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M9 Tohoku earthquake response in Georgia – possible local tremors and hydroseismic effects

T. Chelidze[1], I. Shengelia[2], N. Zhukova[1], T. Matcharashvili[1,2], G. Melikadze[1], G. Kobzev[1]

1. *M. Nodia Institute of Geophysics at Iv. Javakhishvili Tbilisi State University Alexidze str. 0160, Tbilisi, Georgia, tamaz.chelidze@gmail.com*
2. *Ilia State University*

Abstract

Presently, there are a lot of observations on the significant impact of strong remote earthquakes on underground water and local seismicity regimeы (so called nonvolcanic or dynamical tremors). On the other side, teleseismic wave trains give rise to several hydraulic effects in boreholes, namely water level oscillations, which mimic seismograms (hydroseismograms). Both these effects are closely related to each other as one of main factors reducing local strength of rocks is the pore pressure of fluids.

Some evidence of possible dynamic triggering from great Tohoku (M9) earthquake has been obtained recently in the West Caucasus. Besides tremors, clear identical anomalies on the large part of territory of Georgia from Borjomi to Kobuleti in the borehole water levels has been observed at passing S- and Love-Rayleigh teleseismic waves of Tohoku earthquake. We presume that coincidence of possible tremor signal with water level anomaly (oscillation) makes much more reliable event classification as a triggered one. We also report a new observation on water level oscillations during passage of multiple surface Rayleigh waves.

1. Introduction.

Presently, there are a lot of observations on the significant impact of strong remote earthquakes on underground water regime and triggered local seismicity (so called nonvolcanic or dynamical tremors). The stresses imparted by teleseismic wave trains according to assessments are 10^5 times smaller than confining stresses at the depth, where dynamical tremors are generated. Many of such results still are subject of intense scientific discussions due to the weakness of wave trains from remote earthquakes, but nevertheless are quite logical in the light of undisputable strong nonlinearity of processes underlying seismicity: the tremors are generated due to a nonlinear effect of super-sensitivity to a weak impact. On the other side, teleseismic wave trains give rise to several hydraulic effects in boreholes, namely water level oscillations, which mimic seismograms (hydroseismograms). Both these effects, seismohydraulic and triggered tremors are closely related to each other as one of main factors reducing local strength of rocks is the pore pressure of fluids: this is the scope of relatively new direction, so called hydroseismology. Thus we presume that coincidence of possible tremor signal with water level anomaly (oscillation) makes much more reliable triggered seismic event classification.

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 (Hill and Prejean, 2009; Prejean and Hill, 2009). Our laboratory data on stick-slip confirm reality of triggering and synchronization under weak mechanical forcing (Chelidze et al, 2010). According to (Brodsky et al, 2003; Wang, C.-Y., Manga, 2010; Zhang and Huang, 2011) 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 (Parsons et al, 2014), 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 possible tremors triggered by Tohoku earthquake in Georgia

The dynamic triggering due to the great Tohoku M 9 earthquake (2011), Japan was observed in local seismicity all around the globe (Gonzalez-Huizar et al. 2010; Obara and Matsuzawa, 2013). The main characteristic of DT events are peak dynamic values of stress (T_P) or strain (ε_P); for shear waves $T_P \approx G\,(u_P/v_S)$ and $\varepsilon_P \approx u_P/v_S$; here G is the shear modulus, u_P is particle' peak velocity and v_S is velocity of the shear wave. Analysis of the field data gives values of T_P from 0.01MPa to 1MPa (ε_P from 0.03 to 3 microstrain). We assume that such large scatter is due to the impact of another important factor, namely, the local (site) strength of earth material, which is highly heterogeneous. Thus what matters is not the absolute value of T_P or ε_P, but the difference between local stress and local strength or resistance to failure (Chelidze and Matcharashvili, 2013, 2014). This is why in some areas high T_P do not trigger local seismicity and, on contrary, some areas manifest DT even at low peak stresses (Hill, Prejean, 2009;). One of main factors reducing local strength is the pore pressure of fluids, which is the scope of relatively new direction, so called hydroseismology (Costain and Bollinger, 2010).

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. Though it is accepted that extensional tectonics and presence of hydrothermal sources favors dynamical triggering of local tremors (Prejean and Hill, 2009), the latest analysis shows that weak "seismicity rate significantly increases immediately after (~45 min) M7 mainshocks in all tectonic settings and ranges" (Parsons et al, 2014).

Band pass (0.5-20 Hz) filtered records at two broadband seismic stations (s/s) located in Oni (South slope of Greater Caucasus) and Tbilisi (valley of river Kura), separated by the distance 130 km, as well as in Azerbaijan (Altiagach station ATG) were analyzed. The digital records were processed using SEISMOTOOL program (Chelidze et al, 2014) as well as by standard procedure (Chao et al, 2012) are shown in Fig.1. The sequence of triggered events is quite similar at seismic stations of the region.

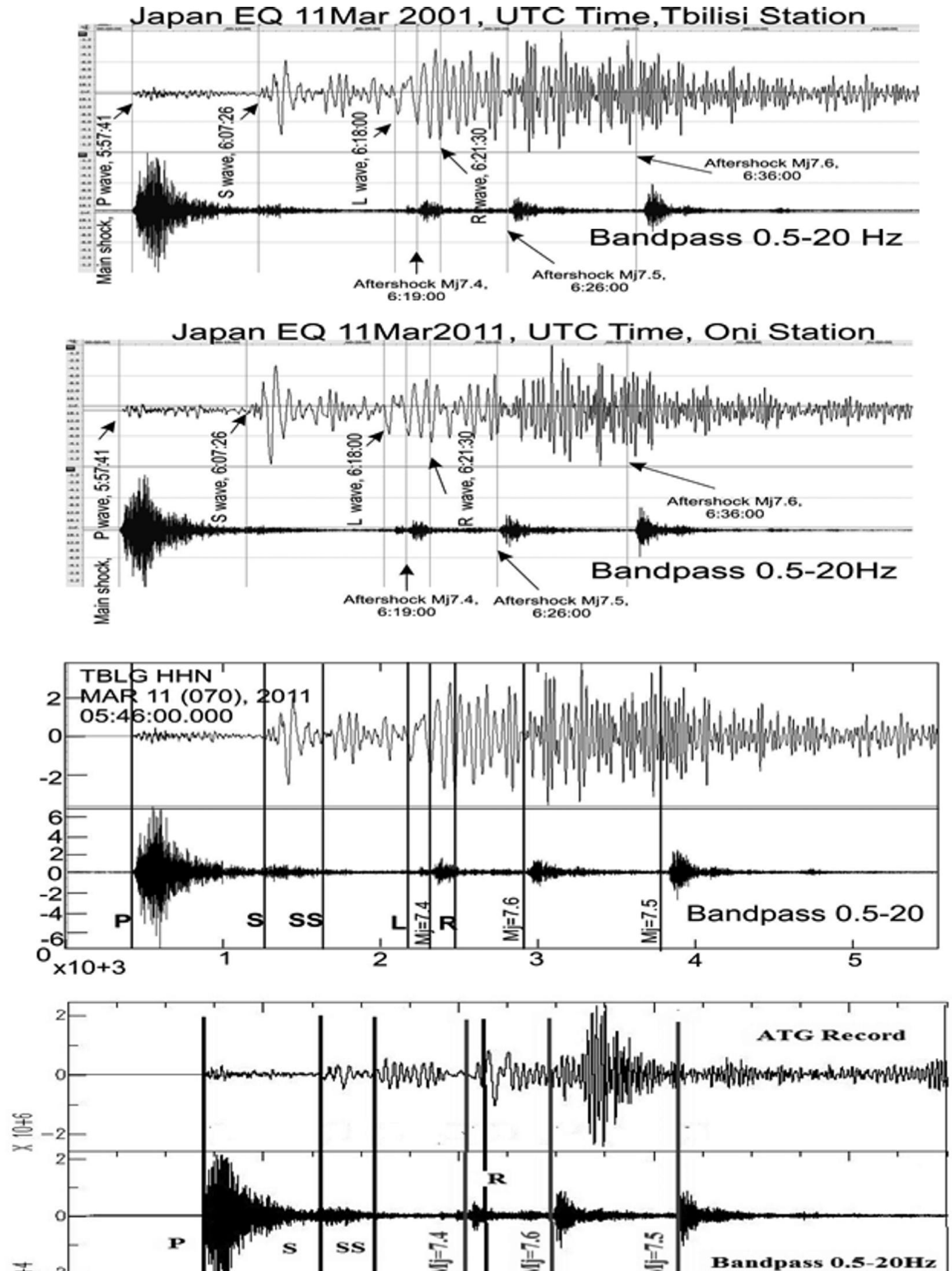

Fig. 1. a, b, c, d. Broadband record of M 9 Tohoku EQ, Japan (11.03.2011) wave train NS-component (upper channel) and the same high-pass band (0.5-20 Hz) filtered record (lower channel). Phases of seismic waves from the mainshock and arrival times of *p*-waves of strong aftershocks are marked by vertical lines. (a) at Tbilisi s/s; (b) the same for Oni s/s; in (a) and (b) the original records were processed using the SEISMOTOOL program (Chelidze et al, 2014). (c) at Tbilisi s/s (NS-component); (d) Z-component at Altiagach s/s (ATG), Azerbaijan. In (c) and (d) original EQ records (upper channels) were filtered by high-pass band 0.5-20 Hz filter (lower channel) using standard filtering procedure (Chao et al, 2012).

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 *p*-wave 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 as well as signals from strong aftershocks can be clearly distinguished in the filtered record.

As the seismic network in Caucasus is not dense and high quality the standard approach to tremors' identification (Chao et al, 2012, 2013; Peng et al, 2010) is not effective here. We suggest the proxy method for discrimination of tremors generated by remote EQ that can be used even at one station; of course this does not allow calculation of location, depth and other details. Our approach is an analog of Reasenberg's spatial β-parameter (Reasenberg, 1985) in the temporal domain. We used the following criteria for presumed tremor discrimination:

i. Deviation by 3 sigma (3 times standard deviation) from the background seismic record scanned for several hours before EQ - this is considered as a lower threshold of presumed tremor signal. An additional condition is that the oscillation amplitudes of tremors should exceed ±0.05 of the maximal amplitude of the considered EQ (Fig.1, b, c) or ±0.25 for EQ record in db (Fig 1 a, b) or the corresponding value in counts /bit during 5 s (500 counts).. In case of Tohoku EQ this criterion corresponds to (2500 counts/bit) for amplitudes during 500 counts.

ii. 3 sigma deviation lasts at least 5 s (500 counts)

The number of such "tremors" increased 4-6 times in both Tbilisi and Oni stations during the first several hours after Tohoku EQ and cumulative curve increases drastically during passage of teleseisms (Fig. 2). Of course the strong aftershocks also can contribute to the statistics and the problem needs thorough consideration. According to USGS data Tohoku EQ from 05.46 11 March 2011 (local UTC) produces 112 aftershocks in the range M4.7-M7.9, which is much more than number of presumable tremors in Oni s/s, which equals 29. The onsets of 13 "tremor" signals coincide with the arrivals of *p*-waves of M6 and stronger aftershocks. The cause of other "tremor" signals is not clear as there were so many aftershocks that their *p*-wave arrival time coincidence with tremor signals can be quite accidental (for example such accidental coincidence was found for aftershock of M4.9, but we know that the isolated events of such magnitude from Japan do not cause tremor-like signals). So we can presume that at least half of "tremors" in Fig. 2 can be of local genesis, but still the significant increment in number of possible tremors remains. Still, in future the problem of discrimination of local tremors' signals from aftershocks should be analyzed in detail to avoid wrong interpretation (http://earthquake.usgs.gov/earthquakes/seqs/usc0001xgp.php).

As the pure seismological information is not enough to recognize local tremors in following we try to involve into interpretation process local hydroseismic data because simultaneous appearance of local tremor-like signal and local WL oscillations during passage of wave trains from remote EQ especially when there are not strong aftershock's can be an solid argument for tremor identification (Brodsky et al, 2003).

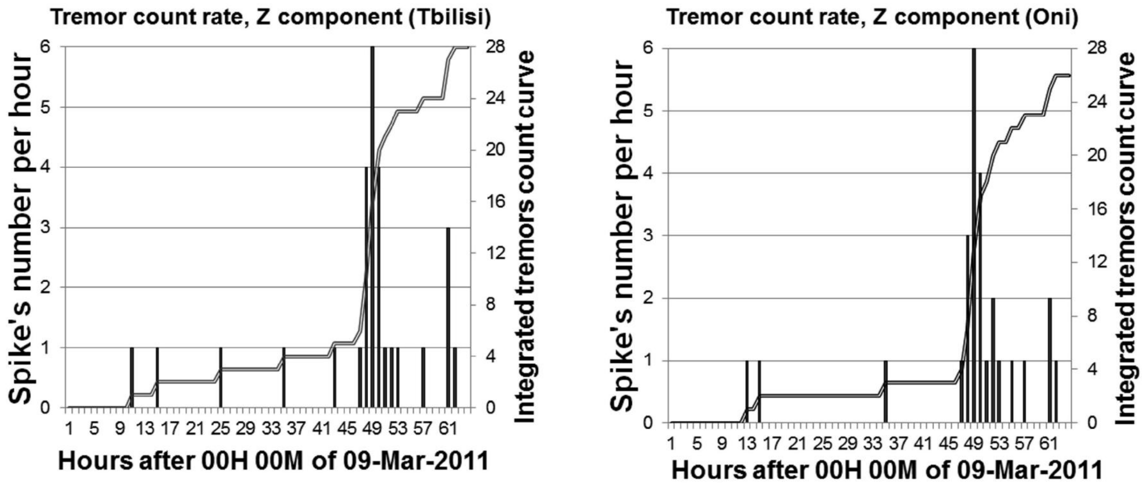

Fig. 2. Tremors' rate (number of presumable local events per hour) and cumulative curve of tremors before, during and after Tohoku event. Tohoku EQ p-wave arrival time is marked by the arrow.

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 (Wang et al, 2009; Zhang, Huang, 2011; Wang, Manga, 2010), WL respond to the EQ wave trains' impact 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.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;

Fig. 3. Network of WL borehole stations in Georgia

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 (Wang, Manga, 2010).
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. The oscillations due to the EQ impact last for 24-12 hours in various wells.

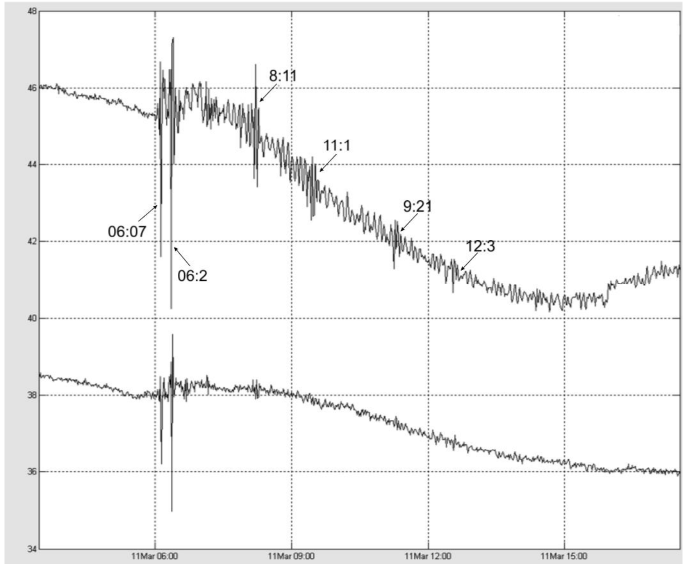

Fig.4. Water Level change in Kobuleti (top), Borjomi Park (middle) and Marneuli (bottom) boreholes before and during Japan M9 earthquake,11 March 2011 in conventional units (1/min sample rate): compressed 24 hour record.

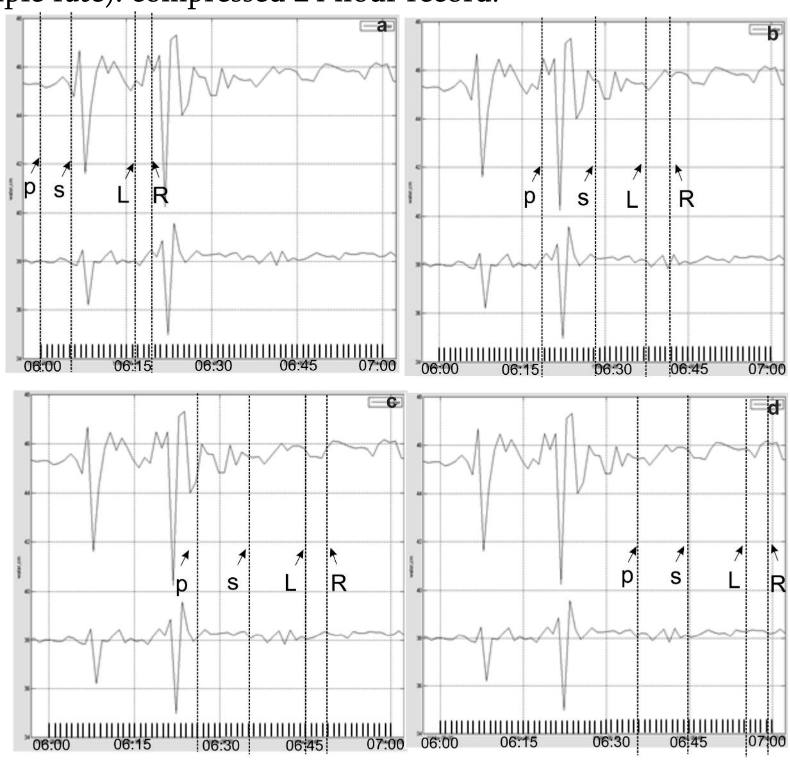

Fig.5 a, b, c, d. Water Level change in Kobuleti BorjomiPark (top) and (bottom) before and

during first 30 minutes of Japan M9 earthquake,11 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).

It was interesting to know whether the wells recording oscillations due to seismic waves respond also to earth tides. In Fig.6 the two-weeks' record of WL in Kobuleti well is presented: upper figure shows original record and lower one – the same record after elimination of atmospheric pressure effect. It is evident that Tohoku EQ oscillations are superimposed on the tidal variations and that both responses are of almost the same amplitude – several (5-6) cm.

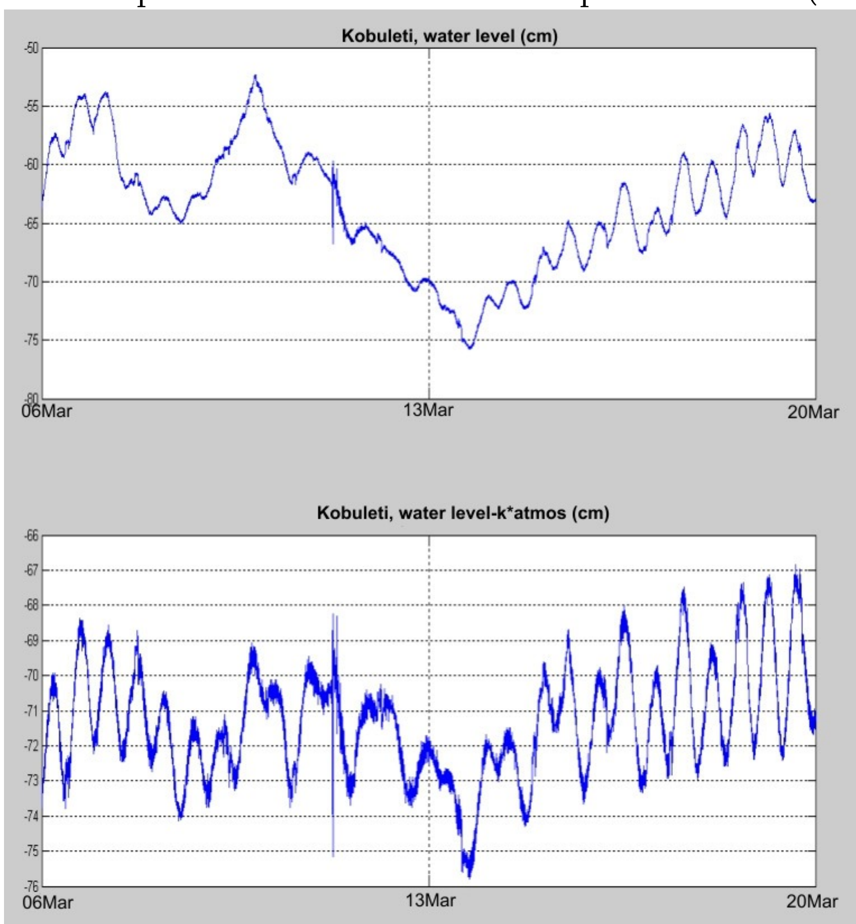

Fig. 6. WL record at Kobuleti borehole 06-20 March 2011: (a) original record of WL, absolute values, cm; (b) WL after removal of atmospheric pressure effect in reduced units; note well-marked tidal variations.

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. Reduced water level value obtained after this manipulation is shown on vertical axes of Figs. 4,5.

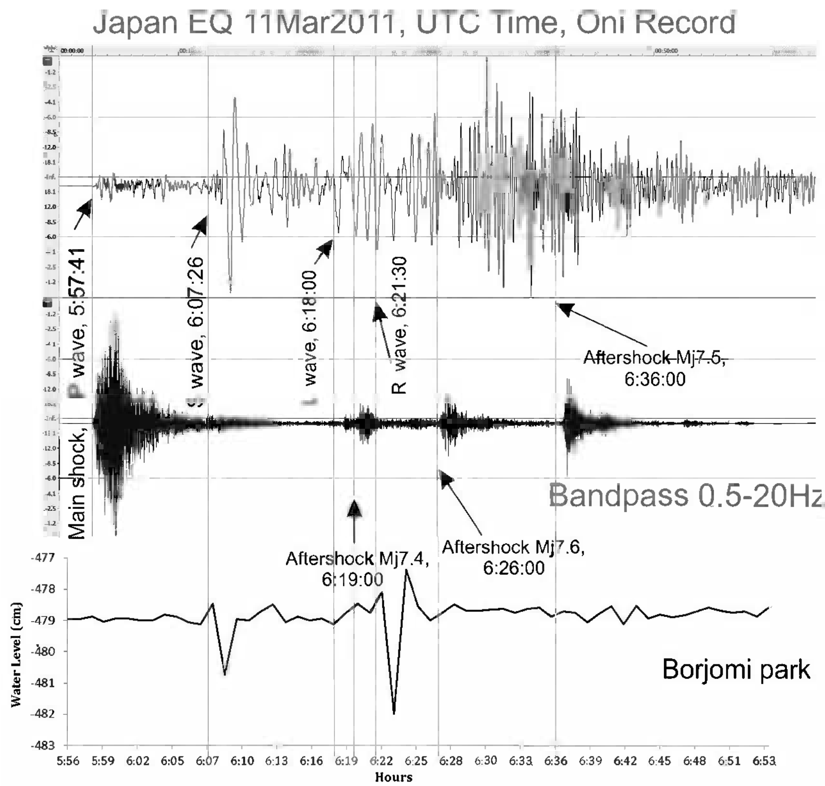

Fig. 7. The integrated plot of seismic and WL events in Georgia during Tohoku event. It is evident that the first strong WL perturbation at 06.07 correlates definitely with S-wave offset; no aftershocks are recorded at that time. The second strongest WL event between 06:19 and 06:22 coincides with both onset of L/R waves' 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 aftershock at 06.36 do not produce any characteristic WL oscillations; thus the most probable explanation of WL effect at 06:19 is the passage of L/R waves.

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_j7.4) 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 (Mj7.6) and 06.36 (Mj7.5) 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 L+R waves. Further, 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). We can conclude that there is good coincidence between teleseismic S waves onsets, some local tremor signals and hydroseismic anomalies. At the same time we cannot affirm that all seismic signal in the filtered record are definitely local tremors – some of them are most probably p-waves of strong aftershocks (Fig.7).

Finally, we conclude that teleseismic S and L+R waves of Tohoku EQ excite significant and quite identical WL anomalies on the whole territory of Georgia. In principle this means that corresponding pore pressure changes can excite DTT though the existing data do not allow making decisive conclusions.

3. Spectrum of WL oscillations following Tohoku EQ and mantle surface waves.

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 10th and 11th March is shown in Fig. 8.

After Tohoku EQ in the spectrum of WL oscillations appear several spikes around frequencies $2.5\ 10^{-3}$; $4.0\ 10^{-3}$; $4.9\ 10^{-3}$; $6.2\ 10^{-3}$; $7.2\ 10^{-3}$ Hz. Highest frequencies seem to be harmonics of the first mode ($2.5\ 10^{-3}$ Hz) 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 10th 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 (Tary et al, 2014). 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.

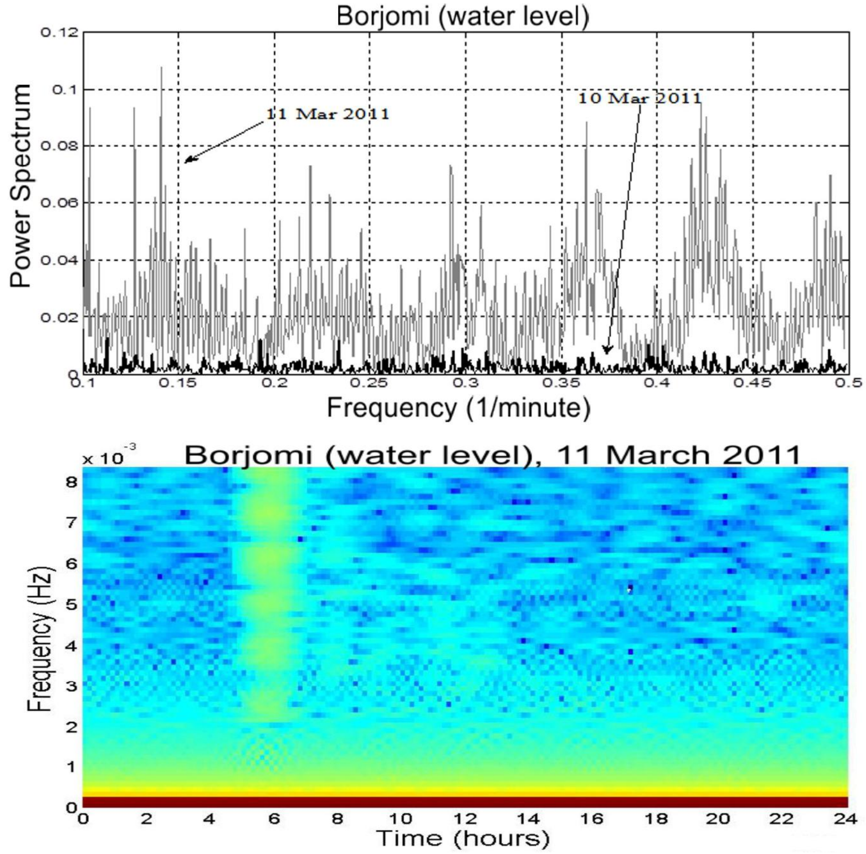

Fig.8. Spectrum (a) and spectrogram (b) of WL oscillations in Borjomi borehole before, during and after Tohoku EQ. The black curve in (a) is a background spectrum calculated for 10 March and grey curve - for 11March. The last one shows several strong spikes at (central) frequencies 2.5 10^{-3}; 4.0 10^{-3}; 4.9 10^{-3}; 6.2 10^{-3}; 7.5 10^{-3} Hz (periods 2-7 min), which are visible in the spectrogram (b) also and probably correspond to Rayleigh waves R1-R5.

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 (Tary et al, 2014). 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 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 (Bormann, 2012), which fits to the observed WL oscillations' frequencies:

410^{-3} to R5; 4.9 10^{-3} to R4; 6.2 10^{-3} to R3 and 7.5 10^{-3} Hz to R2 (Figs. 4, 8). These WL oscillation frequencies are compared to frequencies of Rayleigh waves in Table 2.

Table 2. Comparison of periods in WL oscillations with periods of multiple Rayleigh phases

Rayleigh phases	Periods of Rayleigh phases, s	Periods in WL oscillations, s

R2	110	133
R3	155	161
R4	185	200
R5	220	250

Taking into account wide distribution of observed WL oscillation periods (Fig.8 a, b), the dominant WL periods are close enough to these of Rayleigh phases.

We can conclude that our interpretation on coupling of WL events with multiple surface R-wave phases is confirmed by both good coincidence of WL signals and R-waves arrival times as well as by closeness of their frequencies' ranges.

4. Fusion of seismic and WL effects in Georgia related to Tohoku EQ

In the Table 3 the seismological and WL information on the Toholu EQ impact in Georgia is summed. Here and in the Table 3 $\Delta(WL)_{mR}$ and $\Delta(WL)_{mG}$ are the maximal WL signal (peak-to-peak amplitude of oscillations) for R-group waves and L/G-group waves correspondigly, cm; ΔP_{mG} and ΔP_{mR} are the maximal water pressure change during L/G-waves and R-wave passage, KPa; v_S, v_G and v_R are correspondingly the velocities of S, L/G and R waves in cm/s; $\Delta\sigma_G$ and $\Delta\sigma_R$ are the dynamic stress changes for L/G waves and R waves correspondingly, KPa; ΔL_S, ΔL_L and ΔL_R are accordingly displacements due to 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 in meters $\Delta(WL)_m$ to the particle velocity in the seismic waves v (or its proxy Peak Ground Velocity - PGV), $\chi = \Delta(WL)_m/v$ in units m/(m/s) (Brodsky et al, 2003).

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 (Chao, Peng et al, 2011) we measure the peak ground velocity for the Love and Rayleigh waves in the instrument-corrected NS and vertical component seismograms, respectively (Table 3). 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. These data allow calculating the amplification factor χ, which turns to be of the order of 80±10 m/(m/s). Interestingly, the calculation of the similar factor for tidal response χ_t results very low amplification value: $\chi_t \approx 3.10^{-6}$ m/(m/s) due to a low velocity of deformation.

The different WL responses in different boreholes to practically the same mechanical impact (11 KPa) is explained by the difference in aquifers' transmissivity/storage: large amplitudes of WL are favored by a high transmissivity/low storativity (Wang, Manga, 2010; Brodsky et al, 2003).

Table 3. Seismic and hydraulic reactions to Tohoku (M9) EQ in Georgia

Site name	$\Delta(WL)_{mR}$, cm	ΔP_{mR} KPa	v_S cm/s	ΔL_S cm					$\Delta\sigma$	

					v_L cm/s	ΔL_L, cm	v_R cm/s	ΔL_R, cm	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 (Wang, Manga, 2010). 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 the 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 (Bormann, 2012). Thus, we show that multiple surface R waves can generate not only local microseismicity, but also significant WL signals.

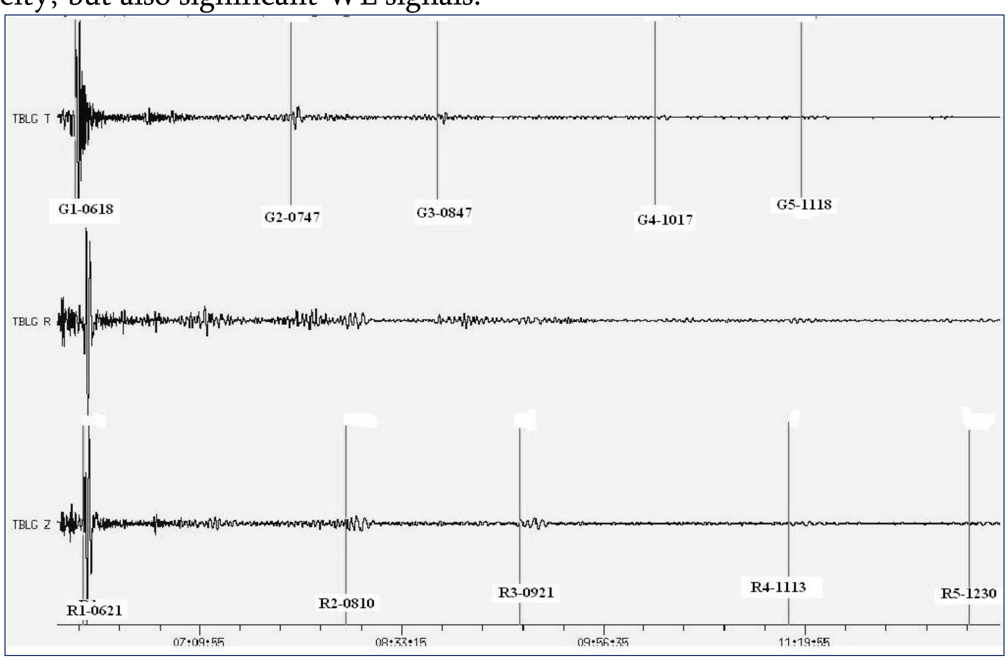

Fig.9. Seismogram with arrivals of multiple surface G and R waves at Tbilisi s/s.

On the other hand WL does not respond to the arrival of Love waves (G1, G2 etc – compare Figs.4, 9). 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, 9). Table 4 summarizes corresponding seismic and WL data.

Table.4. 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 m/(m/s)
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

We can conclude that though the stress change imparted by multiple surface waves of both G and R-groups are comparable (Table 4), 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; Rayleigh waves give rise to volumetric strain what satisfies this model (Wang, Manga, 2010). S and L waves have not volumetric component and accordingly they should not affect WL, but the recent data (Wang, Manga, 2010; Hill et al, 2013; Wang et al, 2009) 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 (Brodsky et al, 2003), permeability enhancement of fractured rocks due to removal of blocking elements by oscillating fluid (Wang, Manga, 2010) 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) document, for the first time, that multiple surface R waves generate not only local microseismicity (Peng et al, 2011), but also significant synchronous WL signals (unlike less efficient multiple surface G waves), see Figs. 4 and 8 (Chelidze et al, 2014).

Conclusions

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.

Thus in future the problem of discrimination of local tremors' signals from aftershocks should be analyzed in detail to avoid wrong interpretation.

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 WL perturbations, which document for the first time that passage of multiple surface Rayleigh waves: R2, R3, R4, R5 imparting dynamic stresses of the order of 0.5-2 KPa, also can affect WL regime. The amplification factor for S and L+R waves is of the order of 80.

Though teleseismic S and L+R waves of Tohoku EQ excite significant and quite identical WL anomalies on the whole territory of Georgia, which means that corresponding pore pressure changes can excite DTT the obtained data do not allow making decisive conclusions related to generation of local tremors by this event.

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.

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References.

Bormann, P. (Ed.) (2012). *New Manual of Seismological Observatory Practice* (NMSOP-2), IASPEI, GFZ German Research Centre for Geosciences, Potsdam; http://nmsop.gfz-potsdam.de;

Brodsky, E., Roeloffs, E., Woodcock, D., Gall, I., and Manga, M. (2003). A mechanism for sustained groundwater pressure changes induced by distant earthquakes. *J. Geophys. Res.* 108, B8, 2390, doi:10.1029/2002JB002321,

Chao, K., Z. Peng, C. Wu, C.-C. Tang, and C.-H. Lin (2012). Remote triggering of non-volcanic tremor around Taiwan, Geophys. J. Int. 188,no. 1, 301–324, doi: 10.1111/j.1365-246X.2011.05261.x.

Chao, K., Peng Z. Gonzalez-Huizar, H., Aiken, Ch., Enescu, B., Kao, H.,. Velasco, A.,

Obara, K. and Matsuzawa, T. (2013), A global Search for triggered tremor following the 2011 Mw9.0 Tohoku earthquake. *Bull. Seis. Soc. America*, 103, 1551-1571, doi: 10.1785/0120120171
Chelidze, T., Matcharashvili, T., Lursmanashvili, O., Varamashvili, N., Zhukova, N. and Meparidze, E. (2010). Triggering and Synchronization of Stick-Slip: Experiments on Spring-Slider System. In: *Synchronization and Triggering: from Fracture to Earthquake Processes*. Eds. V.de Rubeis, Z. Czechowski, R. Teisseyre, pp.123-164.
T. Chelidze and T. Matcharashvili. (2013). Triggering and Synchronization of Seismicity: Laboratory and Field Data - a Review. In: Earthquakes – Triggers, Environmental Impact and Potential Hazards. (Ed. K. Konstantinou), Nova Science Pub. Pp. 165- 231.
T. Chelidze and T. Matcharashvili, (2014). Dynamical Patterns in Seismology. In: Recurrence Quantification Analysis: Theory and best practices. (Eds. C.Webber and N.Marwan). Springer Cham Heidelberg, pp. 291-335.
Chelidze, T., Zhukova, N., Matcharashvili, T. 2014. SEISMOTOOL – an easy way to see, listen, analyze seismograms. *Seismol. Res. Letters*. 85, 889-895.
T. Chelidze, I. Shengelia, N. Zhukova, T. Matcharashvili, G. Melikadze, G. Kobzev. (2014). Coupling of Multiple Rayleigh Waves and Water Level Signals during 2011 Great Tohoku Earthquake observed in Georgia, Caucasus, *Bull. of the Georgian National Academy of Sciences.* v. 8, no. 2, 76- 79.
Costain, J. and Bollinger, J. (2010). Review: Research results in Hydroseismicity from 1987 to 2009. Bull. Seismol. Soc. Am. 100, 1841-1858, doi: 10.1785/0120090288.
Gonzalez-Huizar, H., Velasco, A., Peng, Zh. and Castro, R. (2012). Remote triggered seismicity caused by the 2011, M9.0 Tohoku-Oki, Japan earthquake. *Geophys. Res. Letters.*39, L10302, doi: 10.1029/2012GL051015.
Hill, D. and Prejean, S. (2009). Dynamic triggering. In: *Earthquake Seismology*. Ed. H. Kanamori. Elsevier. pp. 257-293.
Hill, D., Peng Zh., Shelly, D., and Ch. Aiken. (2013). S-Wave Triggering of Tremor beneath the Parkfield, California, Section of the San Andreas Fault by the 2011 Tohoku, Japan, Earthquake: Observations and Theory. *Bull. Seis. Soc. America*, 103, 1541–1550.
Obara, K. and Matsuzawa, T. (2013). A Global Search for Triggered Tremor Following the 2011 Mw 9.0 Tohoku Earthquake. *Bull. Seismol. Soc. Am.,* **103**, 1551–1571.
Parsons, T., Segou, M., Marzocchi, W., (2014). The global aftershock zone. *Tectonophysics.* 18, 1–3.
Peng, Z., Wu, C. and Aiken C. (2011). Delayed triggering of microearthquakes by multiple surface waves circling the Earth, *Geophys. Res. Lett.,* 38, L04306, doi:10.1029/2010GL046373.
Prejean, S. and Hill, D. (2009).Dynamic triggering of earthquakes. In: *Encyclopedia of Complexity and Systems Science*, R. A. Meyers (Ed.), Springer, pp. 2600-2621.
Reasenberg, P. (1985). Second-order moment of central California seismicity, 1969-1982. J. Geophys. Res., 90, 5479-5495.
Tary, J. B., van der Baan, M. and Eaton, D. W. (2014). Interpretation of resonance frequencies recorded during hydraulic fracturing treatments. *J. Geophys. Res. Solid Earth*, *119*. doi:10.1002/2013JB010904.doi:10.1002/2013JB010904.

Wang, C.-Y., Chia, Y, Wang, P-L., Dreger, D. (2009). Role of S waves and Love waves in coseismic permeability enhancement. *Geoph. Res. Lett.* 36, L09404, doi:10.1029/2009GL037330.
Wang, C.-Y., Manga, M. (2010). *Earthquakes and Water*. Springer-Verlag Berlin Heidelberg.
Zhang, Y. and Huang, F. (2011). Mechanism of Different Coseismic Water-Level Changes in Wells with similar Epicentral Distances in Intermediate Field. *Bull. Seis. Soc. America,* 101, 1531-1541.

Отклик на землетрясение Тохоку М9 в Грузии – локальные треморы и гидросейсмические эффекты

Тамаз Челидзе[1], Ия Шенгелия[2], Наталья Жукова[1], Темур Мачарашвили[1,2], Георгий Меликадзе[1], Геннадий Кобзев[1]

1. Институт Геофизики им. М. Нодиа;
2. 2. Университет им. Ильи Чавчавадзе

Резюме

В настоящее время имеется много данных о значительном влиянии сильных удаленных землетрясений на режимы подземных вод и локальной сейсмичности (т.н. невулканические или динамические треморы). Оба эти эффекта часто тесно связаны друг с другом, ибо одним из главных факторов, снижающим локальную прочность пород является поровое давление флюидов.

Некоторые свидетельства в пользу динамического триггерирования локальных треморов сильнейшим землетрясением Тохоку (М9) были получены недавно на Западном Кавказе. Кроме треморов, выявлены ясные аномалии в уровнях вод в скважинах при прохождении телесейсмических S-L-R волн, идентичные на всей территории Грузии от Боржоми до Кобулети. Мы полагаем, что совпадение предполагаемого сигнала от тремора с аномалией (осцилляцией) в уровнях вод в скважинах делает более надежной классификацию локального сейсмического сигнала как триггерированного явления. Обнаружены, видимо, впервые, заметные осцилляции в уровнях вод в скважинах при прохождении кратных поверхностных волн Релея.

დიდი ტოჰოკუს მიწისძვრის (M9) გამოძახილი საქართველოში - ლოკალური ტრემორები და ჰიდროსეისმური ეფექტები

თამაზ ჭელიძე[1],ია შენგელია[2], ნატალია ჟუკოვა[1], თემურ მაჭარაშვილი[1,2], გიორგი მელიქაძე[1], გენადი კობზევი[1]

1. მ.ნოდიას გეოფიზიკის ინსტიტუტი;
2. ილიას უნივერსიტეტი

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