ISSN 0742-0463, Journal of Volcanology and Seismology, 2012, Vol. 6, No. 5, pp. 316–328. © Pleiades Publishing, Ltd., 2012.
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On the Relationships of Water-Level Variations in the E-1 Well, Kamchatka to the 2008–2009 Resumption of Activity on Koryakskii Volcano and to Large (M ≥ 5) Earthquakes

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Received January 25, 2012

Abstract—We discuss the water-level variations in the E-1 well for the time period between May 2006 and 2010, inclusive. A trend towards an increasing level at an abnormally high rate occurred from mid-2006 to December 2009. This increase is regarded as the response of the aquifer of gas-saturated ground water that exists in the volcanogenic–sedimentary deposits of the Avacha volcano-tectonic depression to volumetric compression strain changes during the precursory period and the occurrence of a swarm of small earthquakes ($K_{S_{\max}} = 8.3$) in the area of Koryakskii Volcano and to its phreatic eruption. We estimated the volumetric compression as $\Delta\varepsilon = -(4.1 \times 10^{-6}–1.5 \times 10^{-5})$ from the amplitude of water-level rise using the elastic parameters of the water-saturated rocks. While the strain source was active, we observed a decreasing sensitivity of the hydrodynamic regime in the well to the precursory processes before large (M ≥ 5.0) tectonic earthquakes.

DOI: 10.1134/S074204631205003X

INTRODUCTION

Monitoring active volcanoes in order to predict their eruptions relies on seismological, geodetic, and a set of remote-sensing techniques. In some cases, certain information for eruption prediction can be extracted from observations of the chemical composition, temperature, and heat power of fumaroles. Data from water-level and other kinds of hydrogeological observation are used much less frequently, because observation wells and springs are usually far from the volcanoes of interest and because little knowledge is available at present as to how volcano-tectonic processes affect the hydrologic and geochemical regimes of ground water during the precursory periods of eruptions.

Water-level observation in piezometric wells is considered to be an effective tool for identifying signals of current geodynamic activity, in particular, hydrogeodynamic precursors of large earthquakes. Signals of seismotectonic strain in water-level changes can be entirely quenched or may occur in a distorted form because of the inertia of interaction between ground water in water-saturated rocks and in the wellbore [Boldina and Kopylova, 2006; Hsieh et al., 1987; Rojstaczer, 1988], as well as because of the possible actions of hydrogeodynamic factors that are related to variations in the amount of ground water and the boundary conditions around the well.

The range of periods in the manifestations of inertial water exchange in different wells may vary between a few minutes or a few hours and a few days and is controlled by the isolation of the controlled aquifer and the filtration and elastic parameters of the water-saturated rocks, as well as by how good the connection is between the water-saturated rocks and the wellbore. If there is no influence of hydrogeodynamic factors on water-level changes in the observation well, then the effect of inertial water exchange can be disregarded if the evolution of seismotectonic strain takes at least a day to some tens of days and the water-level changes can be assumed to be directly related to changes in ground water pressure due to deformation of the water-saturated rocks. In that case, one can obtain quantitative estimates of the seismotectonic strain from identified amplitudes of anomalous water-level variations taking the deformation properties of the observation well into account. These deformation properties of a well are controlled by its statically isolated response of the water level to barometric, tidal, and seismotectonic excitation and by the range of periods in which it occurs, as well as by the strain sensitivity of the water level [Kopylova, 2006, 2009].

The monitoring of present-day seismotectonic processes is best done using deep wells that penetrate isolated, dense, water-saturated rocks at depths of at least 200 m [Kopylova et al., 2007]. Such wells also show comparatively low seasonal variations in water level or none at all. The E-1 well in southern Kam-

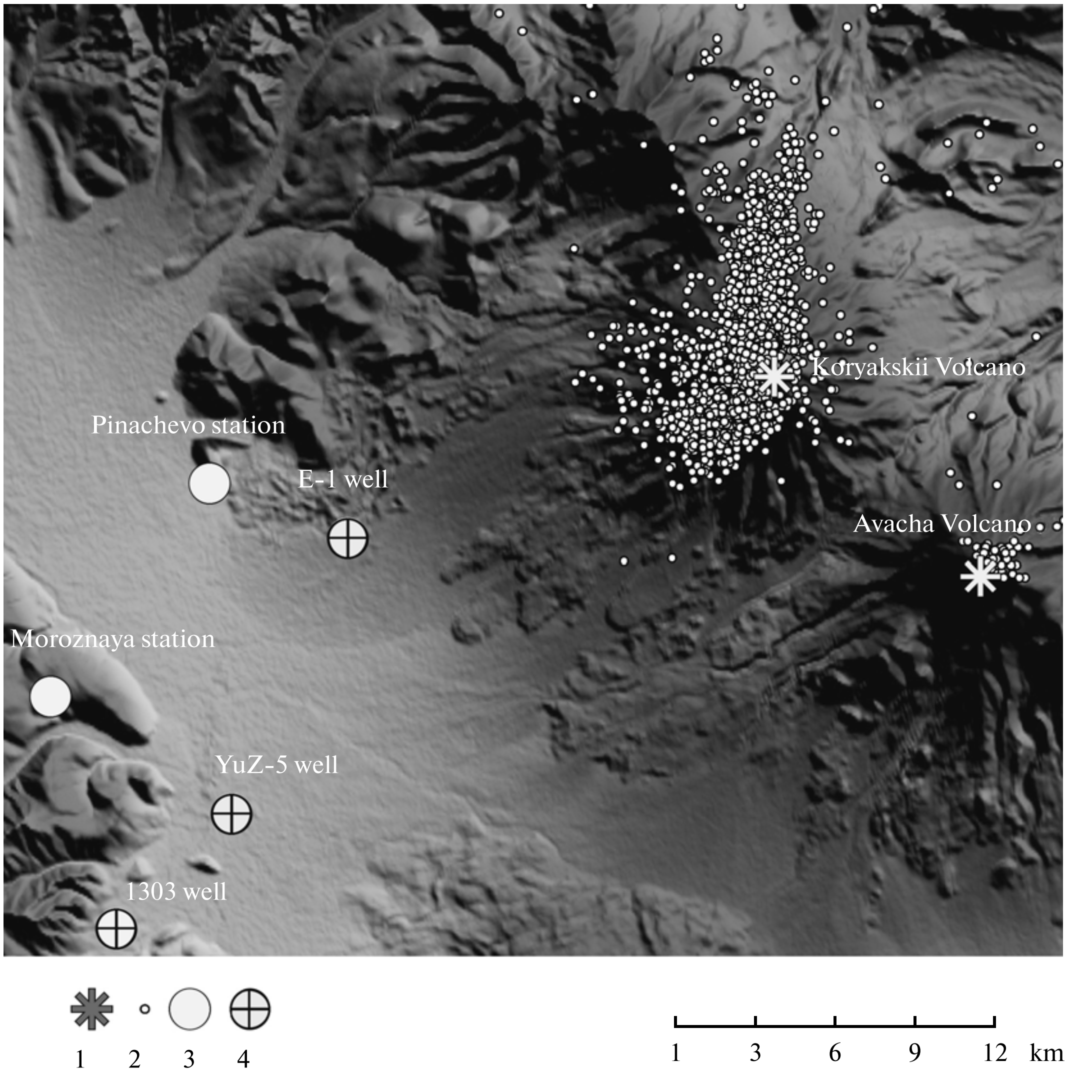

Fig. 1. A map of the Avacha volcano-tectonic depression and its circumference. (1) active volcanoes, (2) epicenters of earthquake swarm in the area of Koryakskii Volcano for 2008–2010, $K_S = 3.1–8.3$, (3) sampling sites to monitor ground water chemical composition, (4) observation wells.

chatka, which is 17 km from the summit of the active Koryakskii Volcano (Fig. 1), is such a well.

An ascending trend of water-level increase at an anomalously high rate was observed at the E-1 well from mid-2006 to December 2009. This increase began 1.9 years before the start of an earthquake swarm ($K_{S_{max}} = 8.3$) within the nearly north–south zone that contains the Koryakskii edifice (see Fig. 1) and 2.5 years before its fumarole activity began to increase. The water-level increase ceased approximately at the same time as the termination of the earthquake swarm and the volcano's fumarole activity. This anomalous water-level variation lasted about 3.5 years; the amplitude of the water-level rise was 1.22 m.

Previously, the data of long-continued observations (from 1987 until the present) were used to find the sensitivity of the hydrodynamic regime at the E-1 well to the precursory process of $M \geq 5.0$ tectonic earthquakes that are produced during the subduction of the Pacific oceanic plate under the Sea-of-Okhotsk plate (of the continental type) [Kopylova, 2001, 2008]. Water-level drops with increased rates during a few weeks to a few months were recorded before about 70% of all such earthquakes at hypocentral distances $R < 350$ km from the well. We observed a lower sensitivity of the well to the precursory processes of tectonic earthquakes during the 2006–2009 intensive water-level increase. Since 2010, i.e., after the water level stopped rising, the E-1 well recovered its sensitivity to the precursory processes of $M \geq 5.0$ earthquakes.

This paper considers the hydrodynamic regime of the E-1 well in 2005–2010 in conjunction with the evolution of the earthquake swarm that occurred in the

Koryakskii Volcano area and with its fumarole activity, as well as $M \geq 5.0$ tectonic subduction earthquakes. The amplitude of the net water-level rise in combination with the estimated elastic parameters of water-saturated rocks was used to assess the volumetric compression strain in the E-1 area during the period prior to and simultaneous with the seismic and fumarole activity increase on Koryakskii Volcano. We discuss possible mechanisms that might be responsible for the source of volumetric compression strain in the area.

The data we are using on the seismic and fumarole activities in the area of Koryakskii Volcano were acquired by the Laboratory of Seismic and Volcanic Activity Research at the Kamchatka Branch of the Geophysical Service (Russian Academy of Sciences), hereafter referred to as KB GS RAS.

DESCRIPTION OF THE STUDY AREA AND ACTIVITY INCREASE ON KORYAKSKII VOLCANO

Koryakskii Volcano (the summit coordinates are 53.32° N, 158.72° E, height 3456 m) is one of the two active volcanoes in the Avacha–Koryakskii volcanic cluster, which is situated within the Avacha volcano-tectonic depression (see Fig. 1). Koryakskii showed increased activity in 2008–2009 in the form of higher seismicity and fumarole steam-and-gas activity [Gordeev et al., 2009; Seliverstov, 2009; Senyukov and Nuzhdina, 2010].

Figure 1 shows the epicenters of K_S 3.1–8.3 earthquakes that occurred from March 2008 to December 2009, which were recorded by seismic stations operated by the KB GS RAS. The earthquake hypocenters were in the depth range between 20 km and the ground surface and also occurred in the volcano's edifice itself [Senyukov and Nuzhdina, 2010]. Most of these earthquakes were concentrated at depths of 5–9 km within a nearly north–south strip 11 km long and 4–5 km wide. Seismicity maxima were recorded in March and October 2008 and in April and August 2009. During those months 600–1000 events with $K_S = 3.1–8.3$ occurred, with the average monthly total of such earthquakes being between 0 and 130 events. The seismic energy that was released within the north–south strip (see Fig. 1) during the period from March 2008 to December 2009 was 5.4×10^8 J.

Increased fumarole activity on Koryakskii Volcano occurred from November 2008 to August 2009 as observed by the KB GS RAS and by the Institute of Volcanology and Seismology of the Far East Branch of the Russian Academy of Sciences and was accompanied by powerful steam-and-gas plumes (Figs. 2a–2c). Resurgent ash was present on some days in the steam-and-gas discharges and plumes. According to thermal visual observations [Gordeev and Droznin, 2010; Droznin and Dubrovskaya, 2010] over 250 days (from the end of December 2008 to the end of August 2009) the maximum temperature of the steam-and-gas discharges was 430°C and the average discharge rate of overheated steam was 35 kg/s. The mass of the steam that came to the ground surface during the eruption was 10^6 t and the total energy of the steam-and-gas discharge was 3×10^9 J.

Figure 1 shows the sites of ground-water regime observation, including the E-1 and YuZ-5 piezometric wells, which are operated by the KB GS RAS, and well 1303 (of the OAO Kamchatgeologiya) where water level and air pressure are measured at intervals of 10 min. Information on the wells can be found in Table 1. Figure 1 also shows two sites where the KB GS RAS observes the chemical composition of ground water. No anomalous changes were recorded in the parameters of the ground water at the observation wells and springs in connection with the activity increase on Koryakskii Volcano, except for the E-1 well.

A DESCRIPTION OF THE E-1 WELL AND WATER-LEVEL OBSERVATIONS

The E-1 well is 17 km from the summit of Koryakskii Volcano and 9 km from its base. The well reached Alnean (Neogene) tuffs with low water content in the 625–647 m depth range (see Table 1 and Fig. 3). A pumping test of the well in 1991 showed that the transmissivity of the water-saturated rocks is $T = 0.005\ \mathrm{m^2/day}$.

Sampling in the 625–647 m range during the drilling in 1984 gave the chemical composition of ground water as chloride–carbonate sodium–calcium, corresponding to the formula $M_{2.5}(CO_3^{2-}78Cl^{-}22)/(Ca^{2+}56Na^{+}40)$, where M is the mineral content in g per liter; the concentrations of anions and cations are given in % equivalent, pH = 12.4. The chemical analysis was carried out at the Central Chemical Laboratory of the Institute of Volcanology of the FEB RAS.

Based on an analysis of the water samples taken in 1998–2011 in the upper part of the water column in the wellbore, the water had a carbonate–chloride sodium composition consistent with the formula

$$M_{1.0-1.5}\left[Cl^{-}65-40\left(CO_3^{2-}+HCO_3^{-}\right)45-33\right.$$
$$\left.SO_4^{2-}6-1\right]\Big/\left(Na^{+}91-78Ca^{2+}11-1K^{+}7-8\right),$$
$$pH=11.5-12.1.$$

The chemical analyses of the water were carried out at the Laboratory of Hydroseismology of the KB GS RAS.

The chemical composition of the gas dissolved in water in the region of Alnean tuffs at depths of 570–1060 m was determined during drilling in the 910–920 m range: 60 vol % for N_2, 29 vol % for CH_4, 10 vol % for O_2, 0.6 vol % for Ar, 0.3 vol % for CO_2, and 0.2 vol % for H_2. Heavy hydrocarbons (from C_2H_4 to C_6H_{14}) were 0.02 vol % and He was 0.006 vol % (as

Fig. 2. Variation in steam-and-gas activity on Koryakskii Volcano in 2009–2011: (a) January 10, 2009, intensive steam-and-gas activity, photographed by D.V. Mel'nikov; (b) April 9, 2009, intensive steam-and-gas activity with ash involved, photographed by S.V. Ushakov; (c) August 27, 2009, intensive steam-and-gas activity with ash involved, photographed by O.A. Girina; (d) October 19, 2009, steam-and-gas activity, photographed by S.V. Ushakov; (e) January 30, 2011, quiet condition, photographed by A.A. Nuzhdaev.

Table 1. The characteristics of observation wells

Well, abs. alt., m	Coordinates, deg		Well depth/water level, m	Wellbore radius in the range of water level fluctuation r_c/ wellbore radius in the range of hole filter r_w, m	Depths range of the open wellbore/ length open part of the wellbore, m	Technical specifications for well—aquifer connection	Water-saturated rocks, age, composition	Pore fluid	Transmissivity T, m^2/day
	N	E							
E-1, 180	53.26	158.48	665/28	0.109/0.109	625–647/22	perforated filter	N, tuffs	water + gas N_2–CH_4	0.005
YuZ-5, 70	53.17	158.41	800/1.5	0.122/0.084	310–800/490	open wellbore	K_2, aleurolites, argillites	water	7.8
1303, 31	53.08	158.21	800/25	0.122/0.046	517–717/200	open wellbore	N, tuff sandstone, tuff breccia, tuff aleurolite	water	0.32

reported by the Central Chemical Laboratory of the Institute of Volcanology, FEB RAS). This composition of dissolved gas shows excessive concentrations of methane and other hydrocarbon gases compared with the gas that is dissolved in water, which is in equilibrium with atmospheric air. This provides evidence for the generation or transport of hydrocarbon–nitrogen gases in Alnean volcanic rocks in the E-1 well area. Later, Pozdeev [2003] demonstrated (using data on the gas composition of deep wells, including E-1 and GK-1P wells at the Pinachevo station, see Fig. 1) that the interiors of the Avacha volcano-tectonic depression concentrate hydrocarbon-bearing gases of Early Cretaceous to Pleistocene ages within the same areas and wells. This reveals the fact that local hydrogeologic structures are sufficiently closed in the interiors of the Avacha depression and that gas can migrate vertically.

Since 1987 we have been conducting water-level observations at the well in order to find hydrogeodynamic precursors of earthquakes [Kopylova, 2001]. In 2002 digital Kedr A2 equipment (manufactured by the Polinom Ltd., Khabarovsk) was installed that has an ultrasound water-level sensor (with a sensitivity of ±0.01 cm) and an atmospheric-pressure sensor (with a sensitivity of ±0.2 hPa). Synchronous measurements of water level and air pressure are made at intervals of 10 min. This observational arrangement and the data-processing technique we are using provide for analysis of water-level variations with amplitudes ≥0.01 cm at periods of a few days [Kopylova, 2001, 2008].

One characteristic feature of the natural hydrogeodynamic regime at the well consists in gradual increases and decreases in water level that last 3–6 years with amplitudes of a few tens of centimeters to 1.5 m and at average rates of below 0.1 cm/day. Against the background of such trends one can see weak barometric water-level variations and water-level changes that are related to large tectonic earthquakes.

In contrast to other observation wells, the water-level changes at E-1 do not contain either tidal variations or annual seasonal variations (see Fig. 1, Table 1). These features in the hydrodynamic regime of the well are due to the fact that the well controls the ground water in a hydrodynamic zone of poor water exchange, as well as being due to methane–nitrogen gas that is present in the ground water of the water-saturated rock and in the wellbore. This is indicated by chemical analyses of dissolved gas sampled in the 910–920 m range (see above) and the presence of gas bubbles that accumulate on the walls of bottles that contain water samples taken from the E-1 wellbore. Our observations convince us that when water is sampled from wells that penetrate into ground waters with dissolved gas that is in equilibrium with atmospheric air, no gas bubbles form on bottle walls.

The presence of gas in the ground water that is in the Alnean volcanic rocks and in the water column of the E-1 wellbore is accompanied by increased compressibility of the fluid phase compared with that of ordinary fresh water. This increased compressibility of ground water may be one of the more important factors why no tidal variations are observed and why the barometric response in the E-1 water-level changes is weaker [Kopylova, 2009].

Table 2 lists the barometric efficiency and tidal sensitivity of the water level at observation wells as estimated from barometric and tidal analyses of mean hourly variations in water level and air pressure. For the E-1 well the barometric efficiency in the range of daily periods is rather small ($E_b = -0.01$ cm/hPa) and reaches the value $E_b = -0.1$ cm/hPa at periods of a few tens to hundreds of days [Kopylova, 2009]. For comparison purposes, we quote the values of barometric efficiency at YuZ-5 and 1303 wells, which are –0.40 and –0.43 cm/hPa, respectively, at periods of a few hours to a few tens or hundreds of days. One signifi-

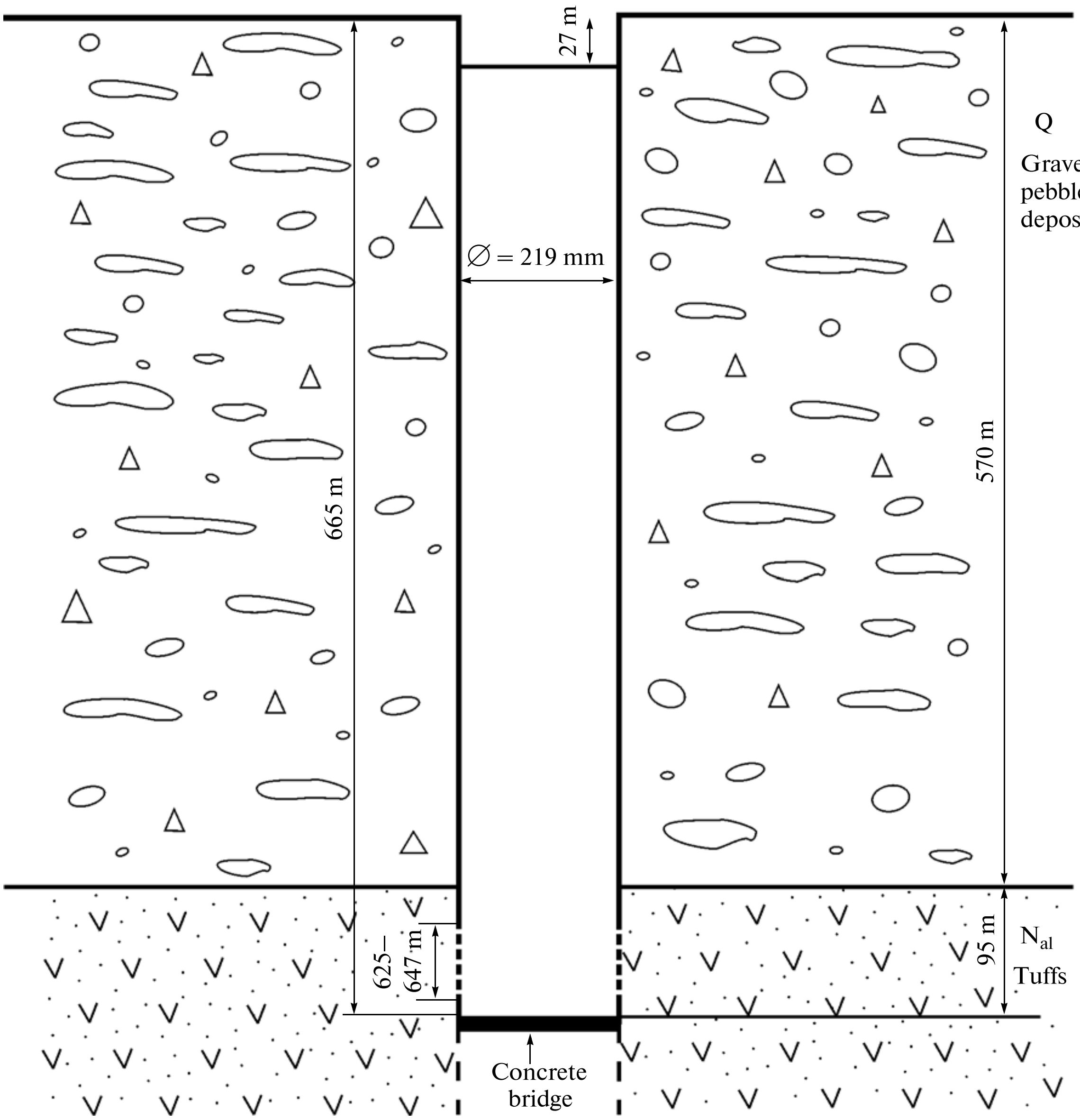

Fig. 3. The structure of the E-1 well and the associated geological section.

cant difference of these wells from E-1 is that they are situated within positive hydrogeologic structures and control the ground water in the zone of active water exchange; as well, there is no gas generation in the water-saturated rocks that they penetrate.

One unique feature of the hydrodynamic regime at E-1 consists in the fact that the hydrodynamic precursor is observed in the form of an increased rate of water-level drop during a few weeks to a few months prior to $M \geq 5$ earthquakes at distances of 350 km or shorter [Kopylova, 2001, 2008]. The 1997–2007 observations showed that precursors could be observed before 70% of such earthquakes [Kopylova, 2008]. The retrospective prediction efficiency after [Gusev, 1974] is 2–3, thus showing that this precursor can improve earthquake prediction by factors of 2–3 compared with the random guess procedure.

Since mid-2006 the E-1 well showed a trend of increasing water level at an anomalously high rate, as high as 0.15–0.20 cm/day (Fig. 4). This water-level increase showed an increase in pore pressure due to the action of a volumetric compression strain source, which might be caused by the precursory period and occurrence of an earthquake swarm and eruption on Koryakskii Volcano. Below, we discuss how the volumetric compression strain can be estimated around the well and consider the origin and location of the strain source, as well as the influence such a source might exert on the appearance of hydrogeodynamic precursors of large tectonic earthquakes.

Table 2. The elastic parameters of ground-water aquifers penetrated by observation wells

Well	Barometric efficiency E_b, cm/hPa	Tidal sensitivity A_S/A_V, m/10^{-7}	Drained skeleton compressibility β, $Pa^{-1} \times 10^{-11}$	Shear modulus G, $Pa \times 10^{10}$	Skempton's coefficient B	Porosity ϕ	Storage of the reservoir S
E-1	0.01/0.1[1]	0.010/0.015	7.59	0.79	0.044–0.17	0.01–0.06	5.8×10^{-5}
1303	0.43	0.143/0.215	7.37	0.81	0.64	0.06	19.6×10^{-5}
YuZ-5	0.40	0.107/0.161	12.5	1.34	0.67	0.11	16.9×10^{-5}

Note: [1] The first numeral gives an estimate of barometric efficiency for the daily range of periods, the second gives an estimate for the range of periods between a few tens and a few hundreds of days [Kopylova, 2009].

ESTIMATING THE VOLUMETRIC STRAIN IN THE E-1 AREA

We estimated the volumetric compression using a relation that connects changes in pore pressure Δp and volumetric strain $\Delta\varepsilon$ for statically isolated conditions, that is, where the water flow in water-saturated rocks and the water exchange between the water-saturated rock and the wellbore can be neglected [Roeloffs, 1988]:

$$\Delta p = -(2GB/3)[(1 + \nu_u)/(1 - 2\nu_u)]\Delta\varepsilon. \quad (1)$$

This relation is controlled by the elastic parameters of the aquifer: Poisson's ratio for zero water runoff (undrained conditions) ν_u, the shear modulus G, and Skempton's coefficient which gives the fraction of total stress in the skeleton of the water-saturated rock that is transmitted to the pore fluid: $B = -3p/\sigma$, where p is the pore-fluid pressure and σ is the total stress in the rock skeleton.

The amplitude of the water-level rise at E-1 from May 2006 to December 2009 was Δh = 122 cm (see Fig. 3). Assuming that the rise amplitude Δh is completely determined by the increase in pore pressure, we may write

$$\Delta\varepsilon = -(\rho g\Delta h)/(2/3GB[(1 + \nu_u)/(1 - 2\nu_u)]), \quad (2)$$

where ρ is water density and g is the acceleration due to gravity. The value of ν_u was assumed to be equal to 0.3, which is the traditional assumption for an aquifer in the upper crust.

When no water runoff occurs (undrained conditions), the elastic parameters G and B can be found from formulas of pore-elastic theory using the tidal sensitivity of water level to areal strain A_S and to volumetric strain A_V (see Table 2) based on the tidal analysis of hourly water-level variations. The relevant relationships are given in [Kopylova and Boldina, 2006; Igarashi and Wakita, 1991; Roeloffs, 1988; Rojstaczer and Agnew, 1989; Van der Kamp and Gale, 1983].

Tidal and barometric variations occur rather distinctly in the water-level changes that were observed at YuZ-5 and 1303 (see Fig. 1, Table 1), which penetrate water-saturated rocks with fresh ground water without any signs of gas generation. For these wells we estimated E_b, A_S, A_V and the elastic parameters and porosity for water-saturated rocks and for statically isolated conditions in the well–aquifer system (see Table 2).

The calculation of Skempton's coefficient B that (2) involves was based on the compressibilities of the rock skeleton β, of the solid phase of the skeleton β_u, and of the fluid β_f. Below, we quote formulas for determining Skempton's coefficient:

$$B = (\rho g A_S \beta)/[1 + \rho g A_S(\beta - \beta_u)], \quad (3)$$

$$B = (\beta - \beta_u)/[(\beta - \beta_u) + \phi(\beta_f + \beta_u), \quad (4)$$

$$B = \rho g \beta_u A_V, \quad (5)$$

where ϕ is porosity, ν is Poisson's ratio for runoff conditions (drained conditions, $\nu = 0.25$), $\beta_u = 0.3 \times 10^{-10}$ Pa^{-1} (quartz) [Roeloffs, 1988]. Relation (3) is after [Igarashi and Wakita, 1991], (4) and (5) after [Rojstaczer and Agnew, 1989]. For the case of the E-1 well, (3)–(5) involve unknown values of A_S, A_V, B, and ϕ.

As noted above, the water level at E-1 does not respond to earth tides. Consequently, we did not estimate the strain sensitivity of the water level that is required for determining the elastic parameters of the aquifer from water-level tidal sensitivity, as was the case for YuZ-5 and 1303, but used data on the hydrogeodynamic precursor prior to the December 5, 1997 Kronotskii earthquake M_W = 7.8 (abbr. KE) as observed at E-1 and YuZ-5. The KE was preceded by bay-like changes in water level at both wells that lasted about 3 weeks, synchronous with the pre-seismic motions at the GPS stations of the Kamchatka network [Kopylova, 2006]. The amplitudes of water-level drops Δh at E-1 and YuZ-5 were –1 cm and –11 cm, respectively. Kopylova [2006] assumed that these water-level drops were due to quasi-elastic strain in the expansion of water-saturated rocks that occurred during the evolution of aseismic movements that preceded the KE. In addition, it was assumed that the pre-seis-

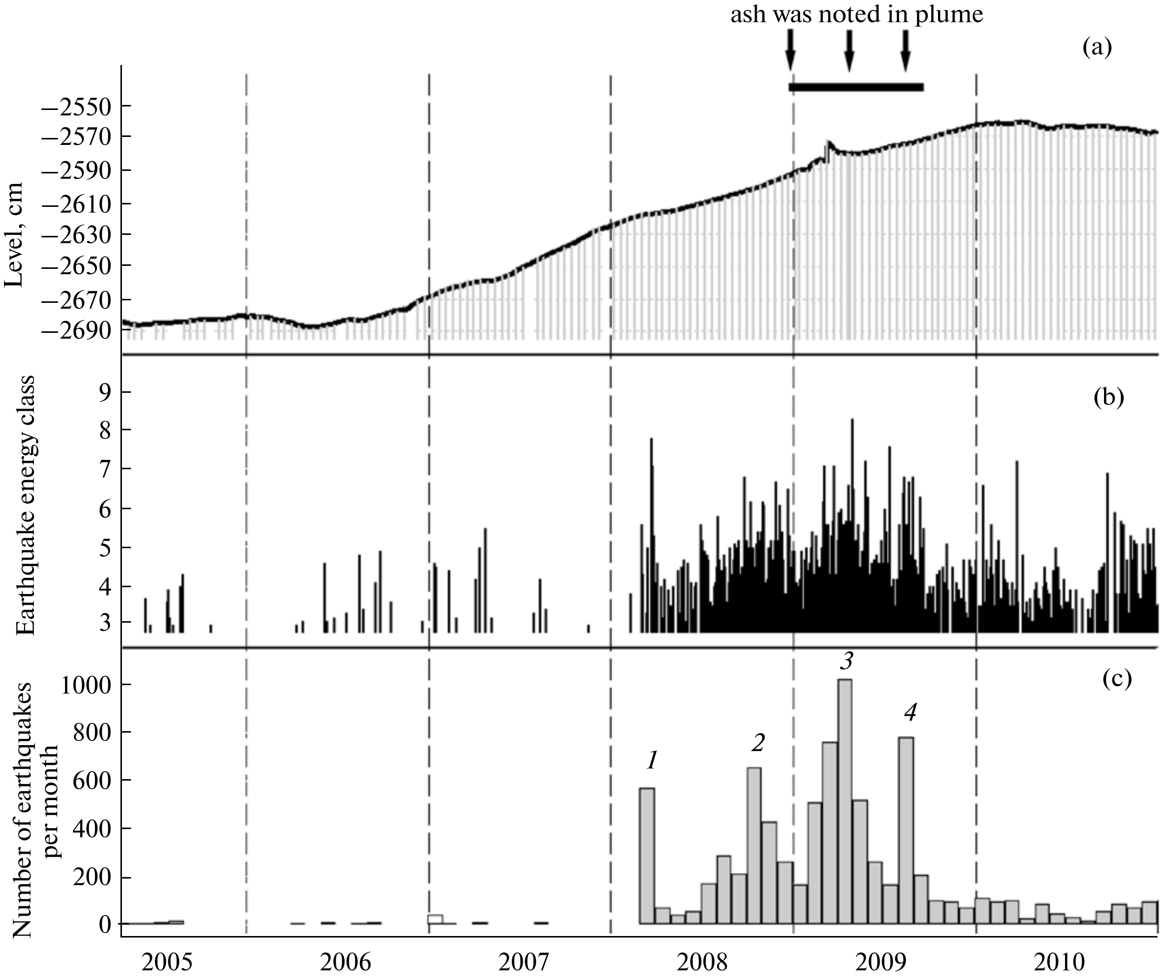

Fig. 4. Water-level variations in the E-1 well in 2005–2010 as compared with the evolution of seismicity and steam-and-gas activity on Koryakskii Volcano: (a) water-level changes based on digital and manual measurements (given in depths of water level below well head); horizontal line shows the time interval of phreatic eruption; (b) earthquakes with energy classes $K_S = 3.1–8.3$ in the area of Koryakskii Volcano; (c) total monthly number of earthquakes (numerals denote the maxima of seismic activity: (*1*) March 2008, (*2*) October 2008, (*3*) April 2009, (*4*) August 2009).

mic expansion strain was about the same around both wells, which are 11 km distant (see Fig. 1).

The approximate value of the strain sensitivity of the water level at E-1 was derived from the relation $\Delta h^{\text{YuZ-5}}/A_V^{\text{YuZ-5}} = \Delta h^{\text{E1}}/A_V^{\text{E1}}$, where $A_V^{\text{YuZ-5}} = 0.161\ \text{m}/10^{-7}$, and was $A_\nu^{\text{E1}} = 0.015\ \text{m}/10^{-7}$.

With 10^{-7} $\nu = 0.25$ for Poisson' ratio, the value of A_S^{E1} is $0.010\ \text{m}/10^{-7}$.

The storage of water-saturated rocks *S* and the fluid compressibility β_f were estimated from the modeled recovery of the water level after an acoustic emission sensor was dropped into the wellbore on March 11, 2009 (see the sudden rise in water level in Fig. 4a). This placement of the sensor was followed by a water level rise of 13 cm with subsequent stabilization during 40–45 days.

Data from 10-min measurements during the stabilization period were compared with the standard curves of the water-level drop for specified transmissivity T and storage S (Fig. 5) as obtained by solving the equation of nonlinear filtration for a perfect well for the degree and character of penetration for a homogeneous, infinite, headwater aquifer [Sindalovskii, 2006].

The modeling gave the best value of ground water compressibility for a transmissivity of $T = 0.004\ \text{m}^2/\text{day}$ and storage of 5.8×10^{-5} (see Fig. 5), viz., $\beta_f = 4.4 \times 10^{-9}\ \text{Pa}^{-1}$. This value is an order of magnitude greater than the compressibility of ordinary water and indicates the low concentration of free gas in the pore and fissure space of the water-saturated rocks.

The compressibility of the rock skeleton β (see Table 2) was obtained from the relation $S = d[\rho g(\beta + \phi\beta_f)]$, where *d* is the aquifer thickness, 22 m. The com-

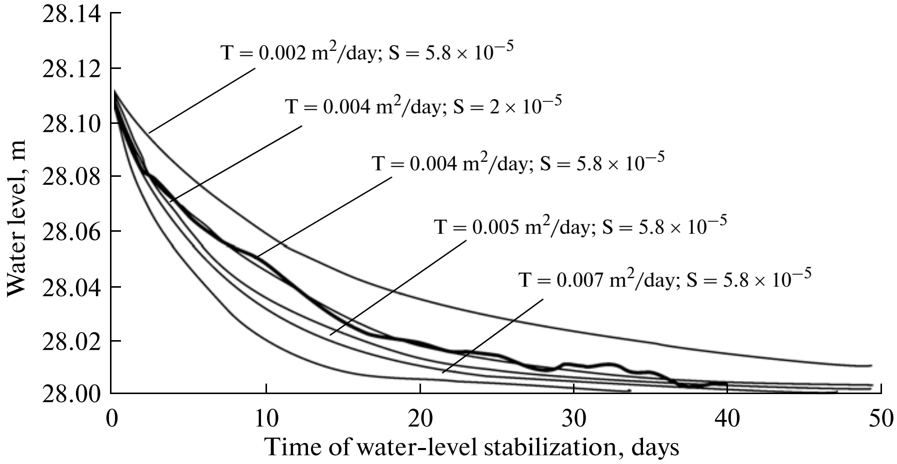

Fig. 5. Stabilization of the water level in the E-1 well after an acoustic emission sensor was installed in the wellbore: heavy line shows water-level changes based on 10-min measurements, light lines show calculated water-level drops for various values of transmissivity T and storage S.

pressibility of the rock skeleton $\beta = 7.37 \times 10^{-11}\text{Pa}^{-1}$ obtained for the 1303 well (see Table 2) was used as an initial approximation to estimate the storage of water-saturated rocks that were penetrated by the E-1 well. The porosity ϕ was assumed to be equal to or below 0.06, also by analogy with the porosity of the Neogene tuffaceous–sedimentary deposits that were penetrated by the 1303 well in the 517–717 m depth range (the average depth is 617 m).

It should be noted that the transmissivity value $T = 0.004\ \text{m}^2/\text{day}$ is quite consistent with that obtained previously (1991) from water-level recovery observations following a pumping test.

The values of B are 0.07 using (3), 0.17 from (4), and 0.044 from (5). To sum up, the estimate of B using (3)–(5) is in the range 0.04–0.17, with the average value being 0.09.

The shear modulus G in (2) was found from the relation $G = 3/2[(1 - 2\nu)/\beta(1 + \nu)]$ [Van der Kamp and Gale, 1983].

Using (2) and the estimated elastic parameters of the aquifer (see Table 2), we obtain an approximate value of the volumetric compression strain $\Delta\varepsilon = -(4.1 \times 10^{-6} – 1.5 \times 10^{-5})$ during the time period of anomalous water-level rise around the E-1 well.

On the Origin of the Compression Strain Source in Water-Saturated Rocks

The water-level rise at E-1 and the seismicity behavior in the Koryakskii Volcano area are considered here as phenomena that show a paragenetic relationship. It has been shown [Seliverstov, 2009; Senyukov and Nuzhdina, 2010] that the earthquake swarm began in March 2008 in the northern sector of the nearly north–south strip zone (see Fig. 1). After relative quiet in April–June 2008 the epicenters moved to its southwestern sector, which contains the volcano's edifice. During the phase of fumarole activity from late December 2008 to August 2009 (see Fig. 2a–2c) earthquakes occurred within the entire north–south strip zone, including the volcanic edifice. The earthquake with the highest $K_S = 8.3$, occurred on April 30, 2009 at a depth of 5 km.

The total seismic energy released during the earthquake swarm occurrence from March 2008 to December 2009 was 5.4×10^8 J, corresponding to an earthquake of energy class $K_S = 8.7$ and a magnitude of about 4. If all seismic energy had been released in a single event, then the area of the "equivalent source" for such an earthquake would be, after Riznichenko [1976], 2 km long and 1 km wide, with an area of 2 km^2. Such dimensions of the "source" are an order of magnitude smaller than the actual area of seismic activation in the area of Koryakskii Volcano (about 40 km^2). This shows that the nearly north–south strip zone of epicenter concentration (see Fig. 1) outlines an activated area of recent tectonic movements in the upper crust of the Avacha volcano-tectonic depression, which contains the edifice of Koryakskii Volcano.

Two hypotheses exist to explain the causes of fumarole activity on Koryakskii. One explains the observed seismic and volcanic occurrences via the emplacement of magmatic material into a hypothetical steam–gas collector beneath the volcano [Gordeev et al., 2009]. The basic assumption is that the area of seismic activity corresponds to the collector position. Under this hypothesis, the source of compressive strain around the well is the increase in fluid pressure in the collector and its elastic transmission to the area of the well. The lowest energy of the associated process can be estimated from steam-and-gas discharge (3×10^9 J after [Droznin and Dubrovskaya, 2010].

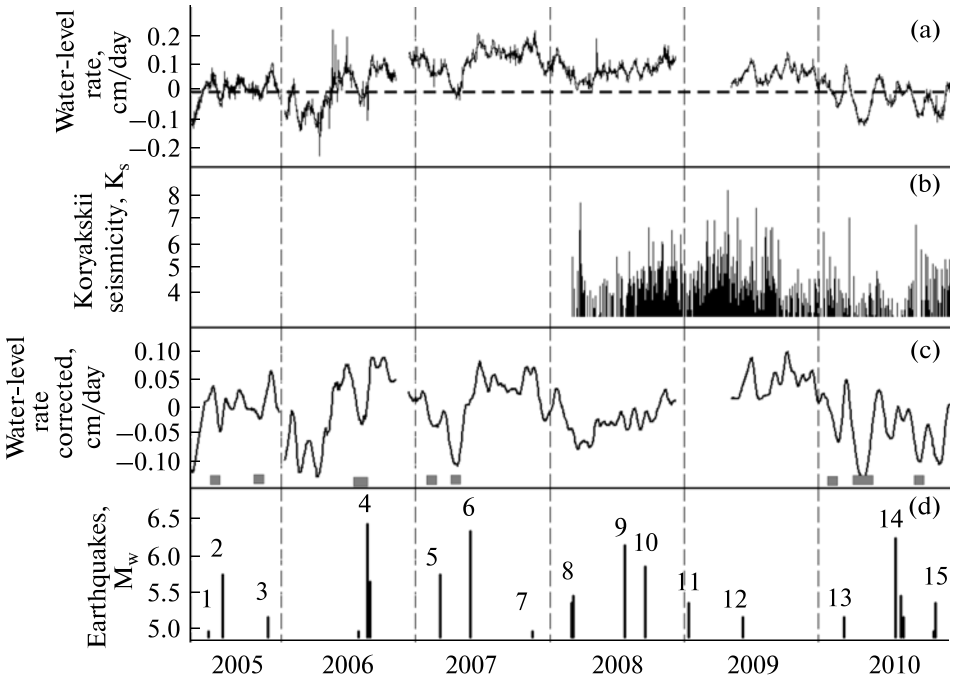

Fig. 6. Water-level variations at the E-1 well in 2005–2010 as compared with the evolution of seismicity in the area of Koryakskii Volcano and large earthquakes: (a) daily rate of water-level change with compensated baric variations (heavy line shows an average of mean daily data in a moving window of 15 days); (b) $K_S \geq 4.0$ earthquakes in the area of Koryakskii Volcano; (c) mean daily rate of water-level change as corrected for the mean trend rate over the relevant fragment (horizontal lines show the times of hydrogeodynamic precursors); (d) $M \geq 5.0$ earthquakes at distances of $R \leq 350$ km from the well: numerals denote identification numbers of seismic events to tally with Table 3.

Gordeev and Droznin [2010] set up a comparison between the hypothetical volume of magmatic material (1.2×10^7 m^3) that might provide the heat discharge of the volcano during the eruption and the earth volume where the earthquakes occurred (4.5×10^9 m^3). The two volumes differ by two orders of magnitude. From this comparison, the authors [Gordeev and Droznin, 2010] conclude that the emplacement of magmatic material is unable to explain the earthquake swarm.

The other hypothesis relates the occurrence of the earthquake swarm in the area of Koryakskii Volcano to the geodynamic setting that controls the generation of the north–south extensional zone [Seliverstov, 2009]. The destructive extensional processes within this zone affected the magma conduits beneath the volcano for some time and were accompanied by increased fumarole activity. In that case, the leading cause of the water-level rise lies in the tectonic extensional processes, which were accompanied by compression of water-saturated rocks in the area of the well.

We prefer the second hypothesis, that is, we believe that the strain source consisted in tectonic tension stresses within the north–south strip zone, which were active both before and during the seismic activation in the area of the volcano and its increased fumarole activity. The seismic energy released within the north–south strip zone (0.5×10^9 J) is of the same order as the energy of the steam–gas discharge (3×10^9 J). Assuming that the released seismic energy is at most a few percent of the total energy of tectonic movements, it becomes obvious that such movements were the main source of compression for the water-saturated rocks around the E-1 well. A similar conclusion about the dominant relationship of seismicity within the north–south strip zone to tectonic processes was also reached in [Gordeev and Droznin, 2010; Seliverstov, 2009; Senyukov and Nuzhdina, 2010].

ON THE OCCURRENCE OF HYDROGEODYNAMIC PRECURSORS DURING THE ACTION OF THE VOLUMETRIC COMPRESSION STRAIN SOURCE AROUND THE E-1 WELL

Figure 6 shows a plot of daily rate of water-level variation at E-1 compared with the seismicity in the area of Koryakskii Volcano and large tectonic earthquakes (Table 3) for the period between May 2005 and December 2010.

The plot of daily rates of water-level variations consists of four fragments; one could compensate for air pressure variations in the water-level variations in each of them. The dashed line in Fig. 6a shows a rate of 0 cm/day. Accordingly, the rate values above the

Table 3. The M ≥ 5.0 earthquakes at distances of R ≤ 350 km from the E-1 well, May 2005 to 2010

Identification number of seismic event*	Earthquake date, dd/mm/yy	Magnitude, M	Energy class, K_S	Distance to well, R, km	Presence of precursor in water-level changes
1	08.06.05	5.0	12.6	329	–
2	26.07.05	5.8	13.3	131	+
3	26.11.05	5.2	12.7	125	+
4	31.07.06	5.0	12.9	219	+
	24.08.06	6.5	14.3	276	
	01.09.06	5.7	12.9	110	
5	10.03.07	5.8	14.3	326	+
6	30.05.07	6.4	13.6	204	+
7	17.11.07	5.0	12.8	110	–
8	01.03.08	5.4	12.9	134	–
	06.03.08	5.5	13.1	270	
9	24.07.08	6.2	14.0	276	–
10	18.09.08	5.9	12.7	167	–
11	14.01.09	5.4	12.5	147	n.d.
12	11.06.09	5.2	12.5	134	n.d.
13	13.03.10	5.2	12.1	139	+
14	30.07.10	6.3	14.1	152	+
	30.07.10	5.4	12.6	153	
	15.08.10	5.5	12.9	145	
	21.08.10	5.2	12.6	296	
15	12.11.10	5.0	12.4	259	+
	16.11.10	5.4	13.3	185	

Note: * seismic events closer than one month in time of occurrence have been combined; –, no precursor in the water level changes was detected; +, a precursor in the water level changes was detected; n.d. data on the water level changes are not available due to technical issues.

dashed line show rising levels and those below the line indicate decreasing levels.

A persistent water-level rise began starting on May 20, 2006. That date can be treated as the beginning of the effective manifestation of the volumetric compression strain source around the well. The highest rates of water-level rise (up to 0.15–0.20 cm/day) were observed during June–December 2007 and preceded the beginning of the earthquakes swarm in March 2008 (see Fig. 6b). The rates were lower (0.05–0.12 cm/day) and varied slightly during the seismic activation and the eruption of Koryakskii Volcano.

Table 3 lists all M ≥ 5.0 earthquakes that occurred between May 2005 and December 2010 within 350 km of the well. Two earthquakes less than a month apart were combined to form a single seismic event. Comparison between the variations in the rate of water-level change and the times of seismic events (see Table 3) showed that hydrogeodynamic precursors appeared before the M = 5.0–6.5 earthquakes in 2005–2006 and in the earlier half of 2007. Since the later half of 2007 and during the subsequent 2.5 years no hydrogeodynamic precursors occurred before such events. This indicates the decreased sensitivity of the well to the precursory processes of large tectonic earthquakes during the action of the volumetric compression strain source.

Since January 2010 we have observed a steady decrease in the daily rate of the water-level rise. In February 2010, the long-term rise was replaced by a drop in water level. This indicated the cessation in the action of the volumetric compression source around the well accompanied by stabilization of the hydrodynamic regime. The hydrodynamic precursor was recorded before all three seismic events in 2010 (nos. 13–15 in Table 3, see Fig. 6c).

CONCLUSIONS

(1) Several remarkable features of the E-1 well exist, which distinguish it from the other observation wells in the Avacha volcano-tectonic depression (see Tables 1 and 2); these are the stagnant conditions of

water exchange in the controlled sequence of water-saturated rocks and their very low transmissivity, increased compressibility of ground water due to the presence of gas, and a low Skempton's coefficient (B = 0.04–0.17).

(2) The water level at the E-1 well gradually rose between May 20, 2006 and December 2009 due to volumetric compression strain of the water-saturated rocks and increasing pore pressure that preceded and accompanied the seismic and fumarole activation of Koryakskii Volcano. The most likely source of compression strain of water-saturated rocks might be the tectonic forces that are related to the generation of a north–south extensional strip zone in the crust around Koryakskii Volcano as stated in Seliverstov [2009]. The increase in tectonic stresses since March 2008 was accompanied by activation of low magnitude seismicity within an extended north–south strip zone (see Fig. 1) and by a weak phreatic eruption (see Figs. 2a–2c).

A detailed study of the rate of the rising trend in the water-level variations at E-1 showed that the phase of maximum pressure increase in the aquifer was observed between June and December 2007 and preceded the beginning of an earthquake swarm within the north–south strip zone. The pressure rose at a lower rate during the swarm and the eruption of the volcano. In January 2010 the pressure in the aquifer stopped rising. This indicates cessation or considerable weakening of the recent tectono-magmatic processes in the upper crustal horizons of the Avacha volcano-tectonic depression (the area of Koryakskii Volcano).

One consequence of the action of the volumetric compression source and the increase in pore pressure was decreased sensitivity of the hydrodynamic regime at the E-1 well to the precursory processes of large tectonic earthquakes from mid-2007 to 2009, that is, during 2.5 years.

In order to draw more definite conclusions about the origin of this strain source it would be helpful to have the focal mechanisms of the larger earthquakes that have occurred in the area of Koryakskii Volcano, as well as to perform multidisciplinary modeling of the evolution of the tectono-magmatic process taking water-level observations, the scheme of the hydrogeodynamic conditions in the interiors of the Avacha volcano-tectonic depression, the dynamics of seismicity, and the heat and mass discharge due to fumarole activity into account.

(2) The example of the E-1 well is the first to demonstrate how it is possible to estimate volumetric strain in water-saturated rocks based on data from water-level observations when there is no tidal response in water level.

The growth of pore pressure in the aquifer was $\Delta p =$ 12.2 kPa or 0.12 bars as inferred from the amplitude of water-level rise from May 20, 2006 to December 2009. Taking the estimated elastic parameters of the water-saturated rocks into account, we found an approximate range for volumetric compression strain around the well: $\Delta\varepsilon = -(4.1 \times 10^{-6}–1.5 \times 10^{-5})$. The estimates of $\Delta\varepsilon$ given above are preliminary, since they were based on several hypothetical assumptions, in particular, those about the approximate equality of volumetric strain values around the YuZ-5 and E-1 wells during the precursory process of the Kronotskii earthquake, about an approximate correspondence between porosity ϕ and the compressibility of the water-saturated rock skeleton β at the 1303 and E-1 wells, and others. In the opinion of one reviewer of this paper, the use of the nonstationary filtration model for a perfect well [Sindalovskii, 2006] in application to the E-1 well for estimating the storage of water-saturated rocks and the compressibility of the pore fluid is not completely justified, because this model does not incorporate the possible influence (on water-level variations) of the skin effect of the well and that due to the position of the filtration zone near an interface between two rock complexes that have different lithologic compositions (see Fig. 3). Such remarks are quite justified and are due, in the first place, to the insufficient development of the theory and methodology for the quantitative interpretation of water-level observations in order to monitor recent seismotectonic and volcano-tectonic processes.

(3) Multiyear monitoring of water-level variations shows the exceptional sensitivity of the hydrodynamic regime at the E-1 well to the precursory processes of large earthquakes that are generated during the subduction of the oceanic Pacific plate under the continental Sea-of-Okhotsk plate and to the tectonic processes in the interiors of the Avacha volcano-tectonic depression accompanied by increases in crustal seismicity and volcanic activity. The example of this well shows that two kinds of present-day geodynamic processes (one is the precursory processes and occurrence of large tectonic earthquakes and the other consists in local movements within the continental crust accompanied by seismic and volcanic activations) may "overlap" and produce responses in the water-level changes that overlap as well.

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