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Coseismic well-level changes due to the 1992 Roermond earthquake compared to static deformation of half-space solutions

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SUMMARY

The M_w 5.4 Roermond earthquake of 1992 April 13 was one of the strongest events during the last 500 years in Central Europe. For the period March–May 1992, we collected records of 194 continuously operating well-level sensors, mostly located within 120 km of the epicentre. Nearly all wells penetrate unconfined or poorly confined Quaternary deposits with high hydraulic conductivities. 81 out of 194 raw data sets show a significant dynamic or step-like response of centimetre amplitude to the passage of seismic waves. Precursory anomalies are not obvious in these records. Coseismic well-level fluctuations could reflect a redistribution of stress and pore pressure in the brittle crust. Systematic analyses of such fluctuations may improve our knowledge of the role of pore fluids in crustal rheology and earthquake mechanics. The rather high number of individual observational records for a single event allows a regional correlation of the signs and amplitudes of the coseismic steps to changes in volume strain caused by the earthquake. The coseismic strain field at the surface was calculated for a homogeneous and a layered half-space. The results show reasonable agreement in the sign of the well-level steps but the amplitudes predicted from the wells' volumetric strain responses are much smaller than those that were recorded. Clearly, the coseismic well-level steps cannot be explained by volume strain changes, as derived from linear elastic models.

Key words: crustal deformation, hydrology, Roermond earthquake, seismotectonics.

1 INTRODUCTION

Seismotectonically induced hydrologic and geochemical phenomena have often been reported and are well documented in numerous cases. They include pre-, co- or post-seismic well-level changes (e.g. Wakita 1975; Igarashi *et al.* 1992; Kissin & Grinevsky 1990; Kissin *et al.* 1993; Kissin *et al.* 1996; Ohno *et al.* 1997; Roeloffs & Quilty 1997; Roeloffs 1998), spring and streamflow discharges (e.g. Koizumi *et al.* 1996), and changes in radon concentration and temperature of groundwater (e.g. Wakita 1981; Mogi *et al.* 1989). Comprehensive overviews of seismogenic hydrologic anomalies are given by Roeloffs (1988, 1996). Although basic ideas about the physical processes leading to the observed phenomena do exist, the exact mechanisms are in general poorly understood. Coseismic well-level fluctuations are believed to reflect sudden pore pressure changes, related to *in situ* volume strain changes and the redistribution of stress in the brittle crust (Böðvarsson 1970; Kümpel 1992; Muir-Wood & King 1993). Wakita (1975) observed coseismic signals following the Izu-Hanto-Oki earthquake of May 1974 in 59 out of 95 wells. They led him to suggest that the sign of a coseismic well-level step indicates whether the static, regionally induced volumetric strain is extensional (well levels fall) or contractional (well levels rise). Similar investigations showed that volume strain changes from simple dislocation models cannot explain all the characteristics of hydrological anomalies in the far field of an epicentre (e.g. Igarashi & Wakita 1991; Koizumi *et al.* 1996; Roeloffs & Quilty 1997; Quilty & Roeloffs 1997). Other possible mechanisms, for example changes of aquifer properties or of local pore pressure induced by seismic waves, failed to explain persistent well-level steps in a well near Parkfield, California (Roeloffs 1998). Clearly, adequate data and further quantitative analyses are needed to understand better the mechanisms involved and in particular the role of pore fluids in crustal rheology.

We here investigate coseismic well-level changes following the 1992 Roermond earthquake, Central Europe. Our data set is one of the most extensive collections of coseismic hydrological signals for a single earthquake. The study includes the search for systematic peculiarities in the data and a comparison with crustal volume strain changes calculated for a homogeneous and a layered half-space.

2 WELL-LEVEL DATA

On 1992 April 13, a M_w 5.4 earthquake hit the Dutch–German border area near the city of Roermond in the Roer Valley Graben in the Netherlands. The earthquake was one of the strongest events during the last 500 years in mid-Europe. Its tectonic, seismological and geological settings have been extensively studied (Braunmiller *et al.* 1994; Davenport & van Eck 1994; Scherbaum 1994; Trivonov *et al.* 1994; Dufumier *et al.* 1997). Through questionnaires we became aware of, and subsequently collected recordings from, a total of 194 continuously operating well-level sensors, mostly located within 120 km of the epicentre (Fig. 1). Nearly all data are from shallow wells drilled into the Quaternary deposits of the Lower Rhine embayment. The wells were not drilled for scientific purposes but for groundwater level monitoring, for example near construction zones. The well depths cover a range between 5.6 and 227 m; 92 per cent of the well depths are less than 40 m. Since neither tidal signals nor barometric pressure variations could be detected in the data, the connected aquifers seem to be unconfined or only partially confined. Nevertheless, 81 out of 194 well-level recordings show a clear response to the seismic event, either a dynamic (double-peak distortion) or a step-like response (drop or rise), or a superposition of both. Fig. 2 displays some level recordings from different well locations. Where occurring, the persistent steps, partly superimposed by dynamic responses, were typically of centimetre amplitude. Within a total of three months of available data for each location before and after the main shock, signals due to fore- or aftershocks could not be identified. We have no reason to interpret any occasional fluctuations before the main shock as precursory anomalies.

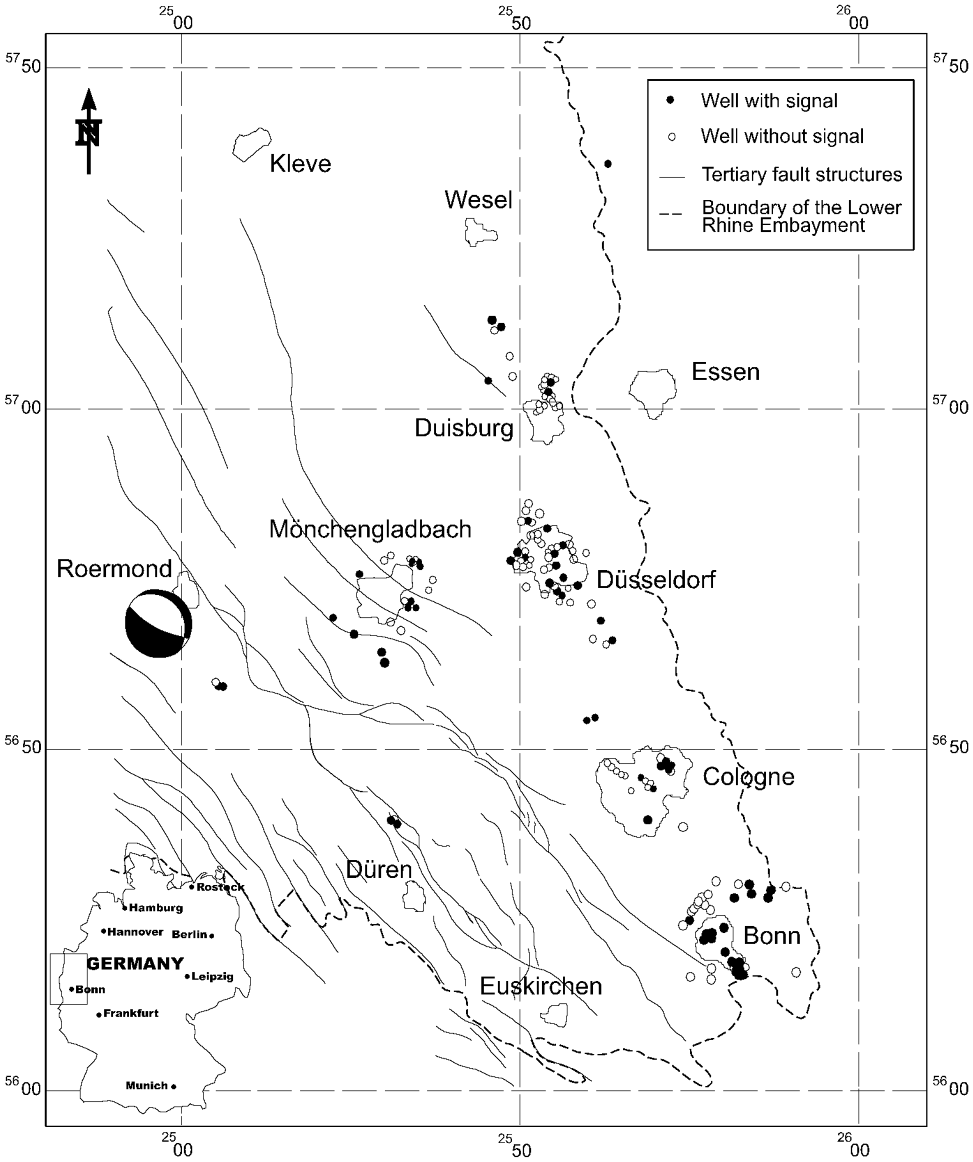

Figure 1. Map of the Lower Rhine embayment near the Dutch–German border in Central Europe, and distribution of well-level sensors showing either a significant hydrological anomaly (filled circles) or no response (empty circles) in the raw data. Well locations from 11 analysed data sets at distances beyond 120 km are not displayed on the map; four of these showed a coseismic step signal in the raw data sets. The epicentre of the Roermond earthquake of 1992 April 13 is indicated by its fault plane solution. Tertiary fault structures after Geologisches Landesamt NW (1988).

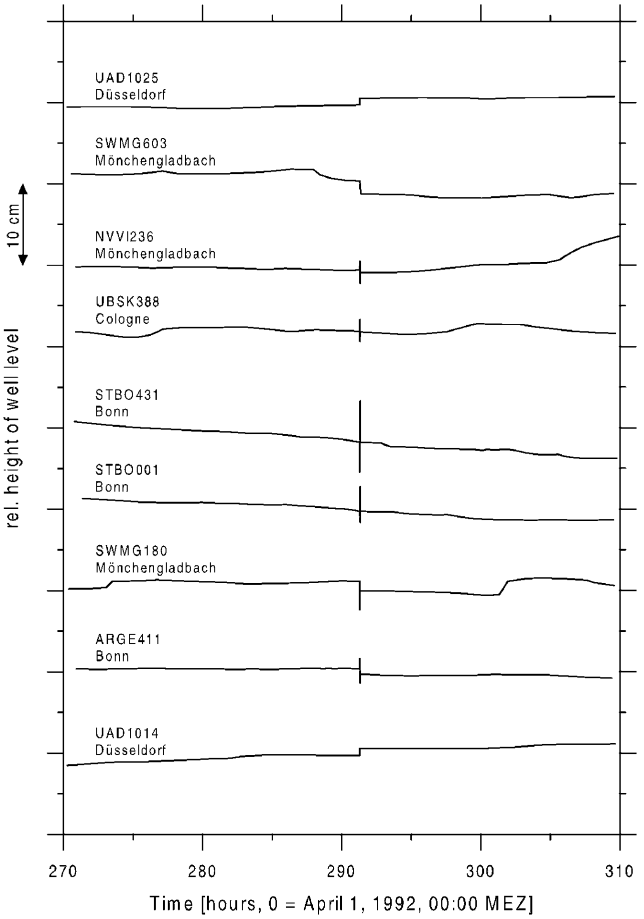

Figure 2. Well-level recordings from various locations, with code names. The timescale covers a range of roughly 20 hr before and after the earthquake. Significant dynamic and step-like responses are obvious.

To find out whether the signals could be due to instrumental effects, laboratory experiments with two mechanical well-level sensors of the type used at most locations were performed. The sensors consist of a float and a counterbalance weight and are mechanically connected to a pen writing on a slowly turning drum. The occurrence of step-like signals, up to several millimetres, could be stimulated by mechanical shock experiments and were shown to result from friction between the float and the well casing (Fischer 1997). This confirms the earlier results of Rexin *et al.* (1962), who observed persistent steps in well-level recordings with signs equal to those of the signal trend and concluded that these displacements were caused by friction. Therefore, small-amplitude (about 1 cm in the water level) step signals of 10 well which showed similar behaviour, that is, signs equal to those of the signal trend, were considered to be doubtful and were disregarded in the following analysis (Grecksch *et al.* 1997). A closer examination of the data sets revealed, in several cases, additional step-like anomalies with amplitudes of about 1 mm on the recording drum. The occurrence of these anomalies and a possible coseismic step with a similar amplitude on the same record may be due to the limited resolution of the respective mechanical writing systems. However, such coseismic steps are also considered as being doubtful. This led us to disregard another 25 coseismic steps.

Table 1 summarizes the signal statistics of the well-level sensors that we got notice of. Step-like responses were identified in 28 wells and dynamic responses in 45. Fig. 3 shows (a) the amplitudes of steps and (b) peak-to-peak dynamic responses as a function of the epicentral distances of the well locations. The uncertainty in amplitude is given by the reading error,

Table 1. Statistics of observed coseismic well-level anomalies: absolute number, percentage and amplitude range of steps and peak-to-peak dynamic responses. Step disregarded: steps probably due to instrumental effects; failure: reported well-level sensor out of operation during the event.

Type of coseismic anomaly	Number	Percentage	Amplitude range
Step	14	7.2	0.4–3.5 cm
Dynamic response	31	16	0.8–15 cm
Step plus dynamic response	14	7.2	—
Step disregarded	35	18	—
	⇒ 28 steps: 9 drops and 19 rises and 45 dynamic responses		
No reaction	95	49	—
Failure	5	2.6	—
Total	194	100	—

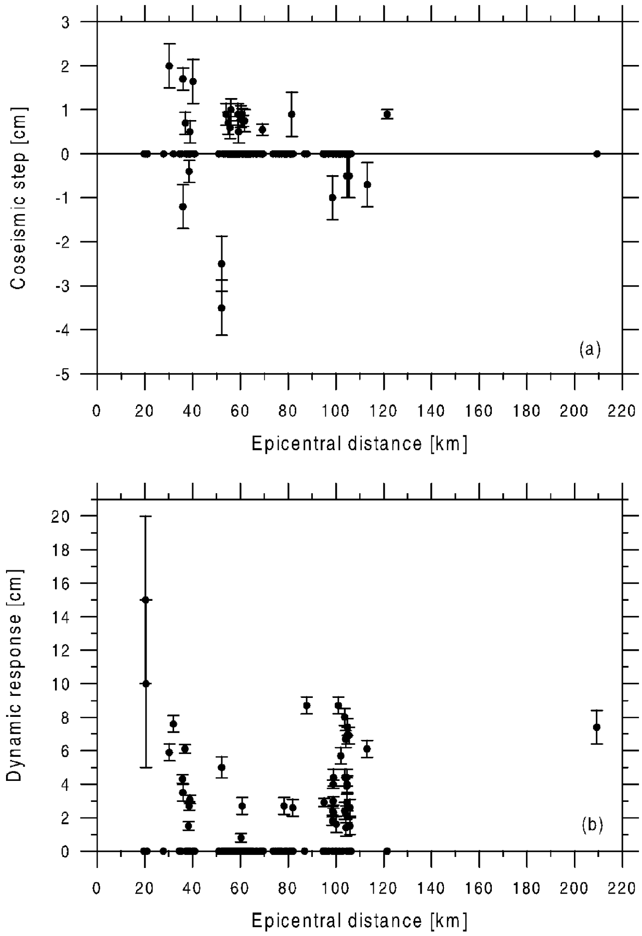

Figure 3. Size of steps (a) and peak-to-peak dynamic responses (b) versus epicentral distance. For non-zero amplitudes, the uncertainty is given by the reading error, which depends mainly on the amplification of the mechanical sensors.

which largely depends on the mechanical amplification of the sensor system. No error bars are plotted for recordings that showed no reaction, although their resolutions vary from ± 0.15 to ± 1.25 cm. In fact, we cannot rule out coseismic signals having remained undetected because of low sensor resolution. A decrease in the step size with increasing epicentral distance is not obvious (Fig. 3a); however, the dynamic response amplitudes do tend to decrease with increasing epicentral distances to about 85 km (Fig. 3b). The higher peak-to-peak amplitudes for epicentral distances around 100 km (area of Bonn) are probably due to a significantly higher macroseismic intensity in that area (Meidow & Ahorner 1994). The reported coseismic signal of a well near Basel (Switzerland) located at an epicentral distance of about 400 km is not displayed. It showed a well-level rise of 0.5 ± 0.25 cm and a peak-to-peak dynamic response of 1.5 ± 0.25 cm.

Dynamic responses that represent seismic oscillations occur preferentially in wells tapping highly transmissive aquifers. Also, suitable values for the height of the water column in the well and the well bore radius should be reached (Krauss 1974; Liu *et al.* 1989). To investigate a possible relationship between the occurrence of anomalies, their amplitudes and the hydraulic conductivities of the connected aquifers, hydraulic tests (slug tests and single-well pumping tests) were conducted in a representative subset of 68 wells. Values of hydraulic conductivities were deduced after Bouwer & Rice (1976) using the program AQTESOLV (Duffield & Rumbaugh 1991). Fig. 4 displays the results. There is no obvious correlation between signal amplitudes and hydraulic conductivities, but note that almost all anomalies occurred in wells tapping aquifers of relatively high hydraulic conductivities (between 10^{-5} and 10^{-2} m s^{-1}). However, the latter result is not significant, since almost all wells without reaction to the earthquake are also connected to aquifers of this range of hydraulic conductivity values. Because of insufficient time resolution, we did not investigate the dynamic well-level signals further but restricted ourselves to the study of the step-like responses.

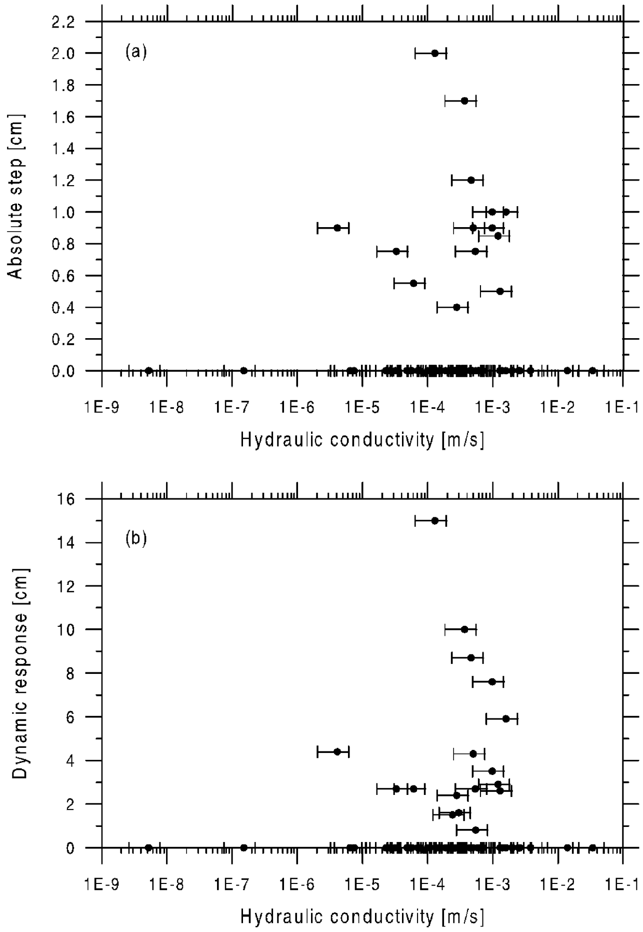

Figure 4. Size of steps (a) and peak-to-peak dynamic responses (b) versus hydraulic conductivities of the tapped aquifers, as revealed by single-well hydraulic field tests. Values of hydraulic conductivities are deduced from Bouwer & Rice (1976) using the program AQTESOLV (Duffield & Rumbaugh 1991). An uncertainty of ± 50 per cent is adopted.

3 STATIC DEFORMATION

Well levels may change in response to deformation of the connected aquifers because of passing seismic waves (e.g. Cooper *et al.* 1965; Liu *et al.* 1989), fault creep (e.g. Wesson 1981; Roeloffs *et al.* 1989), tidal strain (e.g. Bredehoeft 1967; Van der Kamp & Gale 1983) or atmospheric loading (e.g. Rojstaczer 1988; Rojstaczer & Agnew 1989). A coseismic strain imposed by an earthquake is expected to result in a step-like well-level change. In a simple linear model, water levels fall or rise depending on whether the connected aquifer expands or contracts due to the seismogenic redistribution of the regional strain field (Wakita 1975; Roeloffs 1996). To verify whether the signs of the 28 observed steps following the Roermond earthquake could be explained by earthquake-induced deformation, we calculated the static volume strain at the surface of a half-space. As a first approximation, analytical solutions developed by Okada (1992) for an isotropic homogeneous half-space were considered. Since numerous studies have demonstrated the effects of crustal layering on the strain field (e.g. Chinnery & Jovanovich 1972; Ma & Kusznir 1996), we extended our calculations to a layered half-space model using analytical expressions developed by Roth (1990). Brief descriptions and the results of both approaches are discussed below.

3.1 Model calculations for a homogeneous half-space

Okada (1992) presented analytical solutions for internal and surface displacements due to point and finite rectangular sources in a homogeneous elastic half-space for shear and tensile faults. From the displacement values u_i, the components of the strain and stress tensors (ε_{ij}, σ_{ij}, with $i, j \in \{x, y, z\}$) and the volume strain, Δ_V, can easily be calculated. The static volume strain field following an earthquake consists of expanding and contracting zones. For a point source, the spatial distribution of these zones depends critically on the geometry and depth of the focal mechanism. The amplitude and areal extent of volume strain are defined by the seismic moment and the elastic parameters of the formation. Adopting a finite rectangular source, the average dislocation on the rupture area and its areal extent also have to be specified. Tables 2 and 3(a) summarize the parameters that we used for the earthquake's source and the homogeneous half-space, respectively. In the literature slightly different source parameters for the Roermond earthquake can be found (Table 1 in Camelbeeck *et al.* 1994). We here use the parameters given by Camelbeeck *et al.* (1994) and Camelbeeck & Meghraoui (1996). The elastic half-space is defined by adopting some

Table 2. Source parameters for calculating the static coseismic strain field for a homogeneous and layered elastic half-space (after Camelbeeck *et al.* 1994 and Camelbeeck & Meghraoui 1996).

Focal mechanism	normal dip-slip
Dip angle	70°
Strike angle	N125°E
Fault rupture area	11 km^2
Seismic moment	$(1.4 \pm 0.6) \times 10^{17}$ Nm
Focal depth	17 ± 1 km
Average dislocation on rupture area	0.3 m

Table 3. Elastic parameters of (a) the homogeneous and (b) the layered half-space. For source parameters, see Table 2.

(a) Parameters of homogeneous elastic half-space.

P-wave velocity, V_P	5900 m s^{-1}
S-wave velocity, V_S	3400 m s^{-1}
Density, ρ	2550 kg m^{-3}
1. Lamé constant (rigidity), μ	29.5 GPa
2. Lamé constant, λ	29.8 GPa
Poisson's ratio, v	0.25
Compressibility, c	2×10^{-11} Pa^{-1}

(b) Parameters of layered elastic half-space.

Layer	V_P (m s^{-1})	V_S (m s^{-1})	ρ (kg m^{-3})
0–0.5 km	2000	600	2000
0.5–2 km	4700	2500	2400
2–4 km	5500	3200	2500
4–∞ km	6200	3600	2600

average density and seismic *P*- and *S*-wave velocities. From these parameters, other elastic constants (Lamé's constants, Poisson's ratio, compressibility) have been deduced (see Table 3a).

Using these parameters, we calculated the volume strain field on a 1 km surface grid for both a point source and a finite rectangular source. The results are presented in Figs 5(a) and (b). Negative values indicate compressional strain. Locations of wells showing either a coseismic water-level drop (empty circles) or a rise (filled circles) are displayed. The well near Basel, Switzerland, with a reported rise of 0.5 ± 0.25 cm is not shown on the contour plots. Wells with dynamic responses or without signals are not displayed. Except for five wells in the vicinity of a zero-strain nodal line (passing near the cities of Mönchengladbach and Cologne), the signs of the recorded coseismic steps are in agreement with the earthquake-induced static volume strain, although the spatial pattern of

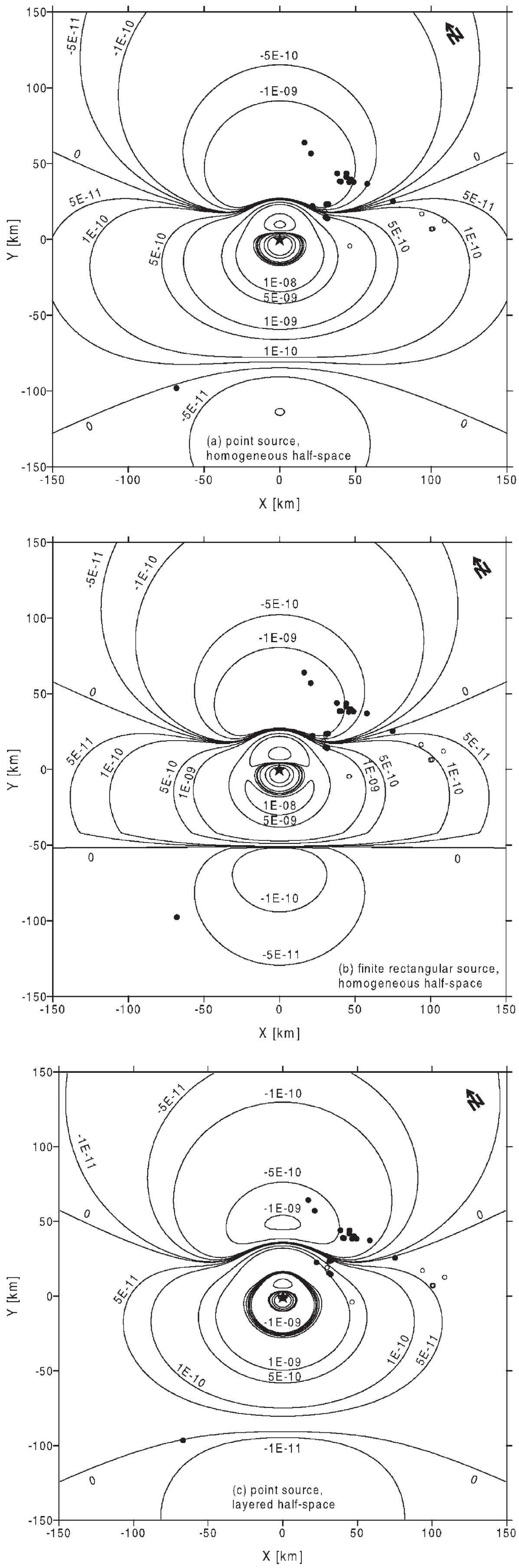

Figure 5. Contour plots of the surface static volume strain field induced by the 1992 Roermond earthquake, as calculated for (a) a point source, (b) a finite rectangular source, both in a homogeneous elastic half-space, and (c) a point source in a layered elastic half-space (positive values indicate extensional strain; model parameters in Table 3). Wells showing a water level rise or drop are indicated by filled or empty circles, respectively (wells with steps probably induced by instrumental effects are not shown, *cf.* Fig. 1). The epicentre is marked by a star.

the extensional area is different in both cases. At the well sites, the surface strain amplitudes obtained from the point source model are about 40 per cent larger than those from the rectangular source model. The zonation suggests that the well levels reflect coseismic volume strain changes.

3.2 Model calculations for a layered half-space

To check how much a layered structure would influence our results for the homogeneous half-space, we expanded our calculation by using a more realistic earth model with four layers. The model was taken from Pelzing (personal communication, 1995) and its parameters are listed in Table 3(b). A thin (500 m), very soft layer with a high Poisson's ratio (0.45) covers three other layers and a homogeneous half-space with elastic constants typical of the upper crust. The soft layer presents the relatively young sediments of the Lower Rhine embayment. The source is located in layer 4, which has the same parameters as the half-space below. The source is assumed to have a static moment equal to that used before and to be point-like, as the emphasis here is on the influence of the layer structure.

The results are shown in Fig. 5(c). It should be noted that the area just around the epicentre is influenced by numerical errors. The other parts can be compared with the results of the point source in the purely homogeneous half-space. The spatial distributions of the two volume strain fields are very similar to each other. Two step data points with negative sign east of the epicentre are now clearly in the dilatational (positive) region, whereas six with positive sign (three more than before) do not match the model now. All of these points are in the immediate vicinity of a nodal line.

In general, the volume strain in the layered model is only 30 per cent of that in the half-space model. This needs some comment. There are no resonance effects in the quasi-static surface strain results; nevertheless, an increase of surface deformation can be expected for individual strain components in certain areas if the surface layer is soft. This is due to the contrast in shear modulus and, in particular, a Poisson's ratio differing from 0.25. However, determination of the volume strain tends to cancel increases in horizontal strain with decreases in vertical strain. A test for a homogeneous half-space using the source layer moduli for the entire half-space and one using the surface layer moduli showed a similar difference of a factor of 3–4 between both volume strain fields in favour of the higher moduli. The homogeneous model here uses some intermediate values for Lamé's constants. Moreover, in this modelling there is the problem of which moment to choose. The layer structure used to determine the hypocentre, magnitude and seismic moment of the Roermond earthquake by Camelbeeck *et al.* (1994; no details of the earth model used are given) is not necessarily the same as that used here to calculate surface volume strain in the layered model. Magnitudes and seismic moments determined from seismograms from seismometers at the surface are biased by the waves having travelled through the layered structure. Similarly, the size of the seismic moment determined from the average coseismic slip and size of the rupture plane crucially depends on the shear modulus of the source layer.

To summarize, the differences between the three models (the extended source and the two point source models) are small with regard to the spatial distribution of the coseismic volume strain changes. The variations found in the amplitudes and those expected from all the uncertainties mentioned above are about a factor of 3–4 on average of the observation points. Small variations (± 1 km) in the focal depth of the point source and the shallowest depth reached by the finite rectangular rupture yield maximum uncertainties in the strain values at the well sites of ± 30 per cent.

4 STATIC CONFINED VOLUME STRAIN EFFICIENCIES

Assuming that the elastic behaviour of the affected aquifers can be described by the linear theory of poroelasticity, we may estimate the expected sizes of water-level steps at the individual well locations. This theory was developed by Biot (1941) to describe 3-D consolidation and was reformulated by Rice & Cleary (1976). A comprehensive review was given by Kümpel (1991), and geophysical applications were discussed by Wang (1993). Following Rice & Cleary (1976), Hooke's linear stress–strain relation for isotropic poroelastic media can be written as

$$2\mu\varepsilon_{ij} = \sigma_{ij} - \frac{\nu}{1+\nu}\sigma_{kk}\delta_{ij} + \frac{3(\nu_u - \nu)}{B(1+\nu)(1+\nu_u)}P\delta_{ij}\,. \tag{1}$$

The deformation of an isotropic poroelastic medium is thus fully described by the four parameters μ, B, ν, and ν_u, where ν and ν_u are the drained and undrained Poisson's ratios, respectively, μ is the rigidity or shear modulus and B is Skempton's pore pressure coefficient, indicating the change in pore pressure per unit change in confining pressure for undrained conditions (Skempton 1954). P is the change in subsurface pore pressure, δ_{ij} is the Kronecker symbol, and subscripts kk follow Einstein's summation convention. From eq. (1), and under the assumption of no fluid flow, an expression for the strain-induced change in pore pressure can be obtained (Roeloffs & Quilty 1997):

$$P = -BK_u\Delta_V\,, \tag{2}$$

where K_u is the undrained bulk modulus of the aquifer. Eq. (2) describes the change in pore pressure, P, as being proportional to changes in volume strain, Δ_V. The quantity BK_u is termed the *static confined volume strain sensitivity* (Rojstaczer 1988). For sufficiently slow changes in pore pressure, the water-level change Δh in a well tapping a confined aquifer is given by $\Delta h = P/\rho_f g$ (Roeloffs 1988), which yields the *static confined volume strain efficiency* (Rojstaczer & Agnew 1989; Roeloffs 1996),

$$\frac{\Delta h}{\Delta_V} = -\frac{BK_u}{\rho_f g}\,, \tag{3}$$

with g and ρ_f denoting the gravitational acceleration and the fluid density, respectively. If BK_u is known, eq. (3) can be used to estimate the volume strain Δ_V that could have produced a coseismic well-level change Δh under confined conditions. Values for BK_u could, in principle, be obtained from the analysis of tidally induced well-level fluctuations, since the tidal forcing is well known (Wilhelm *et al.* 1997).

In the present data, no tidal variations are obvious, either because the aquifer's compressibility is increased by small amounts of gas within the pore space, or because unconfined or poorly confined conditions prevail. The analysis of available well-log information suggests unconfined conditions. We may still accept a confined situation for the generation of coseismic well-level steps since the causative strain change is a much more rapid process (of the order of seismic frequencies) than tidal phenomena (Rojstaczer & Riley 1990; Roeloffs 1996). On the other hand, steps generated under such conditions would not be expected to persist for a long time (e.g. hours). Nevertheless, appropriate values for B and K_u at the well locations cannot be constrained from tidal signals, so further considerations are based on ranges of realistic values for these parameters.

In poroelasticity, we may also make use of the convention that rigidity does not depend on the conditions of draining ($\mu = \mu_u$), so that the undrained bulk modulus K_u may be expressed as

$$K_u = \frac{2\mu(1+\nu_u)}{3(1-2\nu_u)}. \tag{4}$$

Moreover, the undrained Poisson's ratio, ν_u, has to obey

$$\nu_u = \frac{3\nu + B(1-2\nu)\alpha}{3 - B(1-2\nu)\alpha}, \tag{5}$$

with $\alpha = (K_u - K)/(K_u B)$, $0 \le \alpha \le 1$, denoting the coefficient of effective pressure (Rice & Cleary 1976; Kümpel 1991; Wang 1993).

The wells are mostly drilled into Quaternary deposits of the Lower Rhine embayment, which consist mainly of water-saturated sands and gravels. For such sediments, undrained and drained Poisson's ratios of $0.40 \le \nu_u \le 0.49$ and $0.10 \le \nu \le 0.15$ may be adopted (Kümpel 1989); these are in agreement with values derived from wave velocity observations (Hamilton 1971; Domenico 1977). Skempton's parameter is assumed to be $0.77 \le B \le 0.99$ to be consistent with eq. (5). Finally, allowing the range 10^8–10^9 Pa for the rigidity, volume strain efficiencies between 0.04 and 5 mm nε^{-1} appear to be plausible (i.e. 4×10^8–5×10^{10} Pa for BK_u; 1 n$\varepsilon = 10^{-9}$).

Fig. 6 displays expected well-level steps Δh as derived from volume strain Δ_V for different values of volume strain efficiency (in mm nε^{-1}). The observed step amplitudes are displayed for different well-depth ranges versus the volume strains at the well locations ($4.3 \times 10^{-12} \le \Delta_V \le 2.7 \times 10^{-9}$) that result from the homogeneous half-space solutions for the point source. Expected and observed step amplitudes can be compared. For instance, a static confined volume strain efficiency of 0.04 mm nε^{-1}, as is obtained for $\nu_u = 0.4$, $B = 0.77$ and $\mu = 10^8$ Pa, yields well-level changes between 10^{-5} and 10^{-2} cm for the range of volume strains indicated, clearly below the sizes of the observed coseismic steps (between 0.4 and 3.5 cm). Adopting the upper bound of $\Delta h/\Delta_v = 5$ mm nε^{-1}, which holds for $\nu_u = 0.49$, $B = 0.99$ and $\mu = 10^9$ Pa, only a few amplitudes of the observed steps can be explained by earthquake-induced volume strains, namely those reaching values of 10^9 or above (see also Fig. 5). For smaller strain efficiencies (≤ 2 mm nε^{-1}) or volume strains below 10^9, the sizes of the observed coseismic steps are 1–5 orders of magnitude larger than expected. Obviously, $\Delta h/\Delta_V$ depends strongly on the values of rigidity and undrained Poisson's ratio adopted. The discrepancy in observed and expected step amplitudes is independent of the well depths and step signs.

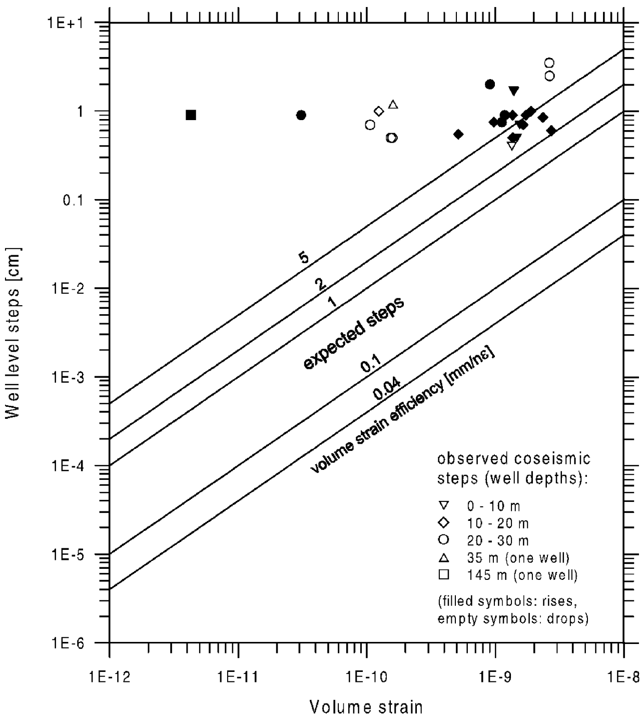

Figure 6. Expected and observed well-level steps, Δh, versus volume strain, Δ_V. The observed coseismic steps cover a range from 0.4 to 3.5 cm. They are displayed for different well-depth ranges and different step signs (filled symbols: well-level rises; empty symbols: drops). Expected well-level steps are calculated for static confined volume strain efficiencies from 0.04 to 5 mm nε^{-1} and for the indicated range of volume strains, as derived from the point source model calculations for a homogeneous half-space (*cf.* Fig. 5a). The volume strains for the other two models are smaller.

5 DISCUSSION AND CONCLUSIONS

Step-like coseismic well-level fluctuations are believed to reflect changes in crustal volume strain. Earlier studies of these phenomena demonstrated discrepancies between step signs and imposed strains, and revealed a lack of understanding of the processes involved (e.g. Igarashi & Wakita 1991; Roeloffs & Quilty 1997). We analysed recordings of coseismic step-like anomalies from 28 out of a total of 194 wells that were induced by the M_w 5.4 Roermond earthquake of 1992 April 13. Three types of model calculations were used to determine the static volume strain field of the earthquake. According to the solutions for a point and a finite rectangular source in a homogeneous half-space, 16 out of 19 wells with rising levels fall into areas of compression and seven out of nine wells with level drops are found within extensional areas. The solution for a point source in a layered half-space gives 13 out of 19 wells with rising levels falling into areas of compression and nine out of nine wells with level drops falling into extensional areas. All the wells with contradictory step signs are located close to a zero-nodal line. Since the signs of the steps mostly agree with the earthquake's static volume strain field, this study suggests that well-level fluctuations reflect coseismic volume strain changes.

Adopting static confined volume strain efficiencies $\Delta h/\Delta_V \le 2$ mm nε^{-1}, the amplitudes of the coseismic steps are found to be much larger than those predicted from the earthquake-induced strains. This confirms earlier findings for precursory anomalies (Roeloffs 1988) that simple elastic models of the crust fail to explain the local variability and amplitudes of hydrological signals. Huang *et al.* (1995) suggested a non-linear response of water levels to coseismic strains that might be due to site effects such as local heterogeneity in geological structures (Koizumi *et al.* 1996). Wang (1997) suggested that deviatoric stress may lead to significant coseismic well-level steps in regions of small volume strain.

The absence of tidal level variations in our data together with log information of the wells indicates unconfined or partially confined conditions. Although volume strain efficiencies derived for confined conditions also hold for short-term disturbances in unconfined aquifers (Rojstaczer & Riley 1990; Roeloffs 1996), the observed well-level steps should not persist for up to several days as in our recordings. In fact, the signal amplitudes obtained for confined conditions give upper bounds for similar effects occurring at unconfined conditions. As for instrumental effects, it would be rather surprising if the sticking of floats in the wells under field conditions were considerably larger than those found in laboratory tests.

Obviously, slow poroelastic deformation cannot account for the observed persistent steps. Thus, despite the new data, the mechanism of hydrological anomalies due to earthquakes remains speculative. As a pore pressure phenomenon, coseismic well-level variations seem to provide valuable information for the study of crustal rheology, earthquake mechanics and hydrological earthquake precursors. It is, however, puzzling that the relatively large data set of step-like coseismic well-level anomalies analysed here show on the one hand reasonable agreement between the signs of observed steps and the earthquake's static volume strain field, and on the other hand step amplitudes significantly larger than those predicted from static confined volume strain efficiencies.

There is order in the disorder, but other case studies, preferably with data from confined aquifers and observed by pressure transducers (instead of mechanical floaters), are necessary to tell us why.

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