades. It was important to find WL anomalous changes and compare them with teleseismic waves' phases as well as to assess pressure and stress changes of correlated seismic and WL signals: according to Brodsky et al (2003) the tremors can be triggered by fluid pore pressure change during teleseismic wave passage. Generally, WL respond to the EQ depends on the distance of the well to the ruptured fault: i. Very close to the fault intensive shaking may increase opening of fractures, i.e. and it cause rock dilatation and consequently, WL dropdown; ii. Outside this zone, but still very close to the fault shaking can consolidate loose sediments causing sudden upraise of WL; iii. In the intermediate field both positive and negative signs of sustained WL change are observed, which are explained by permeability changes; iv. Lastly, in the far field (which is our case) mainly correlated with seismic wave oscillations of WL are observed (hydroseismograms), sometimes accompanied with sustained WL change. As the seismic impact is instantaneous, it is expected that pore water has no time to flow, which in turn means that the WL response is undrained [13].

Figure 3. Network of WL borehole stations in Georgia.

WL monitoring network in Georgia includes the following deep wells: Kobuleti, Borjomi, Axalkalaki, Marneuli, Lagodekhi, Ajameti and Oni (Table 1, Fig.3).

Table 1. Locations and depths of wells in Georgia

Location	Depth of well, meters		Location	Depth of well, meters
Kobuleti	2000		Akhalkalaki	1400
Marneuli	3505		Ajameti	1339
Borjomi70	1339		Lagodekhi	800
Borjomi Park borehole is located on the top of the fault.	30		Oni	255

The sampling rate at all these wells is 1/min (except Oni, where the sample rate is 1/10 min). Measurements are sensors MPX5010 (resolution 1% of the scale) recorded by datalogger XR5 SE-M remotely by modem Siemens MC-35i using program LogXR; datalogger can acquire WL data for 30 days at the 1/min sampling rate. The range of WL measurements by this equipment is 0-100 cm.

Below (Fig.4) we show water level respond to a series of Japan earthquakes 11 March 2011 with following *p*-wave arrival times of the main shock and aftershocks: a) M 9; time - 05: 57; b) Mw7.4, time - 06.19; c) Mw =7.9, time – 06: 26; d) Mw =7.7, time – 06: 36.

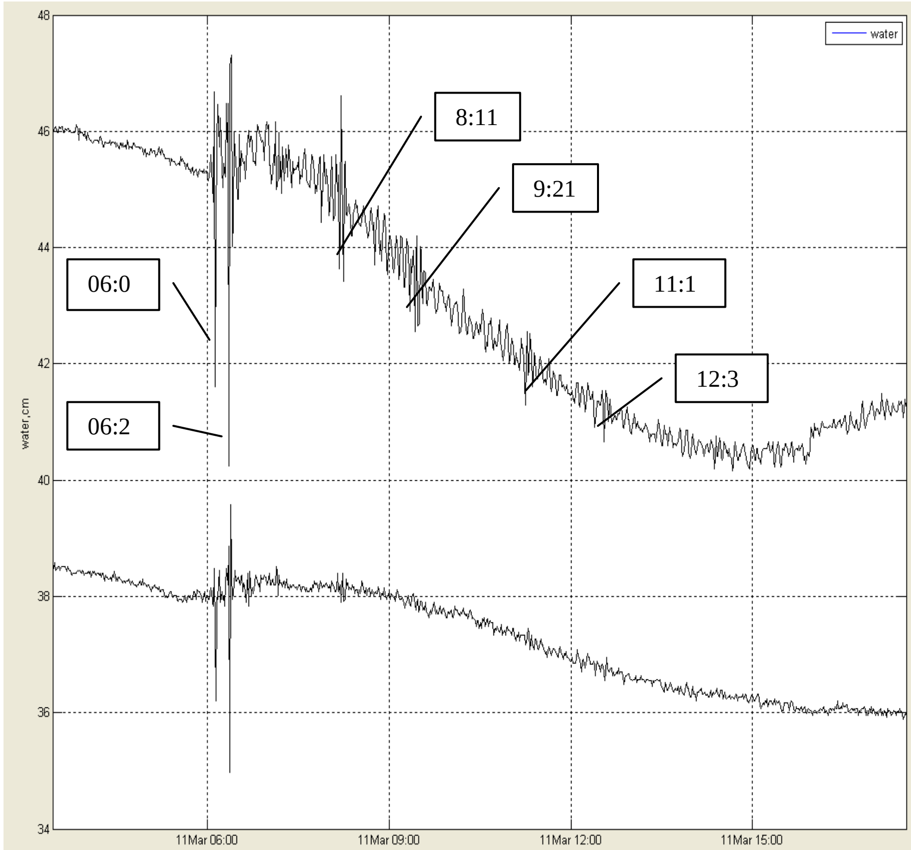
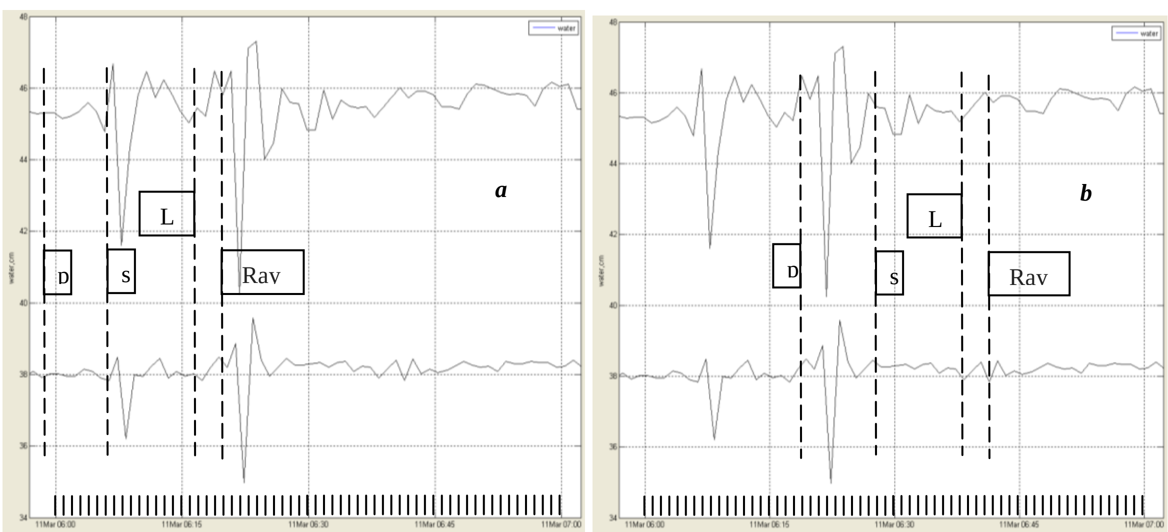


Figure 4. Water Level change in Kobuleti (top) and Borjomi Park (bottom) before and during Japan M9 earthquake,11 March 2011 in conventional units (1/min sample rate): compressed 24 hour record. The lines with time data point to some late teleseismic surface (G-R) waves' onsets.



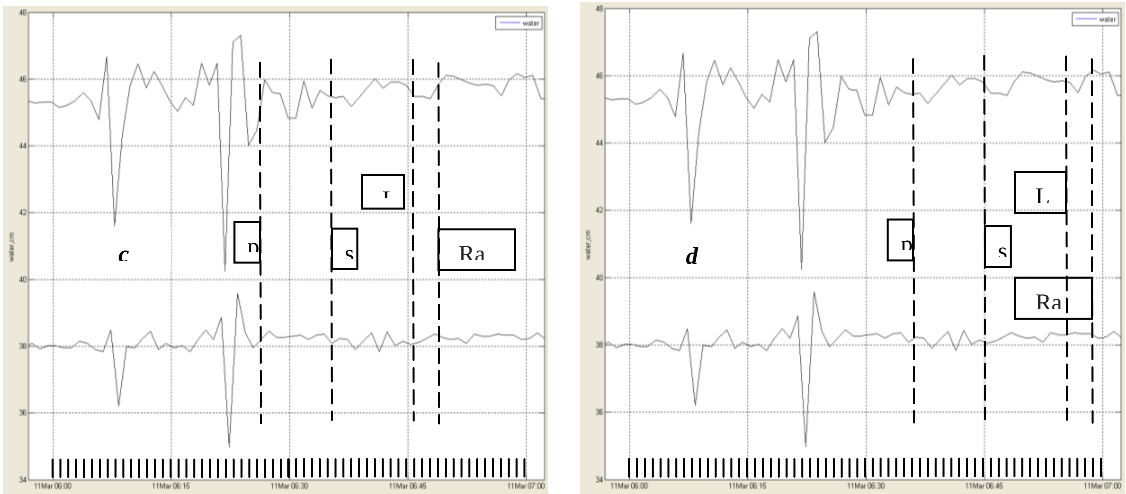


Figure 5 a, b, c, d. Water Level change in Kobuleti BorjomiPark (top) and (bottom) before and during first 30 minutes of Japan M9 earthquake on March 2011 in conventional units (1/min sample rate), aftershocks and seismic phases; expanded records. On (a, b, c, d) the dashed lines mark onsets of the teleseismic p, S, Love and Rayleigh waves generated by the main shock Mw9 (a), and aftershocks Mw7.4 (b), Mw7.9 (c), Mw7.7, (d) correspondingly. The best correlation between teleseismic wave phases and pattern of strong WL signals is for the main shock (Fig. 5a). The most important phases of strong aftershocks (*S*, *L*, *R*) pass to late to cause major WL signals (Figs. 5 b, c, d).

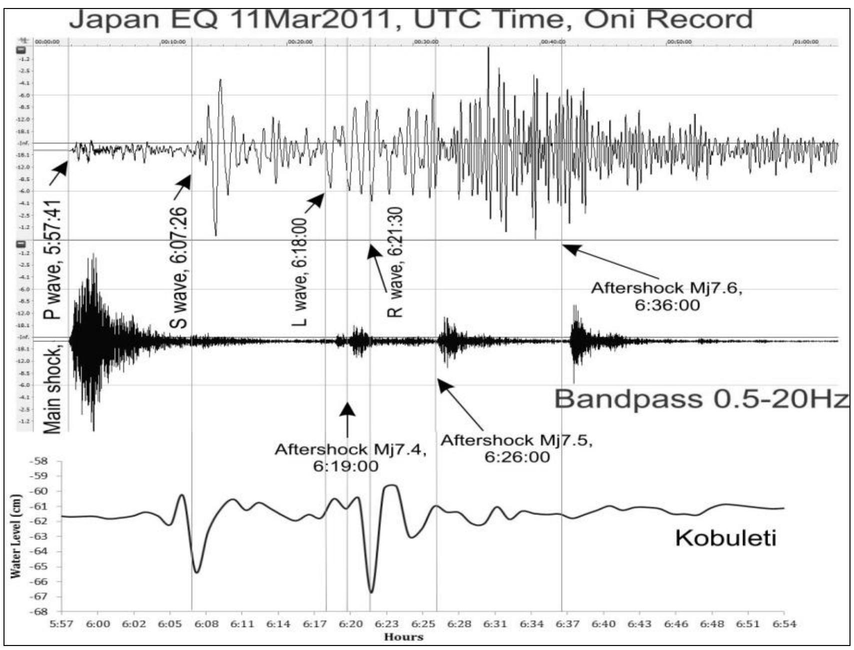


Figure 6. The integrated plot of seismic and WL events in Georgia during Tohoku event. It is evident that the first strong WL perturbation at 06.07 correlates definitely with S-wave offset; no aftershocks are recorded at that time. The second strongest WL event between 06:19 and 06:22 is coincides with both onset of L/R wave package (06:18-06:21) and aftershock M_j7.4 at (06:19). Note, however, that the foreshock of Tohoku event (2011-03-09) of the same magnitude (M_j7.3) as well as stronger aftershocks at 06.26 and 06.36 do not produce any characteristic WL oscillations; thus the most probable explanation of WL effect at 06:19 is the passage of R wave.

As the WL values in different wells change in a very wide range in order to show their reactions on the same plot, the signals from the *i*-th borehole *(WL_i)* are plotted in conventional units, namely, they are shifted along y-axis according to the expression: $(WL_i) = WL_o$–[min(WL_i)]+offset; where WL_o is the observed WL, [min(WL_i)] is a minimum WL in borehole for the year 2011 and the offset is a constant, needed to fit WL curves into the same plot. For example, on the Figs (4, 5) the value of [min(WL_1)] for Kobuleti is -106 cm, the value of offset= 0; for Borjomi [min(WL_2)] is - 523 cm; offset - 6 cm.

WL signals from the Tohoku events are fixed in Kobuleti, Borjomi Park, (Figs. 4, 5, 6), Marneuli and Oni boreholes.

Figs. 5, 6 demonstrate a striking similarity of hydraulic responses to passage of some phases of teleseismic waves from Tohoku event in areas separated by 300 km: namely, to S-wave and to summary impact of Love and Rayleigh waves (as the sampling rate was 1/m, it is impossible to separate reaction to L and R waves). Besides phases of the main shock, the strong aftershocks of Tohoku EQ also can affect WL; the first strong (M_j7.3) aftershock reach Tbilisi 11 March 2011on 06:19. Note, however, that the foreshock of Tohoku event (2011-03-09) of the same magnitude (M_j7.3) as well as even stronger aftershocks at 06.26 (Mw7.9) and 06.36 (Mw7.7) do not produce any characteristic WL oscillations. Thus the most probable explanation of WL effect at 06:19 is the passage of the main shock generated R wave.

The best correlation between teleseismic wave phases and pattern of strong WL signals is for the main shock (Fig. 5a). The most important phases of strong aftershocks (*S, L, R*) pass too late to cause major WL signals (Figs. 5 b, c, d).

4. LOCAL TREMOR OR AFTERSHOCKS' TELESEISMIC WAVES?

Analyzing Fig.1 we can conclude that some signals in the filtered main shock record, which are candidates for identification as local tremors, coincide with the onsets of teleseismic waves from the strong aftershocks of Tohoku. This complicates the diagnostics of filtered seismic signal as a dynamically triggered tremor (DTT). The observation that DTT are generated mainly by *S, L* and *R*-waves can help in discrimination of aftershocks' teleseismic waves from the local tremors. The indirect argument in favor of such approach is that strong WL effects are correlated with main shocks' *S* and *L-R* wave arrivals and not observed at the *p*-wave arrival (Fig. 5).

The additional indication for discrimination of mentioned impacts (DTT or teleseismic wave passage) can be obtained by considering seismohydraulic data also. According to Fig. 5 the most important wave phases of strong aftershocks (*S, L, R*) pass too late to cause major WL signals. The analysis of the Tohoku foreshock (09 March 2011; M7.3), which is in the same magnitude range as the first aftershocks, shows that characteristic fast oscillations are absent in the WL response to the foreshock, which means that the impact of Tohoku EQ both foreshock and aftershocks is too weak to generate WL perturbation.

5. SPECTRUM OF WL OSCILLATIONS FOLLOWING TOHOKU EQ

It is evident that after Tohoku EQ water level undergoes characteristic oscillations, which decay in a dozen of hours (Fig. 4). The spectrum of WL oscillations for 10^{th} and 11^{th} March is shown in Fig. 8.

After Tohoku EQ in the spectrum of WL oscillations appear several spikes around frequencies 2.4 10^{-3}; 4.0 10^{-3}; 4.9 10^{-3}; 6.2 10^{-3}; 7.2 10^{-3} Hz, which seem to be harmonics of the first mode with a multiplier approximately 1.3. The intensity of harmonics is especially high during the first 30 min after EQ. The reverberations are absent in the spectrum for the 10^{th} March (Fig.8a, black curve). The spectrogram of the same WL record also shows intensive signals around above frequencies (Fig. 8b). The observed reverberations in WL hardly can be explained by the excitation of so called Kraukis waves which propagate back and forth along fluid-filled fractures of the aquifer, emitting periodic seismic signal [12]. The frequency of Krauklis wave depends on the fracture width, shear modulus of the solid, fluid density and the ratio of shear and longitudinal waves: in order to be in the observed range, the system should contain unrealistically long and thin cracks.

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and the ratio of shear and longitudinal waves and is of the order of tens of Hz in typical aquifers: in order to be in the observed low-frequency range (Fig. 8), the system should contain unrealistically long and thin cracks.

The most probable explanation of WL oscillations with periods 2-7 min is the impact of mantle surface waves (Love and Rayleigh), which can excite seismic signals with periods up to about 500 s [1], which fits to the observed WL oscillations' frequency range 2.4 10^{-3} - 7.2 10^{-3} Hz (Figs. 4, 8). This interpretation is confirmed by a good coincidence of WL signals and multiple surface R-waves arrival times.

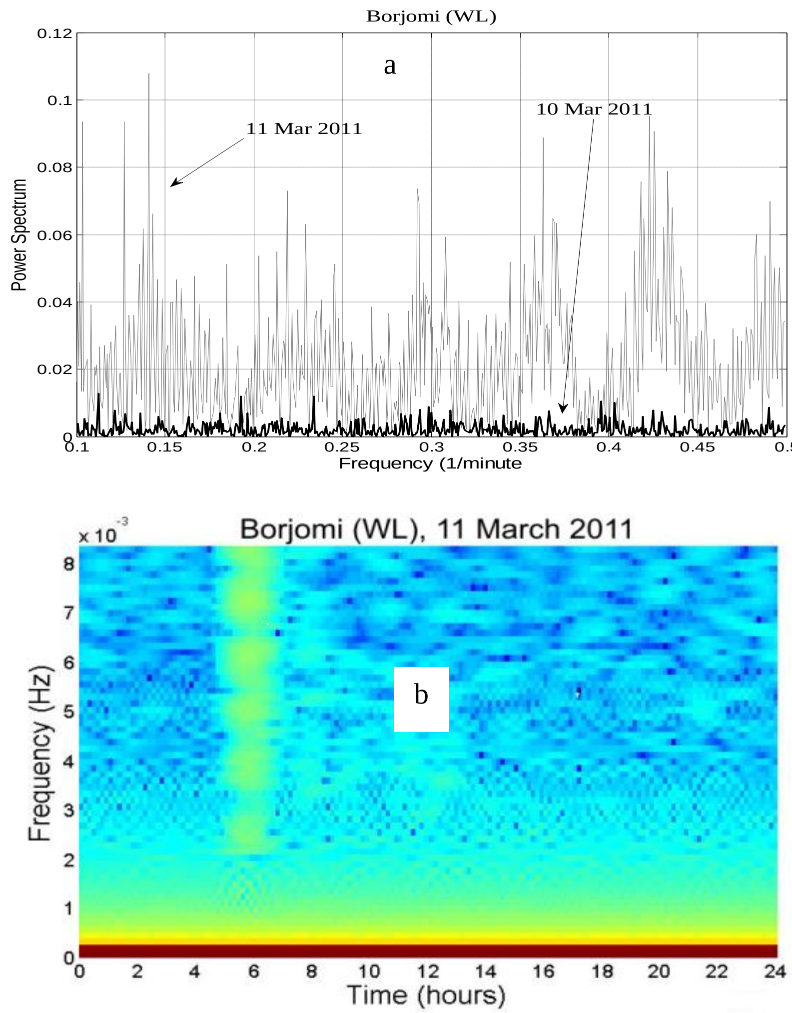

Figure 7. Spectrum (a) and spectrogram (b) of WL oscillations in Borjomi borehole before, during and after Tohoku EQ. The black curve in (a) is a background spectrum calculated for 10 March and grey curve - for 11March. The last one shows several strong spikes at frequencies 2.4 10^{-3}; 4.0 10^{-3}; 4.9 10^{-3}; 6.2 10^{-3}; 7.2 10^{-3} Hz (periods 2-7 min), which are visible in the spectrogram (b) also.

6. FUSION OF SEISMIC AND WL EVENTS IN GEORGIA RELATED TO TOHOKU EQ

In the Table 2 the seismological and WL information on the Tohoku EQ impact in Georgia is summed. Here and in the Table 3 $\Delta(WL)_{mR}$, is the maximal WL signal (peak-to-peak amplitude) for

R-group waves, cm; $\Delta(WL)_{mG}$ is the maximal WL signal (peak-to-peak amplitude) for L/G-group waves, cm; ΔP_{mG} is the maximal water pressure change during L/G-wave passage, KPa;ΔP_{mR} is the maximal water pressure change during R-wave passage, KPa; v_S, v_G and v_R are correspondingly peak ground velocity (wave rate) for S, L/G and R waves in cm/s; $\Delta\sigma_G$ is the dynamic stress change for L/G waves, KPa; $\Delta\sigma_R$ is the dynamic stress change for R waves, KPa; ΔL_S, ΔL_L and ΔL_R are accordingly displacements due to displacements for S, L/G and R waves in cm; χ is the amplification factor of seismic waves in the well calculated as the amplitude of water level oscillations to the particle velocity in the seismic waves, $\chi = \Delta(WL)_m/v$ in units m/(m/s) [2].

Love/Rayleigh phases induce maximal WL displacement (peak-to-peak amplitude), which vary from 4 cm in Borjomi to 10 cm in Oni. The hydraulic effect (displacement) is 4- 10 times larger than seismic L or R wave displacement. In order to estimate dynamic stress [3] we measure the peak ground velocity for the Love and Rayleigh waves in the instrument-corrected NS and vertical component seismograms, respectively. Then we calculate the corresponding dynamic stress ($\Delta\sigma$) based on equation: $\Delta\sigma = G\ (du/dt)\ /v$, where G is the average shear rigidity of crust - 35 GPa, v - phase velocities accordingly 4.0 and 3.5 km/s for Love and Raylegh waves, (du/dt) is a Peak Ground Velocity (PGV) respectively. Measured PGV for Love and Rayleigh waves are 0.09 and 0.1cm/sec, respectively. So the corresponding dynamic stress is about 10 KPa.

The different WL response in different boreholes to practically the same mechanical impact (11 KPa) is explained by difference in aquifers' transmissivity/storage: large amplitudes of WL are favored by high transmissivity/low storativity [2, 13].

Table 2. Seismic and hydraulic reaction to Tohoku (M9) EQ in Georgia

Site name	$\Delta(WL)_{mR}$, cm	ΔP_{mR} KPa	v_S cm/s	ΔL_S cm	v_L cm/s	ΔL_L, cm	v_R cm/s	ΔL_R, cm	$\Delta\sigma_G$ $\Delta\sigma_R$ KPa	χ m/(m/s)
Kobuleti	8	0.8	0.1	1	0.09	1.4	0.11	1.2	11	80
BorjomiPark	4	0.4	0.1	1	0.09	1.4	0.11	1.2	11	89
Oni	10	1	0.1	1	0.09	1.4	0.11	1.2	11	73

Generally, earlier it was accepted that the main impact on WL should cause Rayleigh wave as it provokes volume change. The strong enough response of WL to *S*- and Love waves passage was considered less probable as these wave does not lead to volumetric strain. Nevertheless recent observations document WL coherent oscillations with *S*- and Love waves [13]. Our data also confirm strong impact of S-wave on WL in Georgia boreholes (Figs. 5, 6).

There is also very interesting detail on the WL plot for Borjomi well (Fig.4, trace for Z-component): clear delayed WL perturbations are registered at the following times: 08:11, 09:21, 11:14 and 12:33, which cannot be associated with aftershocks.

The possible explanation of these anomalies is the passage of late teleseismic phases, namely multiple surface waves circling the Earth: according to Peng et al (2011) they also trigger seismic events. The most effective in delayed triggering of microearthquakes are the first three groups of multiple surface waves (G1-R1, G2-R2, etc). Indeed, analysis of seismograms shows that exactly at above mentioned times of WL perturbations arrive multiple surface waves R2 (08.10), R3 (09.21), R4 (11.13) and R5 (12.30), which travelled correspondingly 289, 431, 649 and 791 degrees [1]. Thus, we show that multiple surface R waves can generate not only local microseismicity, but also significant WL signals. On the other hand WL does not respond to the arrival of G-group Love waves (G1, G2 etc 0– Figs.4, 8). Thus the WL signals, recorded at 08:11, 9:21, 11:14 and 12:33 are definitely triggered by passing multiple surface R-waves (Fig.4, 8). Table 3 summarizes corresponding seismic and WL data.

We can conclude that though the stress change imparted by multiple surface waves of both G and R-groups are comparable (Table 3), the WL responds strongly only to R-waves impact. This result is in agreement with the statement that for WL change porous space should consolidate or dilate;

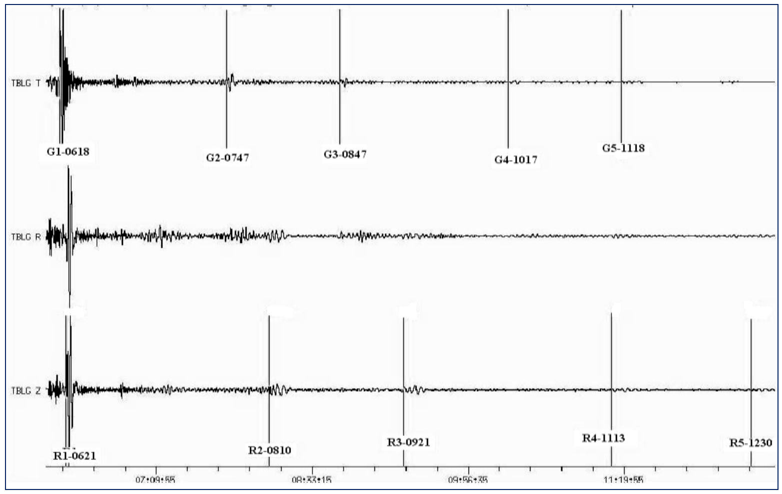

Figure 8. Seismogram with arrivals of multiple surface G and R waves at Tbilisi s/s.

Table. 3. Seismic and hydraulic response to the multiple surface waves (R2, R3, R4, R5 and G2, G3, G4, G5) of Tohoku, M9, EQ in Kobuleti, Georgia.

Site name	$\Delta(WL)_{mR,}$ cm	ΔP_{mR} KPa	$\Delta(WL)_{mG}$ cm	ΔP_{mG} KPa	v_G cm/s	$\Delta\sigma_G$ KPa	v_R cm/s	$\Delta\sigma_R$ KPa	χ
Kobuleti	3.20	0.32	-	-	G2 – 0.030	3.0	R2 - 0.020	2.0	160
	1.65	0.17	-	-	G3 – 0.015	1.5	R3 - 0.018	1.5	90
	1.26	0.13	-	-	G4 – 0.007	0.7	R4 - 0.008	0.7	160
	0.90	0.09	-	-	G5 – 0.003	0.3	R5 - 0.006	0.5	150

Rayleigh waves give rise to volumetric strain what satisfies this model [13]. S and L waves have not volumetric component and accordingly they should not affect WL, but the recent data [8, 13] as well as our results show that S and SS waves also significantly change WL. The mechanisms suggested for explanation of the latter observation include anisotropic poroelastic effect [2], permeability enhancement of fractured rocks due to removal of blocking elements by oscillating fluid [13] or just strong anisotropy/heterogeneity of aquifer rocks, which can add volumetric component to a shear displacement; such effect is absent in isotropic homogeneous material.

Thus our new observation obtained by integrated analysis of seismic and water level records (hydroseismograms) manifests that multiple surface R waves generate not only local microseismicity [10] but also significant synchronous WL signals (unlike less efficient multiple surface G waves) Figs. 4 and 8.

7. Conclusion

The great Tohoku earthquake provokes significant local seismic and hydraulic events in Georgia triggered by passage of teleseismic wave trains, mainly by S and Lave-Rayleigh waves. Some seismic triggered events are masked by offsets of strong aftershocks of Tohoku earthquake. Comparison of WL anomalies with seismic waves' phases can help to discriminate triggered events from aftershock signals. The strong hydraulic events with amplitude 8-10 cm, correlated with passage of *S*- and *L-R* waves are caused by mechanical displacement of the order of 1 cm, i.e WL response to displacement is amplified 8-10 times due to mechanical stress change 11 KPa. It should be noted that the WL response at wells separated by hundreds of km are practically identical. Besides WL response to the first arrivals of S and Love–Rayleigh phases, there are some clear delayed WT perturbations, which closely correlate with the passage of multiple surface Rayleigh waves: R2, R3, R4, R5.

Further development of sensitive devices, dense networks and processing methods will develop a new avenue in seismology, which can be defined as DT microseismology and which will study systematically small earthquakes and tremors, especially events, triggered and synchronized by remote strong earthquakes (magnitudes 7-8). These events at present are ignored by routine seismological processing and are not included in traditional catalogues. At the same time, DT microseismic events contain very important information on geodynamical processes and can give clues to understanding fine mechanism of nonlinear seismic process and may be, even contribute to the problem of earthquake forecast.

8.Acknowledgments

The authors express their gratitude to the Rustaveli National Science Foundation of Georgia (Project FR/567/9-140/12) for financial support.

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