Journal of Georgian Geophysical Society, Issue(A), Physics of Solid Earth, v.19A, 2016, pp. 49-57

Underground water level/temperatureresponse to seismic/tectonic transients: effects of poroelasticity

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Abstract

The problem of underground water flow is discussed in many publications. The upper crust down to approximately 20 km is a sequence of layers of porous/fractured fluid-saturated deformable rocks. When subjected to stress of various duration and mode, the excess transient pressure appears in the pore fluid, which can be observed practically as water level change/oscillations in boreholes.

In the paper are presented elementary theoretical models of underground water level/temperature response to seismic/tectonic transients taking into account effects of poroelasticity (Biot model). Besides, a direct mechanical squeezing of pore water by seismic vibrations (instantaneous flow) is considered.

The problem of underground water flow is discussed in many publications beginning from papers of Branchard and Bayerly (1935), Meinzer (1939), Jacob (1939), Biot (1941), which were followed by publications of Bredehoft (1967), Roeloffs (1988), Palciauskas and Domenico (1989), Brodsky et al (2003), Costain and Bollinger (2010).

Underground water levelresponse to seismic/tectonic transients. The upper crust down to approximately 20 km is a sequence of layers of porous/fractured fluid-saturated deformable rocks. When subjected to stress $\sigma$of various duration and mode, the excess transient pressure P_{ex} appears in the pore fluid. The fluid diffusion (or consolidation) equation defines the response of one-dimensional vertical flow to the appearance of P_{ex}(Domenico and Schwartz, 1994):

$$K_z \frac{d^2 P_{ex}}{dz^2} = \rho_w g(\Phi\beta_w + \beta_p)\frac{dP_{ex}}{dt} - \rho_w g\beta_w \frac{d\sigma}{dt} \qquad (1)$$

Here K_z is vertical hydraulic conductivity, ρ_w is density of water, g is acceleration due to gravity, Φ is porosity, $\beta_w$and β_p arecompressibilities of correspondingly water and pore space, σ is the total stress

$$\sigma = \sigma_{eff} + (P_s + P_{ex}) \qquad (2)$$

whereσ_{eff} is effective stress and P_s is the hydrostatic component of pressure. The coefficient $\rho_w g(\Phi\beta_w + \beta_p)$ is defined as a specific storage S_s and $\rho_w g\beta_p$ is a contribution to S_s, produced by pore compression.

Equation (1) states that the time rate of stress change defines the response of system to a given impact. There are two limiting cases: for stress rate much less than the characteristic time of diffusion of pore fluid, the excess pressure can exist only at small times; it dissipates quickly and the fluid is most of the time under constant pressure. As dissipation of excess pressure results from the rapid redistribution of fluid in the media of high permeability, this behavior is defined as a drained (relaxed) one. At fast loading, pore fluid has no time to escape; the mass of fluid in a given volume remains constant. This response is defined as undrained (unrelaxed). The undrained porous system is much stiffer as the part of load is supported by the uncompressible liquid. Accordingly, the fluid flow term in (1) d^2P_{ex}/dz^2 can be neglected.

Of course, elastic moduli of porous media differ considerably under drained and undrained conditions; in drained mode the porous system is more deformable.

If we convert the diffusion equation into a dimensionless form, a useful parameter, defining response mode of the aquifer can be obtained, i.e. the Fourier number N_{Fo}:

$$N_{Fo} = \frac{S_s L^2 / K}{t_e} = \frac{T^*}{t_e} \qquad (3)$$

Here $T^* = S_s L^2/K$ is the time constant of a given aquifer and t_e is the characteristic time of a transient (or observation time), L^2 is some characteristic length, in this case aquifer thickness.

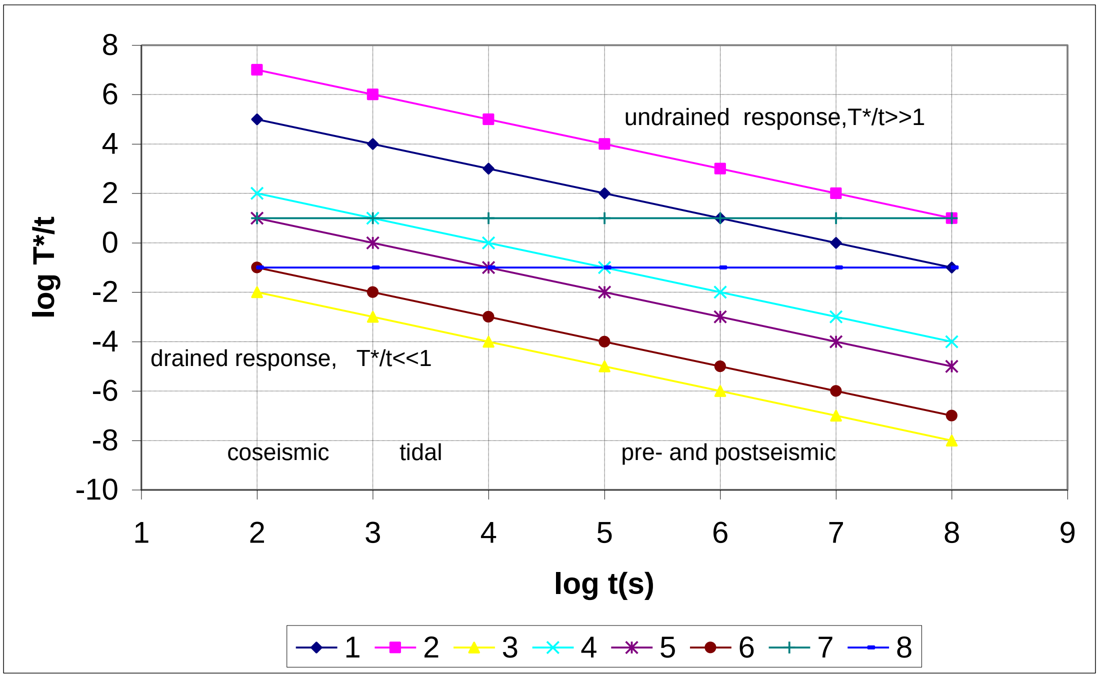

Fig.1. *Hydrodynamic Fourier number* $N_{Fo} = (S_sL^2/K)/t_e = T^*/t_e$ *versus characteristic time of transient (or observation interval)* t_e *for various* S_sL^2/K *ratios:*
1- $K = 10^{-8}$ *m/s;* $L^2S_s = 10^{-1}$*m; 2 -* $K = 10^{-10}$ *m/s;* $L^2S_s = 10^{-1}$*m; 3 -* $K = 10^{-8}$ *m/s;* $L^2S_s = 10^{-4}$*m;*
4 - $K = 10^{-4}$ *m/s;* $L^2S_s = 10^{-1}$*m; 5 -* $K = 10^{-2}$ *m/s;* $L^2S_s = 10^{-1}$*m; 6 -* $K = 10^{-4}$ *m/s;* $L^2S_s = 10^{-4}$*m;*

Large Fourier numbers N_{Fo} correspond to the undrained response; that means that all transients with $T^*>10t_e$ can be recorded. At small Fourier numbers $N_{Fo}<0.1$ or $T^*<10t_e$ the respond is drained, which means that transients, generated by fluid flow will be observable in a drained mode. So, depending on (local) aquifer properties the well can be selectively sensitive to co-seismic (t_e from 10 to 1000 s), tidal (t_e from 10^3 to 10^6 s), air pressure or pre- and post-seismic transient strain signals (t_e from 10^6 to 10^9 s and more).

The plot of “thermal” Fourier number N_{FoT} is identical with N_{Fo} at the condition $S_{sT}/c = S_s$.

All transients with t_e much less than T^* can be recorded by monitoring, but if the characteristic time of a basin T^* is much less than t_e , than the transient will not be observable, because the medium relaxes very fast.

It is evident that on the conceptual level time-dependence of hydraulic and microtemperature responses can be explained by the above theory. Therefore, a different response to coseismic (minutes), tidal (hours and days) or pre- and post-seismic (months and years) processes can be expected.

The plot of N_{Fo} versus characteristic time t_e for various values of S_sL^2, that is, for various time constants of a basin is shown in Fig. 1. The plot allows to predict the probable mode of response of a given aquifer to the impacts of various duration (Fig. 1). For highly permeable rocks ($K < 10^{-2}$ m/s), the coseismic impact (t_e of order of 10^2-10^3 s) N_{Fo} is very small, if the basin parameter S_sL^2 is of the order of 10^{-4}-10^{-5} m (which corresponds to L = 10 m, $S_s = 10^{-6}$ m). Because $N_{Fo}$is small we have to expect drained response in hydraulics and microtemperature, i.e. the coseismic signal will not be steadily observable.

The same can be said on the tidal impact (t_e of order of 10^{-5} s), as well as on pre- and post-seismic signals ($t_e = 10^{-6}$-10^{-7} s).

In rocks of a permeability of $K = 10^{-8}$ m/s, coseismic and tidal transients will be steadily observable, whereas slow pre- and post-seismic processes are observable in drained mode.

Only in tight rocks ($K = 10^{-10}$ m/s) all three effects are steadily observable in unrelaxed mode for $S_sL^2 = 10^{-4}$-10^{-5} m.

If we consider an aquifer with basin parameters of $S_sL^2 = 10^{-1}$ m (e.g. L =100 m, $S_s = 10^{-5}$m) the highly permeable rocks of $K = 10^{-2}$ m/s again seem to be unfavorable for observation of all three transients in undrained mode, but at $K = 10^{-8}$ m/s all of them are observable.

Values of hydraulic conductivity, favoring undrained response, seem to be too high, but in these calculations the porous formation was presumed as homogeneous. Real aquifer can be composed by layers of various hydraulic conductivity, or the porous space can be fractal (Sahimi, 1994); in these cases the channels of lowest permeability may limit the rate of (vertical) fluid flow.

It should be stressed that even in drained regimes the long pulse-like excitation can be observed at initial stage of response before pore fluid attains its drained equilibrium state.

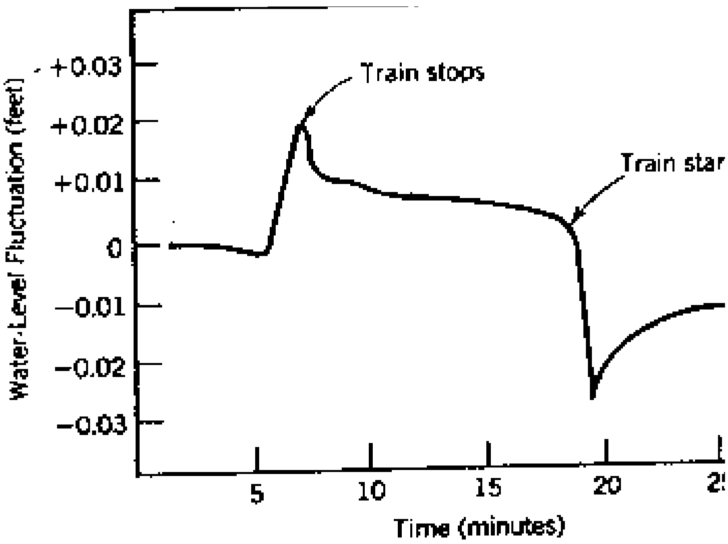

Fig. 2. Water level response to train passing close to a well

The spikes in water level immediately after arrival and departure of a train are transient undrained contributions to the response; the transients decay soon, giving way to the dominant drained contribution.

In the above analysis it was supposed that the deformation is reversible and the impact is transient, which means that the system regains its undrained elastic moduliafter distortion. If the mechanical impact lasts long enough like tectonic strain, even in reversible systems the largest portion of response reflects drained modules, which equals that the resulting "drained" strain should be maximal. Thus, in principle, the response to tectonic strain should be stronger than for short-living perturbation. In real systems, the strain time history can be more complicated(Fig.2); this can be connected with additional factors (saturation state, gas component, nonlinear response, etc.).

Underground water temperatureresponse to seismic/tectonic transients. The above analysis was related mainly to the hydraulic response. Therefore the question arises whether it is directly applicable to the thermal one? In relation to the atmospheric pressure the responses are very similar: both microtemperatures and water level are in a negative correlation with atmospheric pressure. Thus, the above analysis seems to be applicable at least qualitatively, to microtemperatures. A quantitative analysis should be founded on the solution of the heat conduction Equation with mass transfer, which is in the one-dimensional form:

$$\frac{k_e}{\rho' c'}\frac{\partial^2 T}{\partial z^2} - v_z \frac{\Phi \rho_w c_w}{\rho' c'}\frac{\partial T}{\partial z} = \frac{\partial T}{\partial t} \tag{4}$$

wherek_e is the effective thermal conductivity for porous system with moving water

$k_e = \Phi k_w + (1-\Phi)k_s + c v_z$

k_w and $k_s$are water and solid thermal conductivities, $\rho' c'$ is the effective heat capacity for the unit volume

$\rho' c' = n\rho_w c_w + (1-\Phi)\, \rho_s c_s$

c_w and$c_s$are water and solid specific heat capacities respectively, and v_z is the mean ground water velocity in z direction, c is proportionality factor. A simplest, linear dependence of k_e on v_z is assumed. As a rule, quantity v_z in Equ. (14) is considered as Darcy velocity (Domenico, Schwartz, 1994); combination of Darcy Equation with Equation (14) allows to model a thermal transient, which is connected with tectonic/seismic perturbations. It has to be noted that regional flow is a slow carrier of thermal perturbation; its velocity is 10^{-4}-10^{-5} m/s or less. It follows that this approach can be successfully applied to the modelling of the decay phase of co-, pre and post-seismic phase in the earthquake source area when the forced convection, generated by elastic pulse, comes to an end. The build-up (coseismic) phase of signals themselves call for another explanation, because their velocity is not less than 0.2 km/s. Presumably, the build-up phase of strain-related thermal signal results from some of elastic perturbation of medium by an elastic pulse, such as seismic wave, dislocation, plastic or deformation wave, Biot's slow wave, etc. The pulse disturbs the hydraulic and thermal stability of the borehole liquid through mixing of fluids with different temperatures. The analysis of this case can be founded on the solution of the heat equation when the source moves with velocity v_z much larger than the Darcy velocity of regional flow (forced convection).

A further model of fast heat transfer is based on analysis of losses caused by the friction on the solid-liquid interface in porous rocks. Friction generates heat during passing of seismic waves through the borehole area.

In both models the thermal anomaly appears "instantly" after the earthquake or generally after elastic displacement, but relaxation to the stable state will be slow, as the system is driven by conduction and slow convection only.

Dimensionless Eq.(14) reveals, like in the case of hydraulics, some controlling parameters of thermal process, i.e. the Peclet number

$$N_{pe} = (\Phi \rho_w c_w v_z L)/\kappa_e \tag{5}$$

(process is dominated by convection at $N_{Pe} >> 1$), and Fourier number for heat diffusion

$$N_{FoT} = (L^2 \rho' c' / \kappa_e)/t_e = (L^2 S_{sT}/k_e)/t_e = T_t^*/t_e \tag{6}$$

where$S_{sT} = \rho' c'$ can be defined as a "thermal specific storage". The effective thermal conductivity includes also the effect of fluid motion and in this case it is dominated by the convective term.

Plots, similar to Fig.3 can be drawn, using typical values of $\rho' c'$ for rocks. The hydraulic and thermal time response coincide at $S_{sT}/c = S_{sT}$.

Using this formalism we calculate N_{FoT} for various values of governing parameters c, v_z and L (Fig. 3), assuming typical values of thermal parameters of rock components:

c_w = 1 cal/g °C; ρ_w= 1g/cm^2; c_s = 0.2 cal/g °C; ρ_s = 2.5 g/cm^3 ; n = 0.1.

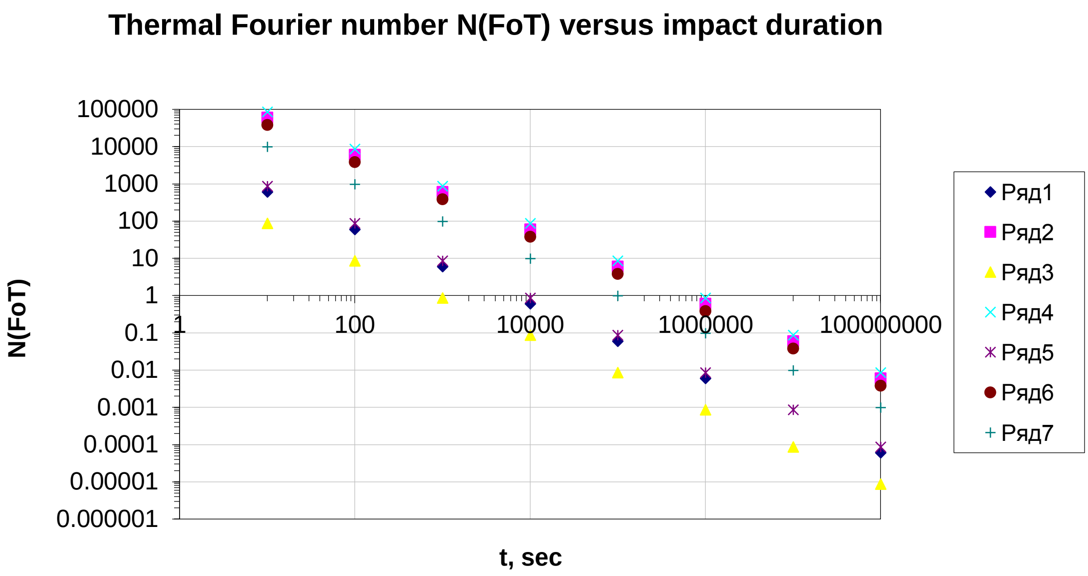

Fig. 3. Thermal Fourier number for heat transfer N_{FoT} versus characteristic time of thermal transient t_e :

1 – c = 10 cal/cm^2, v_z = 1 cm/s, L = 10^3 cm, λ = 170;

2 – c = 10 cal/cm^2, v_z = 1 cm/s, L = 10^4 cm, λ = 170;

3 – c = 10^{3}cal/cm^2, v_z = 1 cm/s, L = 10^3 cm, λ = 1164;

4 - c = 10^{3}cal/cm^2, v_z = 1 cm/s, L = 10^4 cm, λ = 1164;

5 - c = 10^{3}cal/cm^2, v_z = 10^{-3} cm/s, L = 10^4 cm, λ = 1164;

6 –c = 1.0 cal/cm^2, v = 100 cm/s, L = 10^4 cm, λ = 38 000;

7 – c = 1.0 cal/cm^2, v = 100 cm/s, L = 10^{3}cm, λ = 98.

The ratio $k_e/\rho' c'$ in (6) of units L^2/T is referred to as the thermal dispersivityλ ; it includes influence of moving fluid, which is so important in strain-related microthermometry. Thermal dispersivity reduces to thermal diffusivity, if the fluid is resting. Thermal Fourier number, like its hydraulic analogue, allows understanding the response of the formation to thermal transients. The conclusion can be drawn that thermal transients with L^2/λ<<1 are not observable; in other words, observation ofshort thermal transients is favored by large thickness of aquifer and large velocity of moving water.

It seems that in principle hydraulic and thermal decay phases can differ significantly due to inertia of thermal perturbation. If in the build-up phase the thermal relaxation time of the system is governed by thermal dispersion, in the decay phase, at v_z = 0, the heat is dissipated by thermal diffusion. Thus, thermal anomaly can persist even when the pressure transient has gone.

Hydro-seismic response to tides and wave-trains of remote strong earthquakes. The tidal response of water level (WL) in boreholesis well known; its intensity depends mainly on the aquifer properties. The strain-sensitivity of a given well is calculated as ratio of WL change to tidal impact, expressed in units of displacements or stresses (Table 1).

Generally, the reaction of the underground water (Wang et al, 2009; Zhang, Huang, 2011; Wang, Manga, 2010), WL responds 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; 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 close to instantaneous, it is expected that pore water has no time to displace, which in turn means that the WL response is undrained (Wang, Manga, 2010).

Below we present the observation data on hydro-seismic response to tidal strains and wave-trains of remote strong earthquakes recorded by the hydrogeodynamic network of deep wells of the M. Nodia Institute of Geophysics.

Table. 1. Reactions to tidal strain on the water level WL the various boreholes of network.

Name	water level (kPa)	tidal (kPa)	Response (%)	Date
Axalakalki			no	
Ajameti	0.66841	3.6247	18.4	22.01.2016
Kobuleti	0.47743	3.4204	14.0	06.02.2016
Lagodexi	0.46481	3.9388	11.8	11.01.2016
Marneuli	0.71401	3.9549	18.1	11.01.2016
Oni			no	
Gori	0.30677	3.9228	7.8	11.01.2016
Nakalakevi	0.51188	3.9223	13.1	11.01.2016
Borjomi70	1.5125	3.6395	41.6	19.08.2013
BorjomiPark	0.1962	3.3492	5.9	22.08.2013
Borjomi47	1.1473	3.5874	32.0	29.05.2014

Responseis calculated as a ratio of WL stress (KPa) to the theoretical value of the tidal stress (KPa) in percents. “no”means absence of tidal reaction. The ratio is time dependent, so the calculated values are characteristic for corresponding dates.

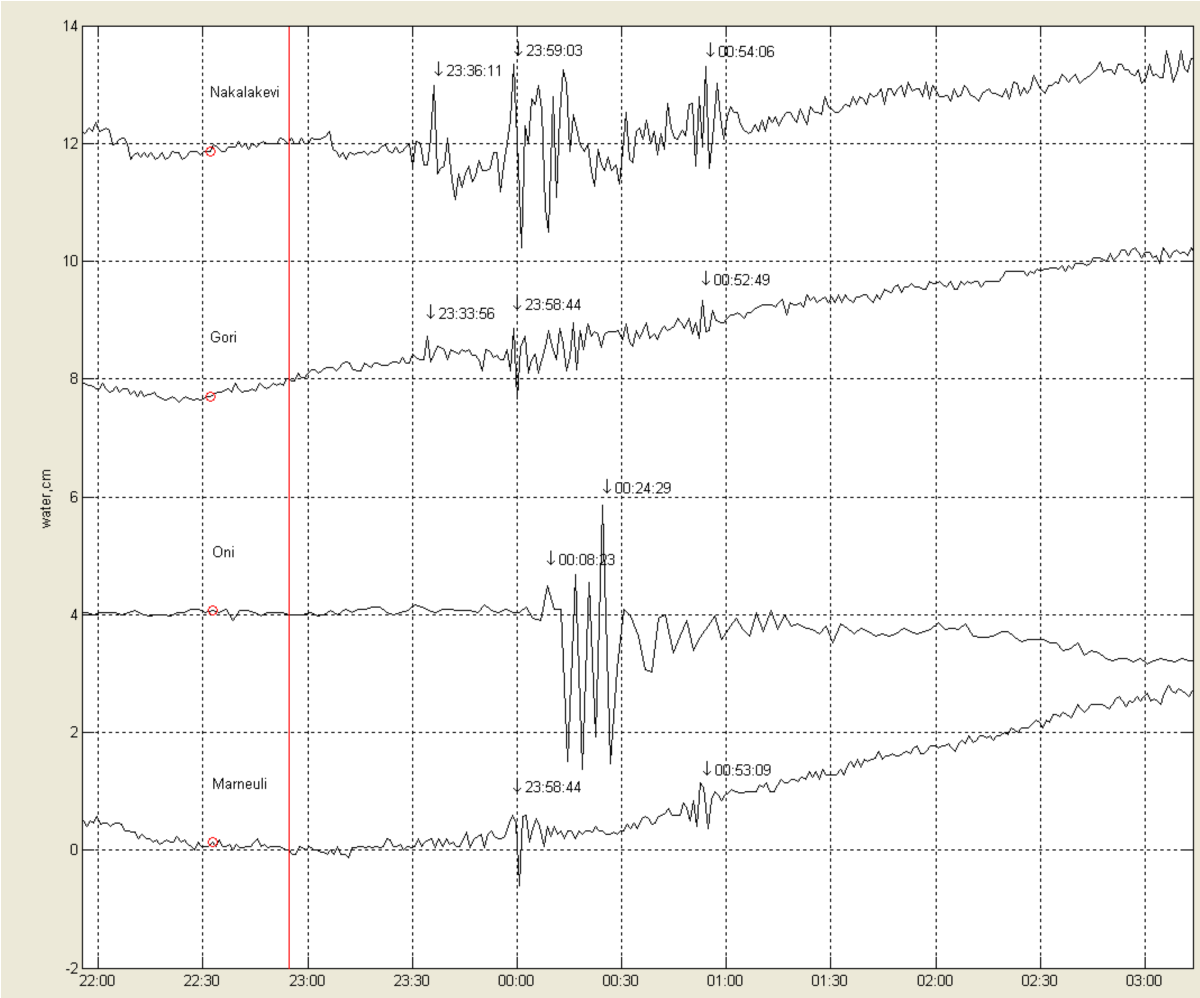

Fig. 4. WL oscillations due to passage of wave-trains of the Chile earthquake 16-09-2015,time:22:54,M8.3.Reactions successively from top to bottom in: Nakalakevi, Gori, Oni,Marneuli.

Fig. 4 shows that WL in the network of wells responses to the passage of seismic waves (here – mainly to direct and multiple Rayleigh waves). This seems to contradict the common knowledge that due to relatively short duration of seismic oscillation the groundwater regime should be undrained, i.e. the WL in the borehole should not change during seismic wave passage: the data of Fig.4 testify against this model. The explanation of synchronous oscillation of WL with fast seismic vibrations should take into consideration, that even in the undrained regime small-scale water displacement in the wellis observed (Fig.4). This effectis connected not with a pore water diffusion in the aquifer (slow flow, Eq.1), but with direct mechanical squeezing of fluid from the porous media, surrounding the well,by the passing seismic wave (Dvorkin, Nur, 1993): this we can call instantaneous flow. In this case again *p*-waves, though compressive, are less effective in generation of WL oscillations. It seems that the most effective impact on the WL manifest Raileigh waves; then follow Love and *s*-waves. Thus, in the instantaneous flow also should be some time constant of the order of dozens of sec for mechanical vibrations to be effective generators of WL oscillations.

Conclusions

We considered theoretical models of underground water level/temperatureresponse to seismic/tectonic transients taking into account effects of poroelasticity (Biot model): we show that the response intensity depends on the Hydrodynamic Fourier number in case of water level change and on the Thermal Fourier number in case of temperature response. Calculations manifest that response depends strongly on the characteristic time of hydraulic/thermal transient.

Besides, a direct mechanical squeezing of pore water by seismic vibrations (instantaneous flow) is considered.

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მიწისქვეშა წყლების დონის/ტემპერატურის რეაქცია სეისმურ და ტექტონიკურ გარდამავალ ზემოქმედებაზე: ფოროელასტიურობის ეფექტები.

თამაზ ჭელიძე

რეზიუმე

ზედა ქერქი დაახლოებით 20 კმ სიღრმემდე წარმნოადგენს ფოროვან ან დაბზარულ ფლუიდით გაჯერებულ დეფორმადი ფენების მიმდევრობას. სხვადასხვა ხანდრმლივობის და ინტენსივობის გარდამავალი დაძაბულობები იწვევს ამ სისტემაში

ფლუიდის ჭარბ ასევე გარდამავალ ფოროვან წნევას, რაც პრაქტიკულად დაიკვირტვება ჭაბურღილებში წყლის დონის ცვლილებასა და ოსცილაციებში. სტატიაში მოცემულია მიწიქვეშა წყლების სეისმურ/ტექტონიკურ გარდამავალ ზემოქმედებებზე რეაქციის ელემენტარული თეორიული მოდელი ფოროელასტიურობის ეფექტების გათვალისწინებით (ბიო-ს მოდელი). გარადა ამისა, განხილულია ფოროვანი სითხის პირდაპირი მექანიკური გამოდევნება სეისმური ვიბრაციებით (წამიერი დინება).

РЕАКЦИЯ УРОВНЯ ПОДЗЕМНЫХ ВОД /ТЕМПЕРАТУРЫ НА ПЕРЕХОДНЫЕ СЕЙСМИЧЕСКИЕ И ТЕКТОНИЧЕСКИЕ ВОЗДЕЙСТВИЯ: ПОРОУПРУГИЕ ЭФФЕКТЫ

ЧЕЛИДЗЕ Т.Л.

Реферат

Верхняя кора приблизительно до глубины 20 км представляет собою последовательность пористых или трещиноватых, подвергнутых деформации пород, насыщенных флюидом. Переходные напряженности разной длительности и интенсивности в этой системе вызывают избыточное, также переходное, поровое давление флюида, что, пректически, наблюдается в скважинах, как изменение уровня воды и ее осциляциях. В статье дана элементарная теоретическая модель реакции на сейсмические/ тектонические переходные воздействия на состояние подземных вод с учетом эффекта пороупругости (Модель Био). Кроме того, рассмотрено прямое механическое выжатие пористой жидкости сейсмическими выбрациями (Мгновенное течение).