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The Geological Society of America
Special Paper 509
2015

Widespread groundwater-level offsets caused by the M_w 5.8 Mineral, Virginia, earthquake of 23 August 2011

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ABSTRACT

Groundwater levels were offset in bedrock observation wells, measured by the U.S. Geological Survey or others, as far as 553 km from the M_w 5.8 Mineral, Virginia (USA), earthquake on 23 August 2011. Water levels dropped as much as 0.47 m in 34 wells and rose as much as 0.15 m in 12 others. In some wells, which are as much as 213 m deep, the water levels recovered from these deviations in hours to days, but in others the water-level offset may have persisted. The groundwater-level offsets occurred in locations where the earthquake was at least weakly felt, and the maximum water-level excursion increased with felt intensity, independent of epicentral distance. Coseismic static strain from the earthquake was too small and localized to have contributed significantly to the groundwater-level offsets. The relation with intensity is consistent with ground motion from seismic waves leading to the water-level offsets. Examination of the hydrographs indicates that short-period ground motion most likely affected the permeability of the bedrock aquifers monitored by the wells.

INTRODUCTION

The M_w (moment magnitude) 5.8 Mineral, Virginia (USA), earthquake of 23 August 2011 offset groundwater levels in bedrock observation wells as far away as 553 km. The earthquake-induced offsets are all unidirectional rises or drops (–0.47 m to +0.15 m) that persisted for hours, days, or longer. The large geographic extent of well-documented groundwater-level offsets is unprecedented for an earthquake of this moderate size.

We show that few, if any, of the observed groundwater-level offsets can be explained by coseismic static strain. Instead, the similar geographic distributions of the groundwater-level offsets and the intensities estimated from reports submitted to the U.S. Geological Survey "Did You Feel It?" (DYFI) website (http://earthquake.usgs.gov/earthquakes/dyfi/events/se/082311a/us/index.html; see Appendix 1) indicate that seismic ground motion changed aquifer properties at the well locations. Previous studies of groundwater-level offsets caused by seismic waves from

Roeloffs, E.A., Nelms, D.L., and Sheets, R.A., 2015, Widespread groundwater-level offsets caused by the M_w 5.8 Mineral, Virginia, earthquake of 23 August 2011, *in* Horton, J.W., Jr., Chapman, M.C., and Green, R.A., eds., The 2011 Mineral, Virginia, Earthquake, and Its Significance for Seismic Hazards in Eastern North America: Geological Society of America Special Paper 509, p. 117–136, doi:10.1130/2014.2509(07). For permission to copy, contact editing@geosociety.org.
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earthquakes (Roeloffs, 1998; Brodsky et al., 2003; Matsumoto and Roeloffs, 2003; Roeloffs et al., 2003) focused on a single well, or group of several wells, that have repeatedly responded to distant earthquakes. These previous studies also identified ground motion as the cause of changes to aquifer properties leading to water-level offsets, since these earthquakes could not impose enough static strain to account for these offsets.

Groundwater-Level Observations

The groundwater-level offsets, which were all too small to be noticed without instrumentation, were recorded by the U.S. Geological Survey and other agencies in 46 observation wells 15–213 m deep, equipped with transducers sampling at intervals of 60, 30, 15, or (in one well) 5 min (Table 1). The resolution of the water-level measurements generally is 0.003 m. All of the groundwater-level data discussed here, as well as detailed information about each site, can be downloaded from publicly accessible U.S. Geological Survey websites (see Appendix).

All the wells are in consolidated bedrock; 26 are in aquifers consisting of sandstone, shale, or a combination of these and intermediate sedimentary rocks. Two are in metamorphic or igneous formations, and two are in brecciated rock. Four are in limestone or dolomite, which can dissolve to form complex conduit networks for rapid groundwater flow. We did not find any relationship between aquifer lithology and the size or nature of the earthquake-induced water-level changes.

The M_w 5.8 earthquake occurred on 23 August at 17:51:04 UTC (Coordinated Universal Time). All of the wells have a post-earthquake water-level measurement 9 min after the origin time at 18:00 UTC. (The well sampled every 5 min has an additional measurement 4 min after the earthquake.) The groundwater-level offsets in 40 of the wells had begun by the time of the first sampling following the earthquake. In the remaining six wells, 158–553 km from the epicenter, no water-level offset was recorded until the next sample after 18:00 UTC. S-waves reached the farthest wells that responded within 4 min, and at those locations significant ground motion lasted <2 min.

No aftershocks were recorded in the 4 or 9 min between the M_w 5.8 main shock and the first postearthquake water-level measurements, implying that all of the groundwater-level offsets were initiated by deformation or seismic waves from the main shock. The largest aftershock was an M_w 3.9 event on 25 August, 05:07 UTC; no measurable groundwater-level offset occurred in response to that, or any of the other M2 to M3 aftershocks that continued to occur through May 2013.

Of the 46 wells with distinct offsets, groundwater levels dropped in 34 and rose in 12 (Table 1). Earthquake-related offsets were more rapid than other groundwater-level variations in the ~10 day period around the earthquake in which no changes were caused by rainfall. All but seven of the offsets were evident as of the first sample following the earthquake, and all had clearly begun within 1 h after that. We most confidently attribute the 36 pronounced offsets of 0.03 m or more to the earthquake, although seasonal and/or rainfall-induced water-level variations are large compared to the earthquake-induced change in every well. Smaller offsets in 10 of the wells started at most 1 h after the earthquake and were the most rapid changes within about a 10 day period, but were less distinct from background variations and closer to the resolution of the measurement.

There is no large-scale spatial pattern of rising versus falling water levels (Fig. 1), and the amplitudes of the groundwater-level offsets did not decrease monotonically with epicentral distance (Fig. 2). The largest water-level rises were 0.152 m in two wells: in Fauquier County, Virginia, 106 km north of the epicenter, and in Fulton County, Pennsylvania, 210 km north of the epicenter. The largest recorded water-level drop was −0.47 m at the Webster County, West Virginia, well, 238 km west-northwest of the epicenter.

Many of the well hydrographs resumed their pre-earthquake trends within a few hours or days after the earthquake. For the other wells, we cannot determine whether the groundwater-level drops were temporary or permanent. For example, in the Snyder County, Pennsylvania, well, 313 km north-northeast of the epicenter, the large groundwater-level rises caused by rainfall from May through November (Fig. 3A) make it impossible to determine whether the much smaller water-level rise caused by the earthquake recovered. Nevertheless, the earthquake-induced change is quite distinct in data for the month of August 2011 (Fig. 3B). Water level in the Snyder County well responds to Earth tides, implying that crustal strain can affect the water level. However, as we show in the following, permanent strain estimated from an elastic dislocation model of the rupture was too small to change water level measurably outside the area shown by the box in Figure 1. Figure 2 shows that neither the sizes nor the signs of the water-level offsets appear to depend upon whether the well responds to Earth tides.

It is possible, although by no means certain, that the water-level offsets in the Juniata County, Pennsylvania, well, 275 km north-northeast of the epicenter, and in the Webster County well, may not have recovered as of June 2013. In both of those wells, the seasonal high water levels in early 2012 and early 2013 are lower than those in the previous 4 yr by about the same amount as the earthquake-induced water-level drops (Figs. 4A, 4B, 5A, and 5B). In the Juniata County well, the water level following the earthquake was lower than at any time since October 2007. In the Webster County well, water level continued to fall for a longer period of time (14 days) than in any of the other wells, contributing to the large net offset and possibly indicating a longer lasting change to aquifer properties. Although the duration of a water-level offset is of interest because it may help diagnose the cause, the available data are insufficient to estimate all of the durations accurately.

Water-Level Offsets and Ground Motion

Seismic waves are the only means by which the earthquake could have affected groundwater levels over such a broad region.

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TABLE 1. GROUNDWATER OBSERVATION WELLS AFFECTED BY THE M_w 5.8 MINERAL, VIRGINIA, EARTHQUAKE OF 23 AUGUST 2011

USGS site number*	Location[†]	Depth (m)	Elevation (m)	Open hole (m)	Lithology	Topographic setting	Water level offset (m)	Sampling interval (minutes)	Delay?	Time of maximum excursion (UTC)	Earth tide response?	Distance (km)	Azimuth (degrees)	ZIP code	DYFI intensity[§]
382008080292801	Webster County, WV (WB)	24.4	944.8	6.1	Sandstone (50%), with silt and shale	Ridgetop	–0.470	60	Yes	Sept. 7	Yes	238	–76	26206	3.9
381002078094201	Orange County, VA (OR)	29.9	146.3	26.5	Schist, marble, and greenstone	Valley	–0.360	15	No	Aug. 23, 19:00	No	39	–28	22942	6.0
404745075184001	Northampton County, PA (NH)	66.4	176.2	51.2	Shale	Gentle terrain	–0.230	60	Yes	Aug. 24, 06:00	No	395	36	18091	4.9
372150079422301	Bedford County, VA (BF1)	61.3	326.1	36.4	Piedmont and Blue Ridge crystalline-rock aquifer	Hillside	–0.210	15	No	Aug. 23, 18:30	Yes	169	253	24174	5.0
402411077374801	Juniata County, PA (JU)	33.5	193.5	28.0	Siltstone grading downward to shale	Valley	–0.210	60	No	Aug. 24, 02:00	Yes	275	6	17051	4.0
421157075535401	Castle Creek, NY (CA)	76.8	337.0	61.0	Mudstone (Schieber, 1999)	Hillside	–0.200	15	No	Sept. 1, 03:00	Yes	500	21	13901	3.6
383423077245901	Prince William County, VA (PW)	149.4	89.9	134.1	Piedmont and Blue Ridge crystalline-rock aquifer; "Cambrian System" (370CMBR)	Gentle terrain	–0.150	15	No	Aug. 23, 22:00	No	92	32	22172	6.1
370812080261901	Christians-burg, VA (CB)	137.2	603.5	125.3	Carbonate fault breccia	Gentle terrain	–0.140	5	No	Aug. 24, 02:00	Yes	242	253	24073	4.1
394024078273401	Allegany County, MD (AL)	34.9	219.5	N.D.[#]	Laminated shale and siltstone	Valley	–0.100	30	No	Aug. 24, 09:00	Yes	200	–13	21766	3.8
385849079563901	Barbour County, WV (BB)	55.5	537.7	49.4	Sandstone	Hillside	–0.090	15	No	Aug. 26, 12:00	No	218	–55	26250	3.8
395322077365301	Franklin County, PA (FR)	61.6	231.6	50.3	Interbanded and interlaminated limestone and dolomite	Gentle terrain	–0.090	60	No	Aug. 23, 23:00	No	220	8	17202	4.7
401843078075401	Huntingdon County, PA (HU)	32.0	295.7	26.5	Sandstone and minor siltstone	Valley	–0.090	60	No	Aug. 24, 10:00	Yes	264	–3	16657	5.2
401804074432601	Mercer County, NJ (ME)	61.0	37.6	45.7	Predominantly arkosic sandstone	Gentle terrain	–0.090	15	No	Aug. 24, 00:00	Yes	392	47	8648	4.1
385638077220101	Fairfax County, VA (FF)	62.5	120.4	51.8	Early Mesozoic basin aquifer; "Manassas Formation" (231MNSS)	Gentle terrain	–0.080	15	No	Aug. 23, 18:00	Yes	129	25	20192	4.6
394655080014301	Greene County, PA (GR)	31.7	304.8	25.8	Cyclic sequences of sandstone, shale, limestone, and coal	Valley	–0.080	60	No	Aug. 24, 02:00	No	279	–41	15349	4.3
400217078281901	Bedford County, PA (BE)	45.7	353.6	31.4	Limestone	Valley	–0.070	60	No	Aug. 23, 19:00	No	239	–11	15537	3.9
414737073563301	Hyde Park. NY (HY)	38.4	51.8	15.5	Limestone	Ridgetop	–0.070	60	Yes	Aug. 24, 19:00	Yes	553	39	12538	4.0
373146078161201	Cumberland County, VA (CU)	61.6	97.5	45.9	Piedmont and Blue Ridge crystalline-rock aquifer; "Paleozoic-Precambrian Erathems" (300PZPB)	Valley	–0.061	15	No	Aug. 24, 00:00	Yes	44	221	23040	5.0
372053078493801	Appomattox County, VA (AP)	91.4	236.2	78.5	Piedmont and Blue Ridge crystalline-rock aquifer; Cambrian System" (370CMBR)	Gentle terrain	–0.061	15	No	Aug. 24, 01:30	Yes	96	237	24522	4.5

(*Continued*)

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TABLE 1. GROUNDWATER OBSERVATION WELLS AFFECTED BY THE M_W 5.8 MINERAL, VIRGINIA, EARTHQUAKE OF 23 AUGUST 2011 (*Continued*)

USGS site number*	Location†	Depth (m)	Elevation (m)	Open hole (m)	Lithology	Topographic setting	Water level offset (m)	Sampling interval (minutes)	Delay?	Time of maximum excursion (UTC)	Earth tide response?	Distance (km)	Azimuth (degrees)	ZIP code	DYFI intensity§
392259077052401	Carroll County, MD (CR)	75.6	167.6	69.2	Piedmont and Blue Ridge crystalline-rock aquifers (N400PDMBRX) national aquifer	Valley	–0.061	30	No	Aug. 24, 05:30	Yes	182	25	21797	5.2
392045076512501	Baltimore County, MD (BA)	76.2	149.2	54.5	Biotite-quartz monzonite	Gentle terrain	–0.061	30	No	Aug. 23, 21:30	Yes	189	32	21163	5.1
411424077462201	Clinton County, PA (CC)	22.9	624.8	11.3	Sandstone and siltstone	Valley	–0.061	60	No	Aug. 24, 02:00	No	363	3	16822	3.2
415323077451301	Susquehanna County, PA (SQ)	53.3	387.1	29.0	Sandstone, siltstone, shale, and mudstone	Valley	–0.055	60	No	Aug. 23, 19:00	Yes	433	2	16948	2.5
384821078472301	Shenandoah County, VA (SH)	59.4	407.5	42.7	Laminated shale and siltstone	Reservoiroutlet	–0.046	15	No	Aug. 24, 12:00	Yes	128	–36	22810	3.9
414159079213601	Warren County, PA (WR)	32.0	368.8	18.0	Sandstone, siltstone, shale, and mudstone	Valley	–0.037	60	No	Aug. 23, 18:00	No	431	–16	16351	2.2
390623077314201	Loudoun County, VA (LO)	163.1	121.9	161.2	Siltstone	Gentle terrain	–0.030	15	No	Aug. 24, 18:00	Yes	140	16	20175	4.7
391308081064201	Ritchie County, WV (RI)	18.4	256.0	9.3	Sandstone and siltstone, some shale, limestone, and coal	Valley	–0.030	30	No	Aug. 23, 18:00	Yes	322	–62	26337	3.3
411223075234901	Monroe County, PA (MO)	29.9	606.6	11.9	Sandstone	Gentle terrain	–0.030	30	Yes	Aug. 23, 18:30	No	426	31	18466	3.1
370604082403901	Wise County, VA (WI)	60.5	507.5	46.2	Shale, with siltstone, sandstone, and coal	Valley	–0.030	15	No	Aug. 24, 07:45	Yes	444	262	24293	3.9
414513077333701	Tioga County, PA (TI)	23.8	399.3	3.4	Sandstone, siltstone, shale, and mudstone	Valley	–0.021	60	No	Aug. 23, 19:00	No	419	5	16922	4.9
413428074085701	Montgomery County, NY (MT)	14.9	125.0	6.7	Shale and argillite with siltstone	Gentle terrain	–0.015	60	Yes	Aug. 24, 01:00	Yes	524	39	12586	3.0
373758079271601	Rockbridge County, VA (RB)	213.4	227.1	182.6	Shale and siltstone, with dolomite and limestone	Valley	–0.015	15	No	Aug. 24, 13:00	Yes	140	262	24555	5.0
410538080280801	Lawrence County, PA (LR)	45.7	317.0	35.7	Sandstone	Gentle terrain	–0.009	60	No	Aug. 23, 22:00	Yes	413	–32	16143	2.0
401157075032001	Bucks County, PA (BU)	120.4	112.8	103.0	Arkosic sandstone	Gentle terrain	–0.005	30	No	Aug. 23, 18:30	No	363	45	18974	3.9
364218078015701	Brunswick County, VA (BW)	135.6	116.5	110.3	Piedmont and Blue Ridge crystalline-rock aquifer; "Paleozoic-Precambrian Erathems" (300PZPB)	Gentle terrain	0.012	15	No	Aug. 23, 18:00	Yes	120	183	23920	4.9
414330076280501	Bradford County, PA (BR)	35.1	228.6	18.3	Mudstone, siltstone, sandstone, and thin conglomerate	Valley	0.012	60	No	Aug. 23, 20:00	Yes	436	17	18848	4.4
390948078001601	Clarke County, VA (CK)	182.9	197.5	N.D.#	Dolomite, limestone, and chert	Gentle terrain	0.015	15	No	Aug. 23, 18:00	Yes	141	–2	22611	4.3
402735077100901	Perry County, PA (PY)	45.7	150.9	40.2	Valley and Ridge aquifers (N500VLYRDG) national aquifer	Valley	0.015	60	No	Aug. 23, 18:00	Yes	288	14	17074	3.2

(*Continued*)

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USGS site number*	Location†	Depth (m)	Elevation (m)	Open hole (m)	Lithology	Topographic setting	Water level offset (m)	Sampling interval (minutes)	Delay?	Time of maximum excursion (UTC)	Earth tide response?	Distance (km)	Azimuth (degrees)	ZIP code	DYFI intensity§
412739080104201	Mercer County, PA (MR)	36.6	399.3	27.4	Siltstone and shale with interbedded sandstone	Gentle terrain	0.037	60	Yes	Aug. 25, 00:00	Yes	434	-26	16114	2.2
391920078032201	Berkeley County, WV (BK)	92.0	175.1	80.6	Valley and Ridge aquifers (N500VLYRDG)	Gentle terrain	0.040	60	Yes	Aug. 24, 04:00	No	158	–3	25413	5.1
422702079005101	Perrysburg, NY (PB)	106.7	487.7	83.2	Siltstone and shale	Gentle terrain	0.046	15	No	Aug. 25, 10:00	No	502	–10	14070	2.6
372224079423601	Bedford County, VA (BF2)	30.8	283.5	15.2	Piedmont and Blue Ridge crystalline-rock aquifers (N400PDMBRX)	Valley	0.052	15	No	Aug. 23, 19:45	Yes	169	254	24174	5.0
372322081241501	McDowell County, WV (MC)	46.3	622.1	40.7	Sandstone, approx. 50%, with some shale, siltstone, and coal	Valley	0.052	60	No	Aug. 24, 05:00	Yes	322	263	24868	3.9
403939076591001	Snyder County, PA (SN)	30.5	225.6	18.3	Siltstone, mudstone, and sandstone interbedded with minor siltstone	Valley	0.076	60	No	Aug. 23, 21:00	Yes	313	16	17853	3.1
384957077481701	Fauquier County, VA (FQ)	106.5	175.0	100.4	Metabasalt breccia	Gentle terrain	0.152	15	No	Aug. 23, 19:00	Yes	106	8	20198	4.6
394755078135001	Fulton County, PA (FU)	37.2	231.6	31.7	Sandstone, siltstone, shale, and mudstone	Valley	0.152	60	No	Aug. 25, 21:00	No	210	–7	17267	4.0

*The latitude and longitude (in degrees, minutes, and seconds) are encoded in the first 13 digits of the USGS site number. For example, the coordinates of site 382008080292801 are 38°20′08″ N, 80°29′28″ W.

†The two letters in parentheses at the end of each location column entry give the abbreviation used for that site on the map in Figure 1.

§The "DYFI intensity" was obtained from the USGS "Did-you-feel it" website (see Appendix).

#N.D.—not determined.

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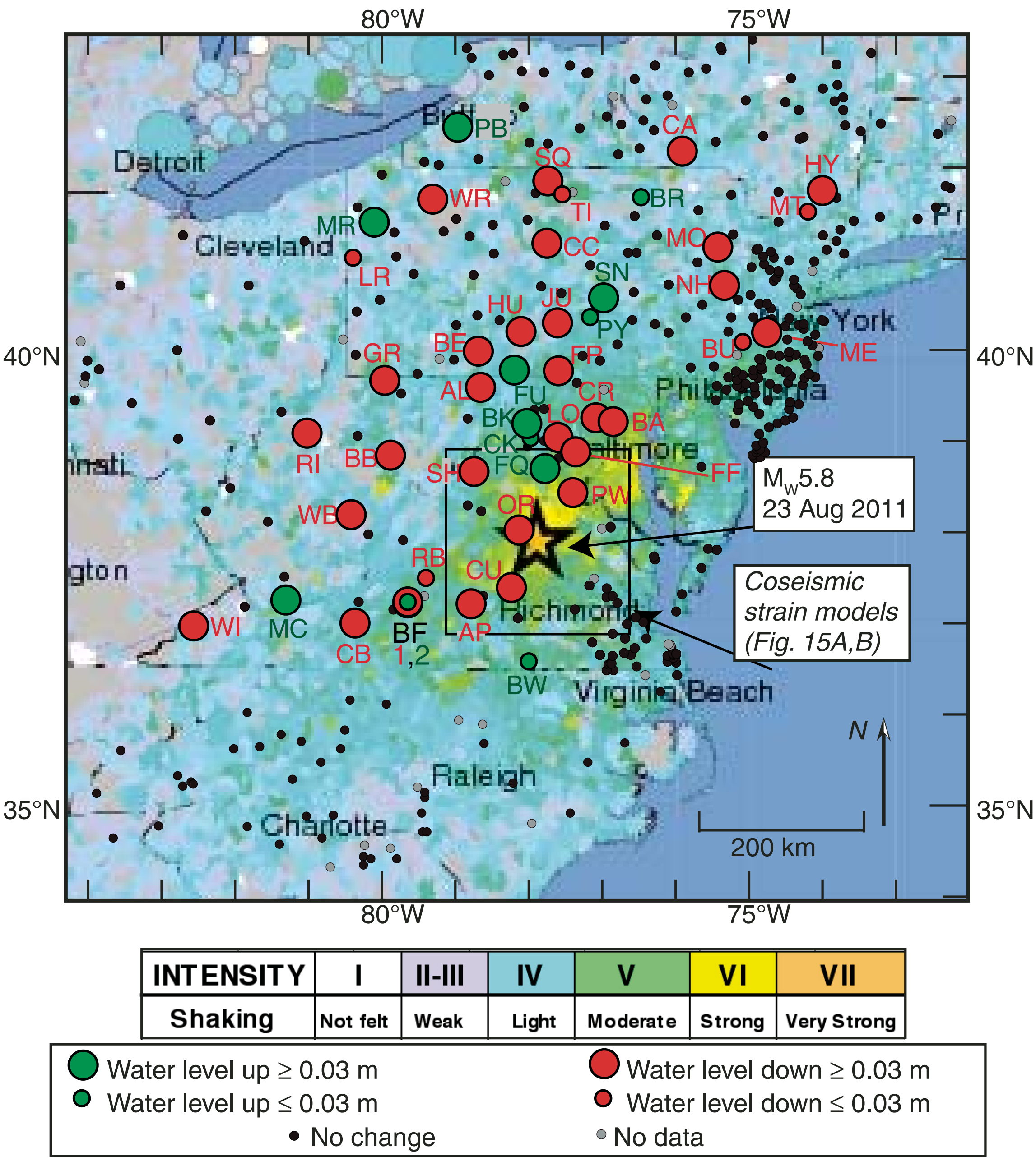

Figure 1. Map showing the locations of U.S. Geological Survey (USGS) monitoring wells (see footnote 1) superimposed on a map of intensities derived from reports submitted to the USGS "Did You Feel It?" website (http://earthquake.usgs.gov/earthquakes/dyfi/). Wells in which earthquake-induced offsets occurred are in red and green. The yellow star denotes the epicenter of the 23 August 2011 M_w 5.8 Mineral, Virginia, earthquake. A rectangle shows the area in which coseismic static areal strains were calculated from an elastic dislocation model of the rupture; outside this rectangle those calculated strains are too small to change water levels measurably.

[1]GSA Data Repository Item 2014365, List of USGS station numbers of wells shown in Figure 1, is available at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

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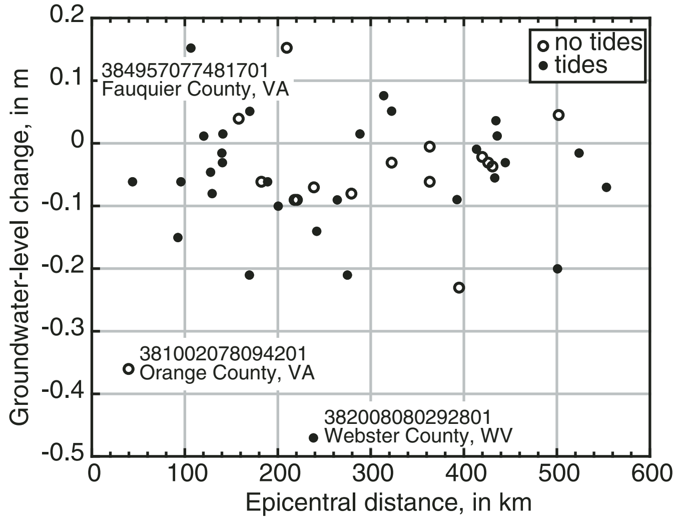

Figure 2. Graph of the amount of water-level offset versus distance of the well from the epicenter of the M_w 5.8 Mineral, Virginia (VA), earthquake. WV—West Virginia.

Within a few tens of kilometers of the epicenter, coseismic static strain may have been large enough to affect groundwater levels, but if ground motion from seismic waves affects distant wells, its influence in the near field also must be considered.

The spatial distribution of groundwater-level changes resembles that of earthquake intensity (Fig. 1), which can, in turn, be related to ground acceleration or velocity (Atkinson and Wald, 2007). Like the locations that experienced intensity 5 or greater on the DYFI scale, the groundwater-level changes extend further northeast than southwest of the epicenter. The northeast fault strike and rupture directivity enhanced accelerations northeast of the epicenter, with critical body-wave reflections from the Moho accounting for the largest accelerations observed further than 100 km to the northeast (Catchings et al., 2012). To the southwest, the spatial distributions of intensities and affected wells both generally follow the trend of the Appalachian orogenic front, which may reflect topographic amplification of ground motion.

Figure 6, a graph of the DYFI intensities assigned to each U.S. Postal Service ZIP code for which an online report was submitted, shows that the intensities generally decrease with distance, but that a range of intensities was experienced at all distances. All but three of the water-level offsets were in ZIP codes assigned DYFI intensity 3 or greater, although some intensity 5 reports were received throughout the distance range in which offsets were observed. Figure 7 is a graph showing the sizes of the water-level offsets versus the DYFI intensity for the nearest ZIP code. With one exception, the maximum absolute offsets increase with intensity for both rises and drops. The exception, in Webster County, is the well with the largest recorded water-level drop. It is also the well in which the water level continued to drop for the longest period of time (until 7 September 2011), and would plot within the range spanned by the other water-level offsets if only the change through 26 August were considered. The well nearest the epicenter, in Orange County, Virginia, experienced intensity 6. The water-level offset could be viewed as consistent with extrapolation of the largest offsets to this intensity, although it could also be

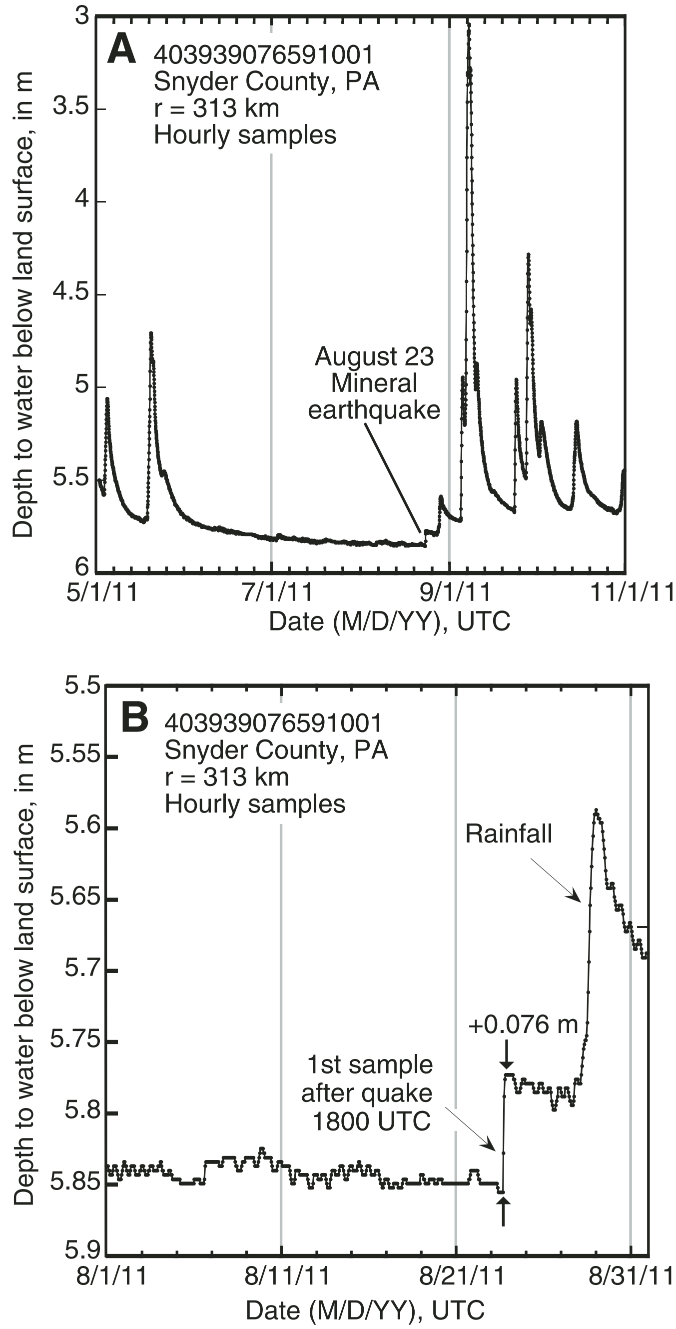

Figure 3. Depth to water in the Snyder County, Pennsylvania, well. (A) 1 May to 1 November 2011. (B) 1–31 August 2011. UTC—Coordinated Universal Time. r—distance from the well to the epicenter of the M_w 5.8 Mineral, Virginia, Earthquake.

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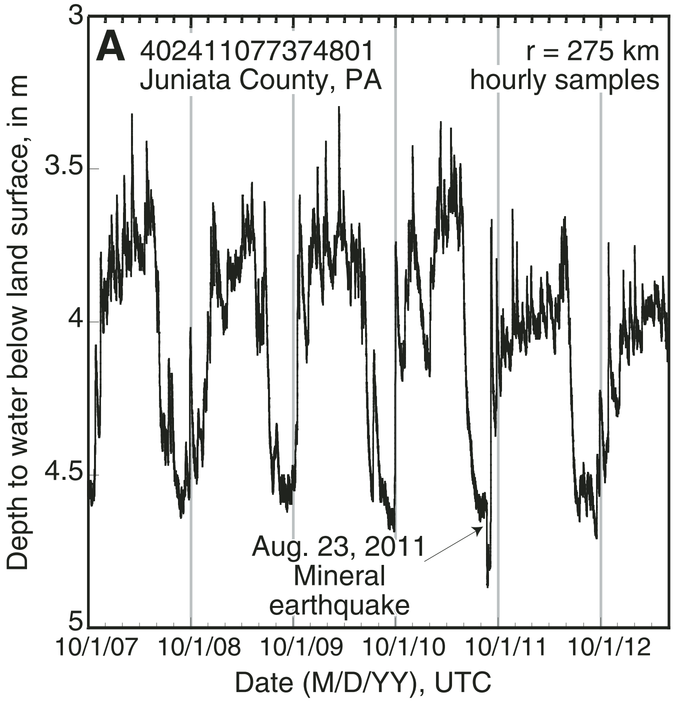

Figure 4. Depth to water in the Juniata County, Pennsylvania, well. (A) 1 October 2007 to 1 June 2013. (B) 1–31 August 2011.

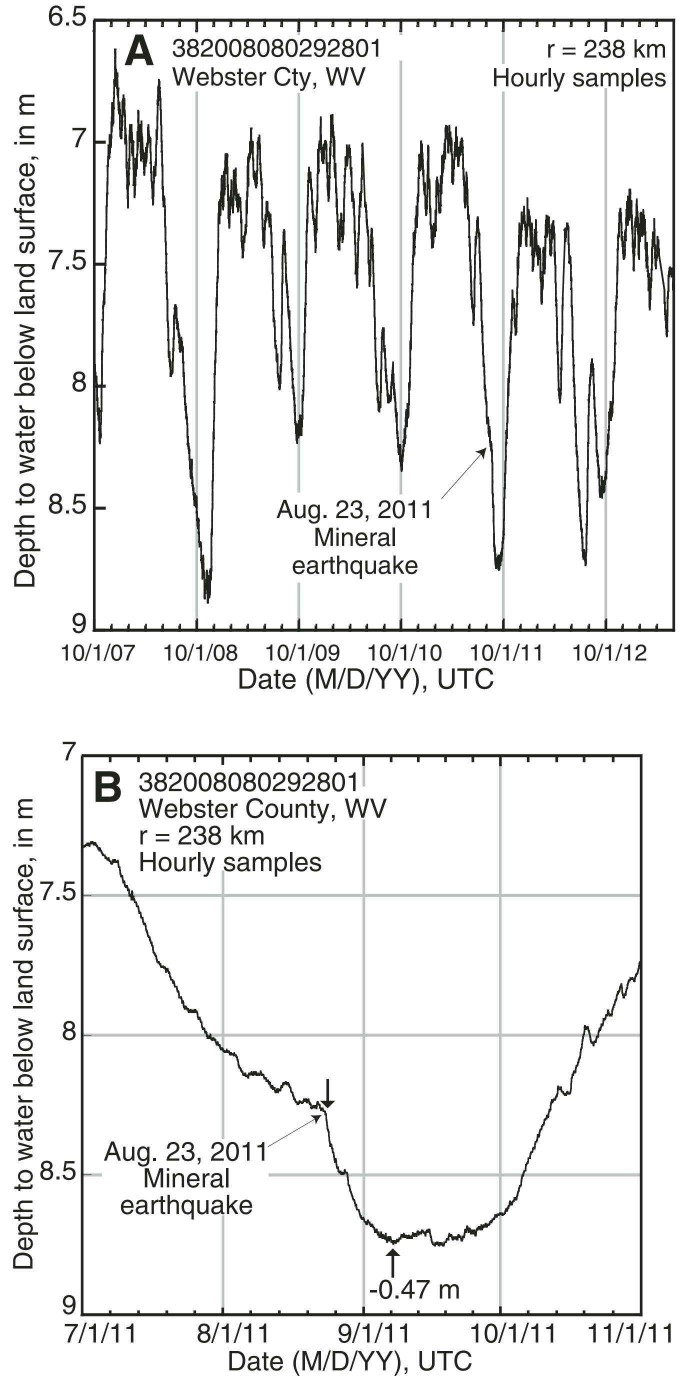

Figure 5. Depth to water in the Webster County, West Virginia, well. (A) 1 October 2007 to 1 June 2013. (B) 1 July to 1 November 2011.

viewed as enhanced by effects local to the near field, such as strain, as discussed herein.

Figure 7 shows only that the largest groundwater-level change at any intensity increases with intensity. The converse is not true, because at all intensities, a range of offsets occurred and some wells were unaffected. The relationship between the groundwater-level offsets and the intensity distribution implies that oscillatory ground motion, lasting at most a few minutes, caused groundwater levels in bedrock aquifers to offset and stay changed for up to two weeks. This in turn implies that the seismic waves affected these aquifers, for example by causing strain, and/or by changing permeability. Because the wells are

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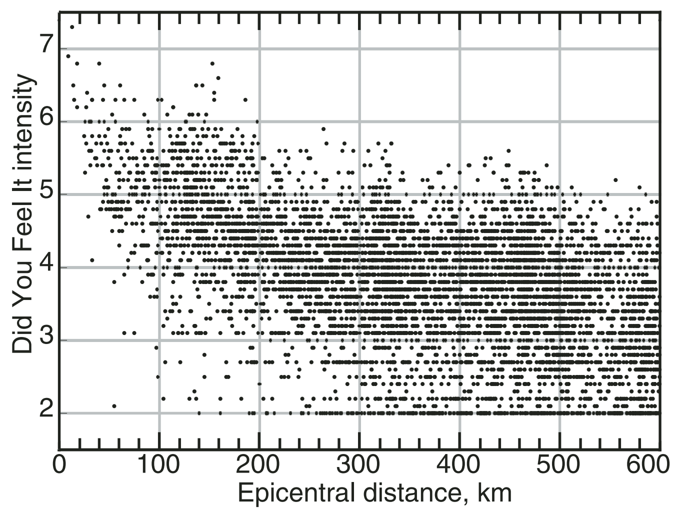

Figure 6. Graph of "Did You Feel It?" (U.S. Geological Survey; http://earthquake.usgs.gov/earthquakes/dyfi/) intensities by U.S. Postal Service ZIP code, versus epicentral distance of that ZIP code.

in different aquifers and in different hydrogeologic settings, the water-level offsets would not be expected to be functions only of intensity.

Aquifer Properties and Water-Level Offsets

The relationships among aquifer properties and time-varying groundwater level can be described by the transient equation of groundwater flow (e.g., see Domenico and Schwartz, 1990). For one-dimensional flow in a uniform porous medium, that equation is

$$\kappa(\rho g/\mu)\frac{\partial^2 h}{\partial x^2} = S_s \frac{\partial h}{\partial t}, \tag{1}$$

in which κ is permeability (dimensions of L^2 [L for length]), ρ is fluid density, g is acceleration due to gravity, μ is fluid dynamic viscosity, and h is elevation of the water surface above a datum (also referred to as hydraulic head). A change Δh in water level represents a change in fluid pressure of $\Delta p = \rho g \Delta h$. S_S is the specific storage (dimensions 1/L); x is distance in the direction of flow, and t is time. The specific storage can in turn be expressed as

$$S_S = \rho g(n\beta_w + \beta_p), \tag{2}$$

where n is porosity, β_w is fluid compressibility, and β_p is compressibility of the bulk aquifer. The properties in Equations 1 and 2 that could be affected by seismic waves are permeability, porosity, and/or aquifer compressibility.

A porosity change large enough to significantly affect groundwater level by acting on the specific storage would probably be detected because it would produce a very large direct change in water level. For example, if pre-earthquake porosity is 10%, a seismic-wave–induced increase to 11% represents aquifer strain of ~10^{-2}. The specific storage would change by <1%. In contrast, hydraulic head in a confined aquifer is typically changed by volumetric strain with a ratio of 0.3 m/microstrain (1 microstrain = 10^{-6}; e.g., see Roeloffs, 1996), so this modest porosity increase would drop water level by many tens of meters.

Damage such as breakage of grain contacts could increase aquifer compressibility, but these changes are likely very small. Elastic modulus changes would be reflected in seismic-wave velocity changes, which have been estimated as being only a few percent in the immediate vicinity of faults recently ruptured by large earthquakes (e.g., see Peng and Ben Zion, 2006). Like porosity, increases in aquifer compressibility would be too small to affect groundwater levels by increasing the specific storage.

Permeability increases with porosity, with especially large changes in low-permeability crystalline rocks, but also in open-structure sedimentary rocks such as sandstones (e.g., see Doyen, 1988). Permeability is therefore expected to increase rapidly with volumetric strain. However, permeability can also be increased without accompanying deformation if seismic waves mobilize suspended material blocking pore channels or break weakly cemented grain contacts. Permeabilities of rocks range over many more orders of magnitude than porosity or compressibility, and therefore permeability is the most likely property in Equation 1 to have been affected enough by ground motion to change water levels measurably. However, Equation 1 shows that a permeability change can lead to a time-dependent water-level change only if there is a nonzero spatial hydraulic gradient.

The preceding discussion does not strictly apply to aquifers where water flows primarily in discrete conductive fractures. Of

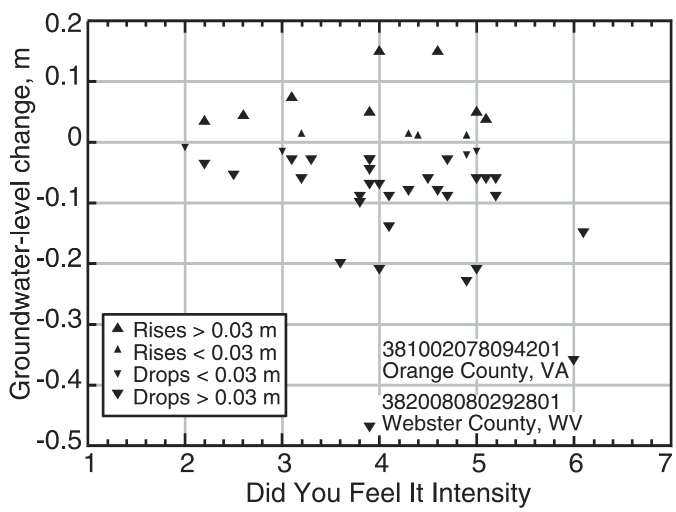

Figure 7. Graph of the amount of water-level offset versus "Did You Feel It?" (U.S. Geological Survey; http://earthquake.usgs.gov/earthquakes/dyfi/) intensity for the U.S. Postal Service ZIP code coordinates nearest each well. VA—Virginia, WV—West Virginia.

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the wells in Table 1, many are in crystalline rock aquifers that bear water in fractures, and flow in sedimentary aquifers may be dominated by bedding-plane fractures. Each discrete fracture has its own transmissivity, that is, the ability to transport a volume rate of water per unit pressure gradient, and may be hydraulically connected to a zone of distinct fluid pressure (e.g., see Johnson et al., 2005). In a well that intersects two or more such conductive fractures, the well water level will be a weighted average of the hydraulic heads in the fractures, weighted by the relative transmissivities of the fractures. If seismic shaking changes the relative transmissivities of the fractures, the well water level will also change. Such transmissivity changes could be promoted by cyclic movement of water into and out of the wellbore that dislodges particulate matter clogging the fractures; this effect could be very local to the wellbore and would not occur were the wellbore not present. We discuss herein the possibility that this mechanism, or another mechanism that depends on the existence of the wellbore, can account for the earthquake-induced groundwater-level changes.

Although the volume of pore space or fractures affects aquifer fluid pressure directly, it is unlikely that the transient stresses accompanying seismic waves could result in enough residual strain to account for the entirety of the groundwater-level offsets. We describe herein calculations showing that even the permanent areal strain due to fault slip reached 10^{-7} only within (at most) a 75 km radius of the epicenter. Moreover, the Mineral earthquake changed groundwater levels in wells that presumably have low strain sensitivity, because they do not respond to Earth tides (Table 1).

PREVIOUS OBSERVATIONS AND STUDIES OF WATER-LEVEL OFFSETS CAUSED BY SEISMIC WAVES

Permeability changes caused by seismic waves have been hypothesized to be the reason that certain wells consistently exhibit sustained water-level offsets in response to earthquakes too small and/or distant to produce significant aquifer deformation (Rojstaczer et al., 1995; Roeloffs, 1998; Brodsky et al., 2003; Matsumoto and Roeloffs, 2003). Elkhoury et al. (2006) deduced that earthquakes repeatedly enhanced permeability near each of two wells in fractured granitic rocks, based on identification and analysis of phase shifts in Earth tide–induced water-level fluctuations, and found that the permeability increases were proportional to peak ground velocity.

The epicentral distance range over which the Mineral earthquake changed groundwater levels is comparable to that at which the wells in previous studies consistently respond. To investigate whether any of the wells in Table 1 have shown water-level offsets in response to large, distant earthquakes, we examined online data for the times of the three M > 8.5 earthquakes that have occurred globally since 2007 (M_w 8.8 Chile, 27 Feb 2010; M_w 9.0 Tohoku, Japan, 11 March 2011; M_w 8.6 off Sumatra, 11 April 2012). At two of the wells, significant downward offsets occurred at the arrival times of seismic waves from the 2011 M_w 9.0 Tohoku earthquake: −0.043 m in the Castle Creek, New York well (421157075535401, 500 km from the Mineral earthquake), and −0.036 m in the Susquehanna County, Pennsylvania, well (415323077451301, 433 km from the Mineral earthquake). Neither of these wells responded to the other M > 8.5 events. These wells are among the furthest wells at which the Mineral earthquake caused groundwater-level offsets. It is possible that, like the wells in the previously mentioned studies, these wells are unusually responsive to seismic waves.

Seismic shaking can increase permeability greatly in the near field (approximately one rupture length from the fault), which has been demonstrated by abundant stream-discharge increases, accompanied by hilltop water-level drops, in the 1989 M_w 6.9 Loma Prieta, California, earthquake (Rojstaczer and Wolf, 1992); the 1995 M_w 6.9 Hyogo-ken Nanbu (Kobe, Japan) earthquake (Koizumi et al., 1996); the 1998 m_b (body wave) 5.0 Pymatuning, Pennsylvania, earthquake (Fleeger et al., 1999), and the 1999 M_w 7.5 Chi-Chi, Taiwan, earthquake (Wang et al., 2004). These phenomena occurred in hilly or mountainous terrain and the streams and springs that drained them, and, for the 3 M > six examples, within the distance ranges of landsliding and liquefaction. After the Loma Prieta and Pymatuning earthquakes, the hilltop water-level drops lasted for years, and may have been permanent. For all of these earthquakes, it is plausible that damage to water-bearing formations, caused by shaking, was the mechanism of permeability increase.

In their study of the 1998 Pymatuning earthquake, Fleeger et al. (1999) used numerical modeling to support their hypothesis that the water-level drops on the ridge top, water-level rises along the base of the ridge, and discharge increases in rivers draining the ridge could be explained by order-of-magnitude increases in the vertical permeability of shale confining beds as deep as 100 m below the ridge top. They attributed increased permeability to widening of preexisting fractures by seismic shaking, such as subvertical stress-relief fractures on ridge margins where lateral support is absent. The combination of topographic amplification of ground motion with steep subsurface hydraulic gradients in the ridge appears to have led to large permeability increases that in turn caused large water-level drops.

Although increased permeability is consistent with most of the observations described in these studies, two sets of laboratory experiments measuring the effects on permeability of oscillatory stress and fluid pressure found opposite directions of permeability change. Liu and Manga (2009) found that application of an oscillatory axial load under undrained conditions almost always decreased permeability of laboratory-fractured sandstone samples. Adding small amounts of silt to the sample resulted in larger permeability changes. The conclusion was that mobilization of loose material, either the added silt or particles left from the fracture process, could clog pores and reduce permeability. However, Elkhoury et al. (2011) applied fluid pressure oscillations to laboratory-fractured sandstones under undrained conditions, and found that permeability invariably increased. Because

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the permeability gradually returned to the previous level, they argued for a reversible mechanism, again invoking mobilization of loose material in conductive pathways. Our observations of both positive and negative offsets in water levels may indicate decreased or increased permeability and are consistent with these laboratory experiments. However, the very large water-level offsets and discharge increases following the Loma Prieta, Kobe, Pymatuning, and Chi-Chi earthquakes indicate enhanced, not decreased, permeability.

Wang and Chia (2008) noted that even if the effect of seismic waves on permeability is always to increase it, both groundwater-level rises and drops can be explained depending on whether the permeability increase enhances the connection to a location of higher or lower permeability. It has been observed that wells that consistently exhibit sustained water-level offsets in response to earthquakes consistently change in the same direction, regardless of earthquake focal mechanism or direction relative to the well (Roeloffs, 1998; Brodsky et al., 2003; Matsumoto and Roeloffs, 2003; Matsumoto et al., 2003; Roeloffs et al., 2003). Wang and Chia (2008) attributed that feature to each well's fixed position along an established hydrologic flow path.

PERMEABILITY CHANGES AND HYDROLOGIC SETTING

Changing permeability can change groundwater level only where a pressure gradient exists. Topography typically controls pressure gradients. Precipitation infiltrates at ridges and hilltops, where pressure gradients favor downward flow, while in valleys and creek beds, gradients favor upward flow and groundwater discharge. Therefore, beneath local topographic highs, enhanced permeability would be expected to speed subsurface flow toward discharge areas, causing water levels to drop. Wells in valleys could have increased pressures if enhanced permeability occurs upgradient, as observed following the 1998 Pymatuning earthquake (Fleeger et al., 1999), but if permeability is increased between the well and a discharge area, a pressure drop accompanied by increased discharge could result. By this reasoning, enhanced permeability could cause a water-level drop in most wells, but could not cause a water-level increase in a well that is on top of a hill or ridge-top recharge area, because, absent active and constant recharge, there is no area of higher pressure to which it is connected.

To investigate whether enhanced permeability can explain the groundwater-level changes caused by the 2011 Mineral earthquake, we compared the directions of the groundwater-level changes with topography. Table 1 lists brief classifications of the topography around each wellhead; Appendix 1 explains how to access online topographic maps for all of the well locations. Of the 46 wells, 41 are in valleys or areas of low topographic relief, and water levels dropped in 36 of these. The five wells on hillsides or ridge tops, however, all exhibited water-level drops of 0.09 m or more. The largest water-level drop occurred in a ridge-top well (Webster County), and the most distant water-level drop (Hyde Park, New York, 553 km from the epicenter) was also in a well on a narrow ridge. Of the 12 wells in which water levels rose, six are in very gentle terrain and six are in valleys. This relationship with topography is consistent with the direction of the water level change being determined by whether connection to higher or lower pressure is enhanced, and further suggests that ridge tops may be especially susceptible to seismic-wave–enhanced permeability and subsequent water-level decreases.

If seismic ground motion enhances permeability, a further question is whether the enhancement occurs only near the monitoring well, or instead more broadly throughout an aquifer. Elkhoury et al. (2006) showed that the postearthquake phase shifts of Earth-tide–induced water-level fluctuations could be explained by permeability enhancement limited to 200 m from each well. Evidence for permeability changes throughout an aquifer would be best obtained with two or more wells located in the same basin at different elevations along a groundwater flow path.

Hydrographs from a pair of wells 1.2 km apart in Bedford County, Virginia, 170 km southwest of the epicenter, suggest that a permeability change affected much of a hillslope aquifer (Figs. 8A, 8B). Well 372224079423601 is in the valley of South Fork Goose Creek at an elevation of 283 m and is open 15 m below the creek bed; the Mineral earthquake caused the water level to rise in this well. Well 372150079422301 is due south at 326 m elevation, and is open to the formation below the level of the creek; in this well the earthquake produced a water-level drop (Figs. 9A, 9B). The earthquake-induced drop in the higher elevation well and smaller, slower rise in the valley well had both begun as of the measurement 9 minutes after the earthquake, consistent with increased permeability along all or parts of subsurface flow paths on both sides of the creek. The relationship between these two hydrographs is somewhat complex; for example, some rapid water-level rises in the higher elevation well are not reflected in the valley well, possibly because the wells are on opposite sides of the creek. Moreover, the 23 August Mineral earthquake is not the only time when water level fell in the higher elevation well and rose in the valley well; for example, this situation also occurred 3–6 September 2011, although those unexplained groundwater-level changes occurred a few days apart (Fig. 9A). Nevertheless, the responses of these two wells to the Mineral earthquake resemble a small-amplitude version of the dramatic effects of hillslope permeability increases observed for previous earthquakes (e.g., Fleeger et al., 1999).

WHAT FREQUENCY OF GROUND MOTION CHANGED THE GROUNDWATER LEVELS?

Previous studies (Roeloffs, 1998; Brodsky et al., 2003; Matsumoto and Roeloffs, 2003; Matsumoto et al., 2003; Roeloffs et al., 2003) have shown that certain wells consistently exhibit sustained water-level changes in response to earthquakes thousands of kilometers away. Short-period accelerations for these events are negligible, suggesting that the changes are caused by surface waves with periods of 10 s or more. In at least one of the wells,

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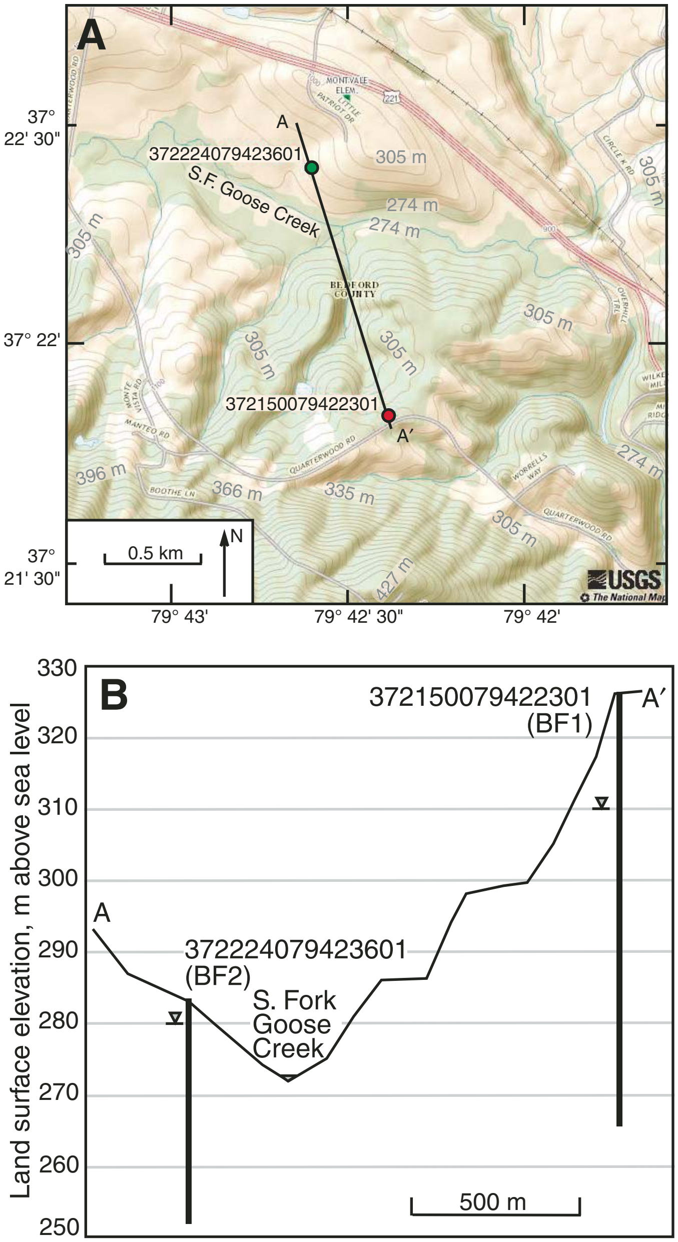

Figure 8. (A) Map showing the locations of the two Bedford County, Virginia, wells denoted as "BF 1,2" in Figure 1. The map symbol for BF 1,2 is larger than the area shown here. S.F.—South Fork. (B) Cross section showing topography and well and water-level depths of the two wells.

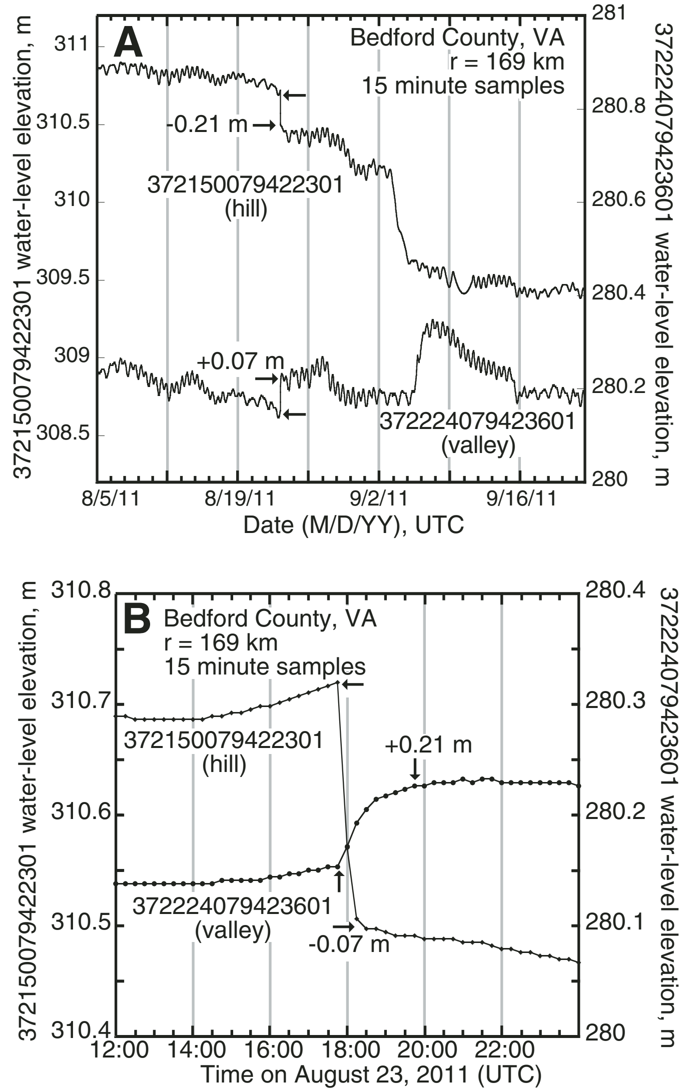

Figure 9. Depth to water in the two Bedford County, Virginia, wells. (A) 5 August to 23 September 2011. VA—Virginia. (B) 12 h approximately centered on the M_w 5.8 Mineral, Virginia, earthquake. UTC—Coordinated Universal Time.

however, water-level rises occurred in response to relatively close events smaller than M5 (Roeloffs, 1998), indicating that short-period ground motion, the predominant effect of the Mineral earthquake, can affect aquifer properties. Coseismic groundwater-level increases in the upper 10 m of a densely instrumented alluvial fan within 30 km of the fault ruptured by the 1999 M_w 7.5 Chi-Chi earthquake are consistent with causative ground motions having periods of 1 s or longer (Wang et al., 2003; Wong and Wang, 2007). The water-level offsets analyzed here, however, are in much deeper bedrock wells, include drops as well as rises, and occur over a much larger area.

In the Christiansburg, Virginia, well (370812080261901), the Mineral earthquake imposed changes distinct from the seismic water-level oscillations that occur in this well in response to shallow M > 6 earthquakes worldwide. Seismic oscillations in wells have been attributed to aquifer pressure cycles caused by strain carried by surface waves with periods of 10 s or more (e.g., Brodsky et al., 2003). Like the persistent offsets, only certain wells exhibit pronounced seismic oscillations, which require high strain sensitivity, high permeability, and resonant fluid movement in the well casing. We found that two other wells in Table 1 exhibited seismic oscillations, much smaller than those

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at the Christiansburg well, for all 3 of the M > 8.5 earthquakes since 2007: 364218078015701 (Brunswick County, Virginia), and 401804074432601 (Mercer County, New Jersey). Seismic oscillations have left little, if any, water-level offset in any of these three wells.

The Christiansburg well hydrograph for May through October 2011 (Fig. 10A) exhibits seismic oscillations in addition to water-level rises associated with rainfall. The well responds to Earth tides and atmospheric pressure, and these effects can be analyzed and removed in the period indicated in Figure 10A, a period relatively free of rainfall effects. Figure 10B shows the result of removing the tidal and atmospheric pressure effects, which makes the responses to the distant earthquakes quite evident. None of the responses to distant earthquakes have been followed by offsets of the groundwater level. Figure 11A shows the Christiansburg well response to the M_w 8.8 Maule, Chile,

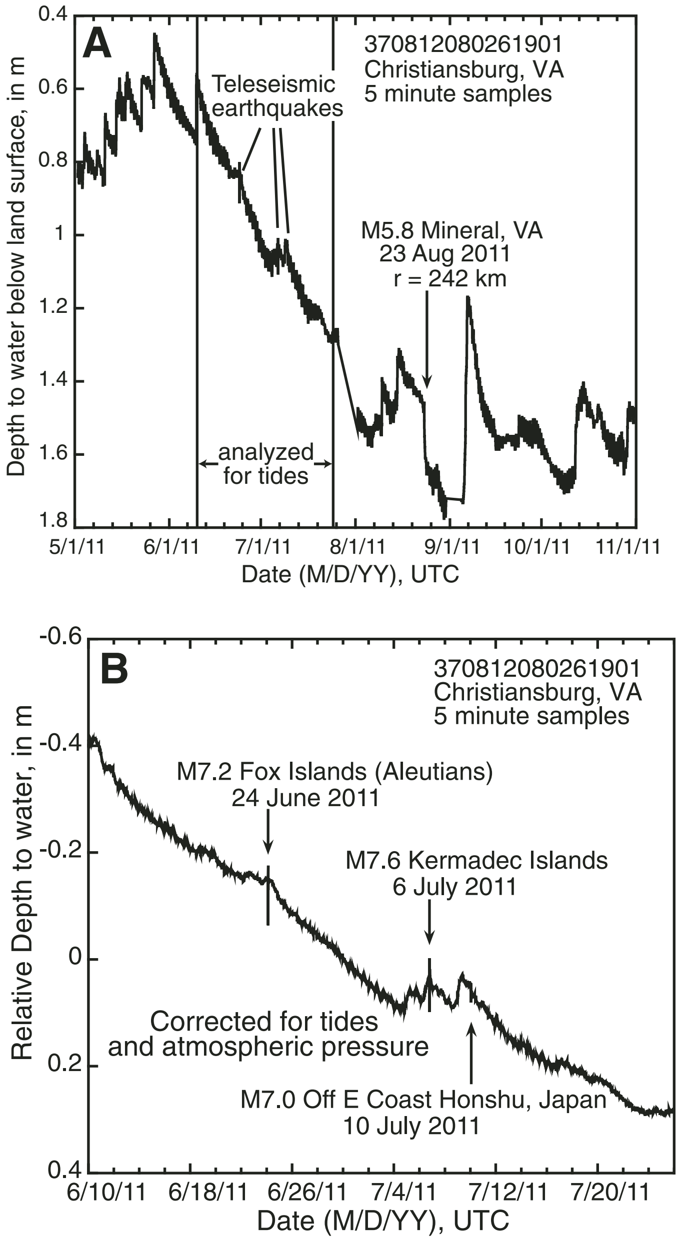

Figure 10. (A) Depth to water in the Christiansburg, Virginia (VA), well for May through November 2011. (B) Depth to water (relative) in the Christiansburg well for 10 June to 26 July 2011, with atmospheric pressure and earth-tide fluctuations removed.

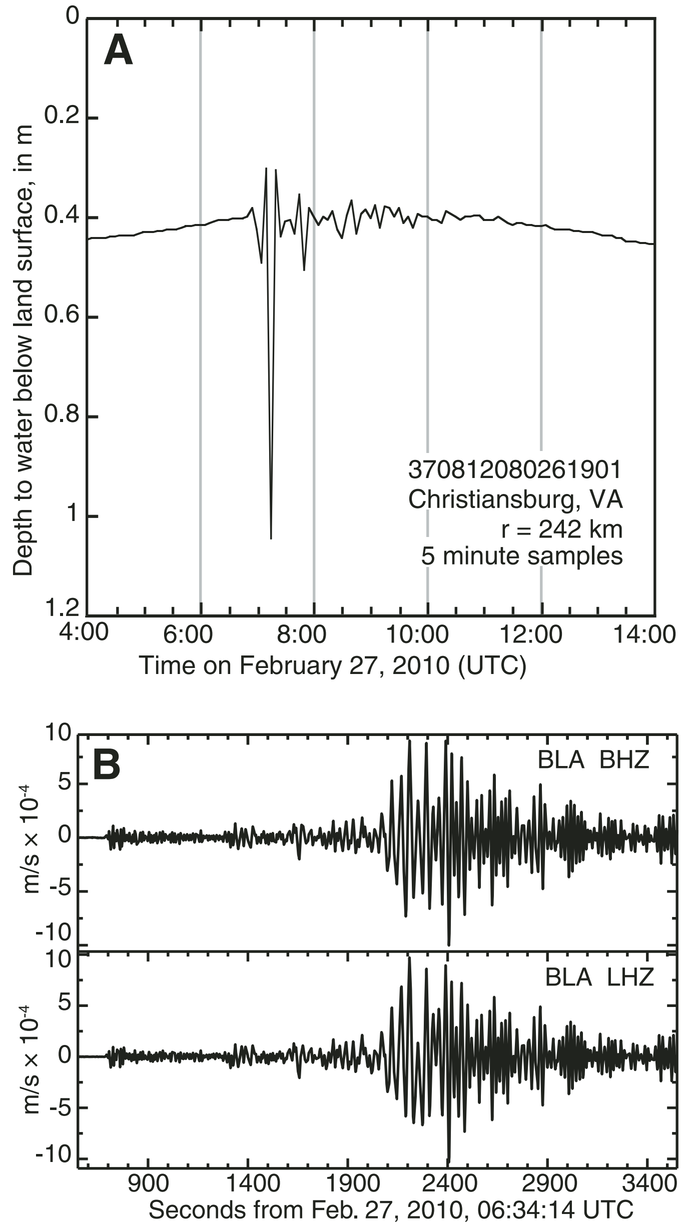

Figure 11. (A) Response of water level in the Christianburg, Virginia (VA), well to the M_w 8.8 Maule, Chile, earthquake of 27 February 2010. UTC—Coordinated Universal Time. (B) BHZ (40 samples per second) and LHZ (1 sample per second) vertical seismograms from broadband station BLA, 10 km north of the well.

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earthquake of 27 February 2010, for comparison with vertical seismograms from broadband station BLA, 10 km north of the well (Fig. 11B). For this distant earthquake, the LHZ seismogram recorded at one sample per second (sps) is essentially identical to the BHZ seismogram, recorded at 40 sps, demonstrating the lack of short-period shaking, as expected for an event so far away. Both seismograms contain large phases with periods of 10 s or more.

In contrast, no seismic water-level oscillations were observed for the M_w 5.8 Mineral earthquake in the Christiansburg well, or in any other well examined for this study. Oscillations in the other wells may have been missed because groundwater levels were sampled every 15, 30, or 60 min; however, the absence of oscillations more likely reflects an absence of pronounced long-period surface waves, which were not generated by the relatively small Mineral earthquake.

Instead, the Mineral earthquake caused groundwater-level offsets in the Christiansburg well (Fig. 12A), as well as in the other two wells that had exhibited seismic oscillations in the past. Water level in the Christiansburg well continued to drop for 12 h, while the BLA seismograms show that significant ground motion lasted <4 min. If the shaking had produced enough permanent aquifer strain to cause the water-level drop, these wells should have responded immediately, since seismic oscillations can only occur in wells with high permeability. Therefore, a permeability change in the aquifer, which could occur with little or no deformation, better explains the groundwater-level drops.

For the M_w 5.8 Mineral earthquake, the BHZ velocity seismogram (40 sps) from Blacksburg is more than twice as large as the 1 sps LHZ seismogram (Fig. 12B), indicating abundant short-period ground motion captured at the higher BHZ sampling rate, and neither seismogram shows any large phases with periods as long as 10 s. We conclude that aquifer properties at the Christiansburg well, most likely permeability, changed in response to short-period ground motion, and did not recover for at least 12 h after seismic shaking ended.

ROLE OF EARTHQUAKE EFFECTS LOCAL TO THE WELLBORE

Could the water-level changes reflect action of the seismic waves on the wells, rather than on the aquifers? Accumulated material in the wellbore wall (a "wellbore skin"), or in fractures intersecting the well, can impede flow between the formation and the wellbore, causing wellbore water-level changes to differ from formation (or fracture) pressure.

By dislodging a wellbore skin, seismic shaking could allow the well water level to equilibrate with formation pressure. Removing a wellbore skin would change the water level in the direction needed to equilibrate with increasing or decreasing formation fluid pressure, as observed for 31 of the wells in Table 1. For example, in the Perrysburg, New York, well, water level was rising when the earthquake occurred. It cannot be ruled out that this groundwater-level change could reflect the well water level equilibrating to formation pressure after seismic shaking dislodged small particles clogging the borehole wall.

Lacking detailed characterization of the fractures that may be intersected by any of these wells, it is difficult to assess the possible role of near-wellbore fracture-transmissivity changes caused by cyclic movement of water into and out of the well. However, in the Christiansburg well, numerous earthquakes worldwide have repeatedly caused water to flow across the wellbore wall, but even very large oscillations (e.g., the 1 m oscillations from the M_w 8.8 Chile earthquake in 2010) have not produced the persistent type of water-level offset caused by the Mineral earthquake.

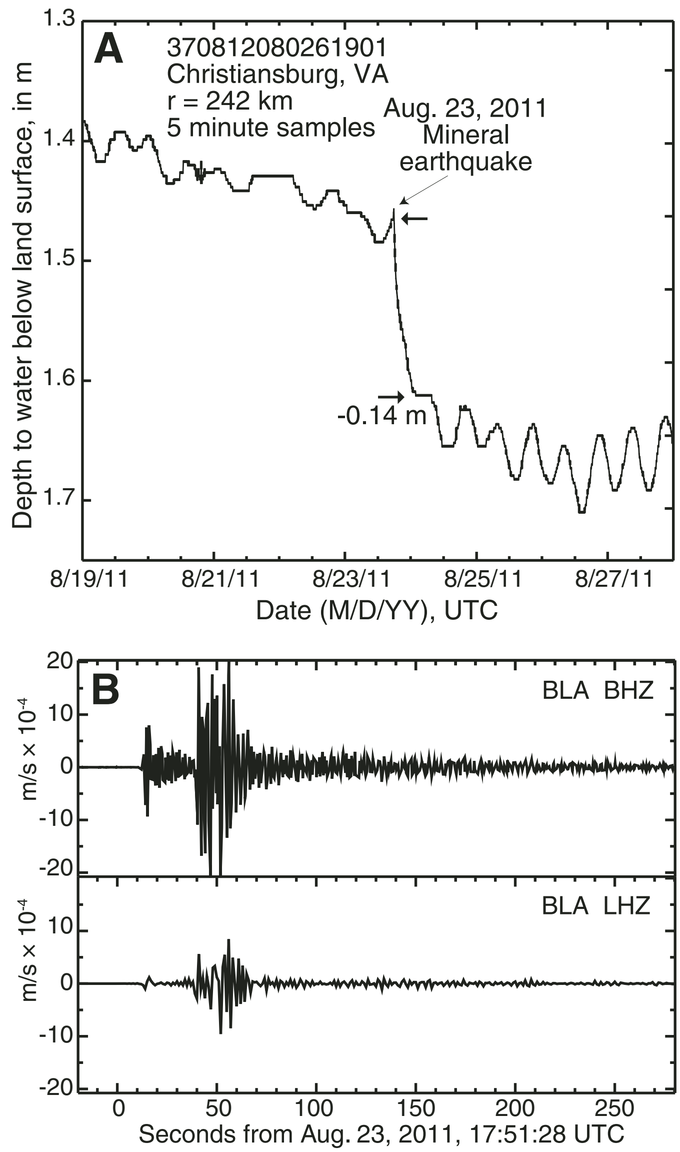

Figure 12. (A) Response of water level in the Christianburg, Virginia, well to the M_w 5.8 Mineral, Virginia (VA), earthquake of 23 August 2011. (B) BHZ (40 samples per second) and LHZ (1 sample per second) vertical seismograms from broadband seismic station BLA, 10 km north of the well. UTC—Coordinated Universal Time.

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The distance to which cyclic fluid motion extends from the wellbore would be expected to be smaller for the short-period waves dominating the Mineral earthquake ground motion than for the longer duration surface waves from large distant earthquakes.

Mechanisms that affect only the immediate vicinity of a well would not be expected to show a relationship between the direction of groundwater-level offset and hydrologic setting. To some extent, the water-level offsets caused by the Mineral earthquake are consistent with topographic and hydrologic settings, suggesting that they extend beyond the immediate vicinities of the wells. In particular, the two Bedford County wells discussed here demonstrate an effect of hydrologic setting on the direction of water-level offset.

Earthquake-induced groundwater-level changes reported in the literature have affected flow paths of 1 km or more (e.g., see Fleeger et al, 1999). Earthquake-induced spring-discharge changes have also been widely reported (e.g., see Rojstaczer and Wolf, 1992). An important area for further research is the extent to which seismic waves affect not just well water levels, but also larger scale aquifer properties.

HOW DOES GROUND MOTION ACT ON PERMEABILITY?

No direct observations support any single mechanism by which seismic ground motion could enhance permeability, and no single mechanism necessarily operates at all locations. Some proposed mechanisms, such as freeing of gas bubbles or loosening of colloidal material, could take place with little or no aquifer deformation, would not reduce the strength of the aquifer rock, and could be temporary. Other mechanisms, such as breaking of fragile intergranular cementation, constitute damage that weakens the rocks, even with little or no deformation, and may be permanent.

Buech et al. (2010) speculated that topographic amplification of short-period seismic ground motion damages rocks composing bedrock hills, based on data from seismometers at different locations on a small hill in New Zealand. The hill is 210 m high, 500 m wide, and 800 m long; the topographic relief is comparable to that at most of the well locations discussed here, and it is composed predominantly of sandstone, as are many of the aquifers affected by the Mineral earthquake. Studying relatively weak ground motions (from earthquakes ranging from M1.8 at 10 km to M7.4 at 1500 km), Buech et al. (2010) found the highest amplifications consistently occurred on the hill crest and for horizontal motion at frequencies near 5 Hz. Although they did not measure rock damage, they noted that the greatest amplification occurred near fields of broken bedrock. This study suggests that short-period weak motion may cause damage similar to, albeit less severe than, that caused by strong motion.

The idea that ground motion from the Mineral earthquake changed permeability by damaging rocks is consistent with the great extent of landslides and rockfalls induced by this event (Jibson and Harp, 2012); these occurred at maximum distances of 158 km to the northeast and northwest and 245 km to the southwest. Like the felt reports and groundwater-level offsets, the landslides and rockfalls occurred far beyond distance limits estimated from past earthquakes. It is reasonable to suppose that permeability changes from the Mineral earthquake extended beyond the limits of macroscopic ground failure.

The observed groundwater-level offsets require that seismic waves changed properties in bedrock aquifers at the depths of the wells (15–213 m), where peak ground velocities and accelerations are usually much lower than at the Earth's surface. Previous work has documented earthquake-induced aquifer-property changes at comparable depths. Bower and Heaton (1978, p. 338) investigated a groundwater-level offset in a well near Ottawa, Canada, caused by seismic waves from the 1964 Alaska earthquake, ~5200 km distant; they concluded that the groundwater level dropped due to volumetric expansion $>10^{-7}$ at depths below 200 m in the fractured limestone aquifer, expansion that would have had to last at least several days after the earthquake and must therefore have been "a nonlinear effect induced by strong ground motion." In Matsumoto and Roeloffs (2003) and Elkhoury et al. (2006) it was concluded, on the basis of subtle changes in well response to Earth tides, that seismic waves changed bedrock aquifer permeabilities as deep as 211 m. At the time of the Mineral earthquake, water levels in many of the affected aquifers were within 10 m of the surface, corresponding to effective overburden pressures between 0.2 and 3.3 MPa. It is plausible that low effective overburden pressure contributes to a subsurface aquifer being vulnerable to seismic ground motion.

PERMANENT STRAIN (NEAR FIELD) OVERWHELMED BY GROUND MOTION

Fault slip permanently deforms the Earth's crust, applying static strain that can change subsurface fluid pressure. Groundwater-level changes in wells that respond to Earth tides can be expressed as strain changes (e.g., Quilty and Roeloffs, 1997), possibly measuring deformation caused by the 2011 M_w 5.8 Mineral earthquake, for which no geodetic data exist. We have shown that the maximum groundwater-level offsets caused by the Mineral earthquake increased with seismic intensity, and that the offset in the Orange County well, closest to the epicenter, can be viewed as consistent with this trend. Nevertheless, the Orange County and Fauquier County water-level time series have features consistent with responses to coseismic static strain.

Figure 13 shows about four weeks of data for the Fauquier County well, spanning the time of the Mineral earthquake. The water level in this well rose to its maximum earthquake-related excursion in 1 h and remained at that level with no indication of recovery until 6 September, when rainfall led to a rapid rise of 1 m. The Fauquier County well has a strain response of 0.68 m/microstrain, based on tide analysis (Table 2). The rapid rise to a sustained level is the expected response of a confined aquifer to an abruptly applied contractional strain step (e.g., Roeloffs, 1996).

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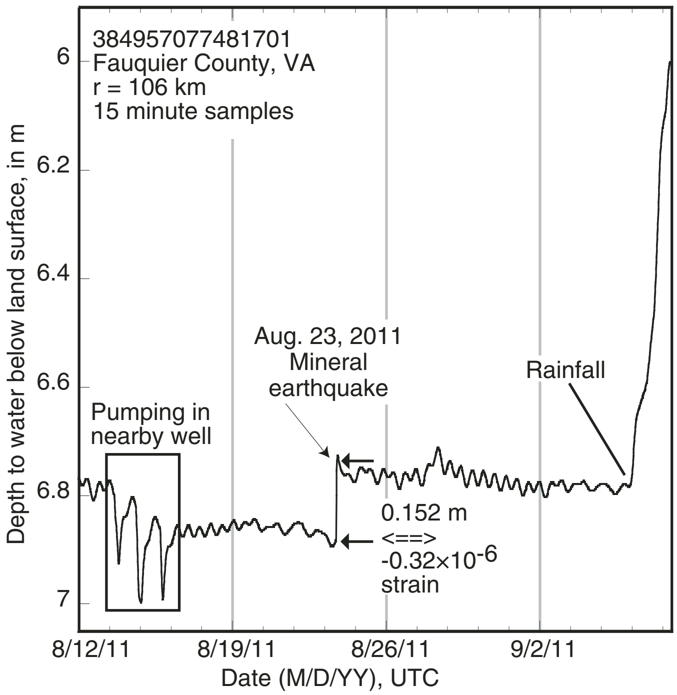

Figure 13. Depth to water in the Fauquier County, Virginia (VA), well, for 4 weeks spanning the time of the M_w 5.8 Mineral, VA earthquake.

Figure 14 shows ~10 days before and after the earthquake for the Orange County well, the closest to the epicenter (39 km), with the second-largest observed offset. This water-level drop recovers within 14 h, but apparently not all the way to the pre-earthquake trend. The Orange County well responds to atmospheric pressure, but does not exhibit a tidal response. This, in addition to the well's shallow depth (29.9 m), likely indicates that the well is poorly confined: aquifer pressure may change in response to strain, but strain-induced changes dissipate by equilibrating with the water table over a time scale no longer than the period of the semidiurnal tide. With this interpretation, only the part of the water-level drop that recovered would represent the well's response to coseismic strain.

Calculated Strain Field

We considered two possible distributions of areal strain at the Earth's surface from the M_w 5.8 main shock (and possible afterslip), calculated using codes by Okada (1992). These codes approximate the Earth as a uniform, isotropic, linearly elastic half-space. The pattern of expansion and contraction depends on the fault strike and dip, the direction of slip, and the depth range in which rupture occurred. For a given fault geometry, strain amplitude is proportional to the amount of slip. Beyond several rupture dimensions, the shape of the rupture surface becomes unimportant, and the strain amplitude is proportional to the product of fault area and slip. Because the fault slip imposes a specified displacement, the strain amplitudes are independent of the elastic shear modulus. A Poisson's ratio of 0.25 was assumed; the calculated strain distribution does not change appreciably for Poisson's ratios in the range 0.17–0.27.

We used the fault geometry found by Chapman (2013): a reverse fault striking N29°E and dipping 51° to the southeast, with rake 113°, i.e., a slight component of right-lateral motion. The rupture initiated at 8 ± 1 km. Chapman (2013) identified three subevents in the main shock rupture, the total moment of which was 2.72×10^{17} N·m. Our first model (Figs. 15A, 15B) presumes that the fault is a 2 × 2 km square with net slip corresponding to that seismic moment, and with the main shock hypocenter at the lower southwest corner (because early aftershocks were almost all shallower than 6 km). Figure 15B is a map of the areal strain distribution corresponding to this model.

TABLE 2. AMPLITUDES OF WATER-LEVEL FLUCTUATIONS CAUSED BY THE M_2 EARTH-TIDE STRAIN IN THE CHRISTIANSBURG, VA, WELL, AND IN THE WELLS SHOWN IN FIGURES 15A AND B FOR WHICH THE EARTH-TIDE RESPONSE CAN BE RESOLVED

USGS site number	Location	Water-level offset (m)	Time of maximum excursion (UTC)	Barometer*	M_2 tide in water level (m)	Calculated M_2 tide (nano-strain)	Water level/strain (m/microstrain)	Water level change converted to microstrain
372053078493801	Appomattox County, VA	–0.061	Aug. 24, 01:30	Lynchburg 724100	0.0083	16.51	0.505	0.121
384957077481701	Fauquier County, VA	0.152	Aug. 23, 19:00	Dulles 724030	0.0077	15.98	0.482	–0.316
384821078472301	Shenandoah County, VA	–0.046	Aug. 24, 12:00	Shenandoah 724105	0.0025	16.46	0.150	0.305
372150079422301	Bedford County, VA	–0.210	Aug. 23, 18:30	Roanoke 724110	0.0213	16.94	1.257	0.167
372224079423601	Bedford County, VA	0.052	Aug. 23, 19:45	Roanoke 724110	0.0087	16.94	0.512	–0.101
370812080261901	Christians-burg, VA	–0.140	Aug. 24, 02:00	Roanoke 724110	0.0169	17.24	0.980	0.143

*Name of airport where barometer is located, followed by the National Climatic Data Center Weather Bureau Army Navy (NCDCWBAN) identifier for that weather station. See Appendix for website from which barometer data were obtained.

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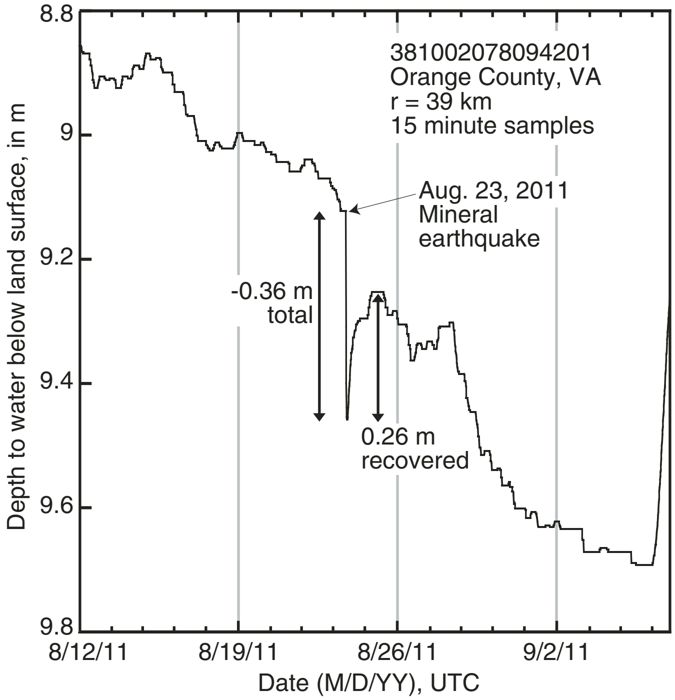

Figure 14. Depth to water in the Orange County, Virginia, well, for 20 days approximately centered on the time of the M_w 5.8 Mineral, Virginia, earthquake.

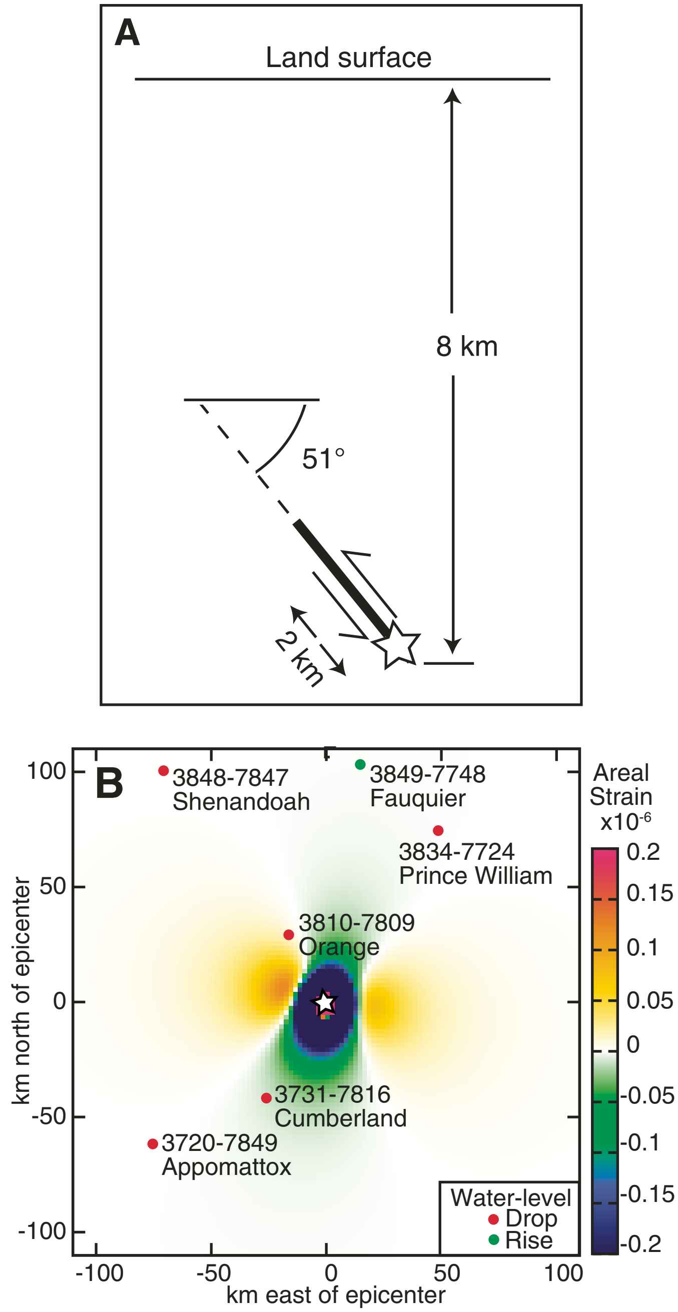

Figure 15. (A) Cross section perpendicular to fault strike showing dip and along-dip extent of rupture for the Mineral, Virginia, earthquake, based on results of Chapman (2013). (B) Areal strain distribution based on a 2 × 2 km square fault with net slip of 2.27 m (corresponding to a seismic moment of 2.72×10^{17} N·m), with the main shock hypocenter at the lower southwest corner, and cross section as shown in A. Star represents the main shock hypocenter.

An alternative model (Figs. 16A, 16B) accounts for possible afterslip following the main shock. Owing to the lack of nearby geodetic instruments, aseismic afterslip would not have been measured for the Mineral earthquake, but is not implausible given that aftershocks defined a dipping plane, aligned with the main shock rupture, extending upward to within 3 km of the surface, and along strike for 5 km from the main shock. Assuming that the amount of slip in the main shock extended throughout this area yields the strain distribution shown in Figure 16B, in which strains are about six times larger than those in Figure 15B. The strains in Figure 16B are likely an upper bound on strain in the 4 or 9 min between the main shock and the first groundwater-level measurements.

Well Responses to Strain

In porous formations within consolidated bedrock, fluid pressure is expected to decrease or increase in response to expansion or contraction of the formation, respectively, and to be unaffected by shear strain (e.g., Roeloffs, 1996). Comparing the areal strains in Figures 15B and 16B with the groundwater-level changes shows that, at four of the six closest wells (Orange, Fauquier, Shenandoah, and Appomattox Counties), the sign of the calculated strain is consistent with the directions of the water-level offsets. However, although in locations subjected to contractional strain, the Cumberland County and Prince William County wells had water-level drops, which were probably caused

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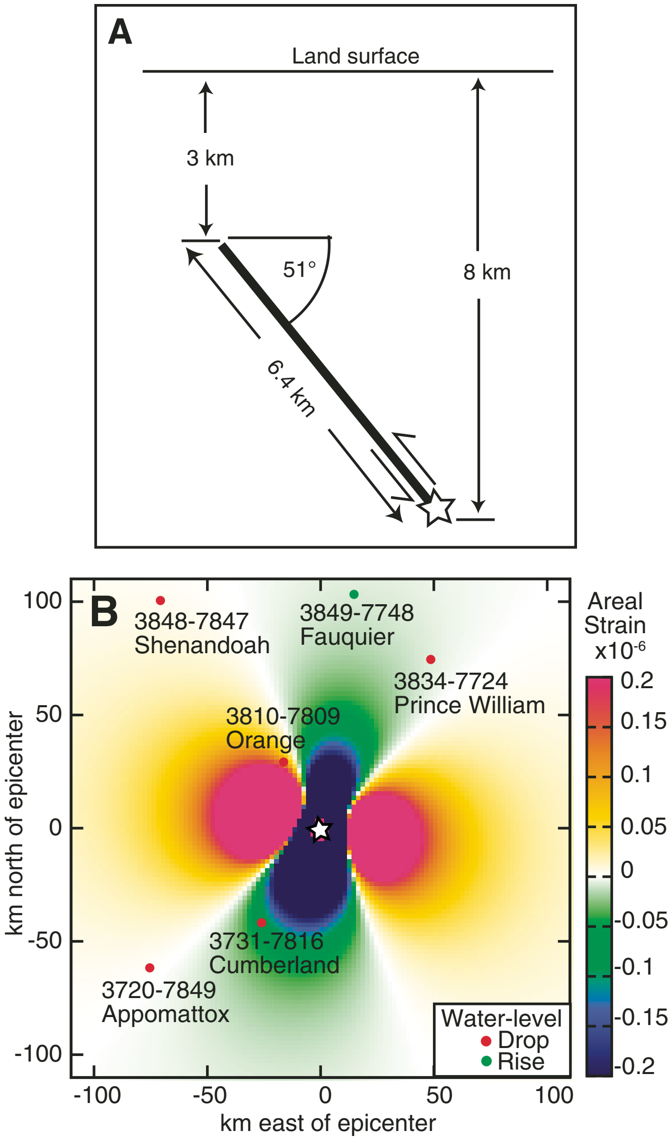

Figure 16. (A) Cross section perpendicular to fault strike showing dip and along-dip extent of rupture for the Mineral, Virginia, earthquake and possible afterslip. (B) Areal strain distribution based on a 6.4 × 5 km square fault with 2.27 m net slip, with the main shock hypocenter at the lower southwest corner, and cross section as in A. Star represents the main shock hypocenter.

by ground motion, supported in part because these two wells do not respond to Earth tides.

The Appomattox County, Fauquier County, and Shenandoah County wells respond to Earth tides. We used the Earth tide responses of these three wells to estimate coefficients relating water level to strain. Tidal analysis was done using the longest suitable data record (30–113 days) during the period from 1 May 2011 until just prior to the Mineral earthquake. In that time period, excursions caused by rainfall were much smaller than during the weeks following the earthquake. The wells also respond to atmospheric pressure changes. We corrected for that effect using barometer data downloaded from the National Oceanic and Atmospheric Administration (NOAA; see Appendix 1) for the nearest available locations (Table 2). For each well, Table 2 gives the amplitudes of the M_2 Earth tide constituent in the groundwater-level data and in the theoretical areal strain tide, including the effects of the ocean load (Agnew, 1996, 1997). The coefficient relating water-level change to areal strain is the ratio of these two amplitudes. The values of 0.50 and 0.48 m/microstrain for the Fauquier County and Appomattox County wells are typical of those for rocks in which porosity does not have a preferred orientation; the value of 0.15 m/microstrain for Shenandoah County is at the low end of that range. Using these strain sensitivities to convert the observed groundwater-level changes to strain yields the strains that would be needed to produce the observed water-level changes.

Even the upper-bound strains shown in Figure 16B are an order of magnitude too small to account for the water-level changes in the Appomattox County, Fauquier County, and Shenandoah County wells. If, however, a strain sensitivity of 0.5 m/microstrain is assumed for the Orange County well, then the strain corresponding to the part of the groundwater-level drop that recovered (0.26 m) would be reduced to 0.13 microstrain, which agrees with the calculated strain shown in Figure 16B.

The effects of ground motion on the wells nearest the earthquake can account for all of the water-level changes observed in that area. Coseismic strain may account for part of the change observed in the Orange County well, but only if the amount of slip and fault area are much larger than the inferred main shock rupture plane. The Fauquier County well exhibited a water-level rise that resembles a response to strain and is consistent with its location in an area subjected to coseismic contraction, but calculated strains cannot account for the size of that water-level rise, even if substantial slip beyond the main shock fault plane is included. Although in other studies groundwater-level changes have been reconciled successfully with imposed static strain, the results shown here imply that the effects of ground motion must be considered in making such comparisons.

CONCLUSIONS AND FUTURE DIRECTIONS

The 2011 M_w 5.8 Mineral earthquake resulted in water-level offsets in wells as far as 550 km away. The earthquake resulted in seismic intensities as great as 5 on the DYFI scale to within 550 km of the epicenter, and >6 within 200 km of the epicenter, as well as static strains that could have affected groundwater levels within at most 100 km of the epicenter. We attribute all or most of the groundwater-level offsets to ground motion, based on the resemblance of the spatial distribution of the offsets to the distribution of online reports assigned intensity 3 or greater. For this earthquake, the amplitudes of the largest groundwater-level offsets at any intensity increased monotonically with increasing intensity.

In some wells, water level may have changed only because seismic waves dislodged material clogging the wellbore walls, or flushed particulate matter from fractures intersecting the

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wellbore. However, such mechanisms are unlikely to explain water-level offsets in wells that have shown oscillations, without offsets, in response to distant earthquakes, and does not explain the correlation of the direction of groundwater-level offsets with topography, observed in a pair of wells in Bedford County and to some extent in the complete set of groundwater-level offsets.

The wells are in consolidated bedrock and are too far from the epicenter for pore pressure to increase by undrained consolidation (Wang and Chia, 2008). In the absence of any other mechanism for transient seismic shaking to directly change fluid pressure, we conclude that the seismic waves affected aquifer properties. Because seismic-wave motion was significant for only a few minutes, while the groundwater levels continued to change for at least an hour and as much as several months, we conclude that the effects of ground motion on these bedrock aquifers persisted much longer than the ground motion. In at least two wells, it is not certain that these changes have yet recovered. The most likely aquifer property to have been affected by the seismic waves is permeability. Increased permeability along some parts of hydrologic flow paths would cause well water levels to adjust more closely to the (higher or lower) ambient aquifer pressure to which a hydrologic connection had been enhanced. Support for this mechanism includes the occurrence of water-level rises only in valleys or areas of low topographic relief; the existence of a pair of wells for which water level in an upland well falls while that in a nearby valley well rises; and the location of the largest water-level drop in a setting with high topographic relief, which not only amplifies seismic shaking, but also provides a steep hillslope pressure gradient.

The Mineral earthquake groundwater-level offsets appear to have been caused by short-period ground acceleration, rather than by strains accompanying propagation of long-period surface waves. Seismic waves could have increased permeability temporarily by mobilizing fine loose particles, colloids, or gas bubbles blocking flow, or more permanently by damaging fragile grain-to-grain contact. The aquifers that were affected are as deep as 213 m, where ground acceleration and velocity would be expected to be much lower than at the Earth's surface. We speculate that the pore pressure in the aquifers increases their vulnerability to damage by reducing the effective overburden stress, noting that only minor aquifer damage may be needed to affect permeability.

Several of the water-level offsets are in wells that respond to Earth tides, and in two of the wells nearest the epicenter, the offsets have features consistent with being caused by earthquake-induced strain. However, only in the nearest well can part of the change be reconciled with estimated strain, and then only if significant afterslip followed the earthquake. If other wells responded to strain, these were inelastic strains caused by ground accelerations or dynamic stresses. The influence of ground motion should be taken into account when inferring strain from groundwater-level changes associated with earthquakes.

For the Mineral earthquake, the distribution of seismic ground motion is similar to the distribution of water-level offsets in aquifers in the upper few hundred meters of the crust. These ground-motion effects range from small water-level changes caused, in turn, by small changes of permeability, to very large water-level drops in hillslope aquifers. It is possible that the tendency of earthquakes to raise aquifer pressure in valleys may help promote liquefaction. The underlying mechanisms of permeability changes caused by seismic ground motion may be useful in understanding seismic wave attenuation, a parameter critical in estimating seismic hazard.

The possibility that seismic waves can affect permeability at greater depths, under conditions of low effective pressure, requires investigation. Evaluation of the effects of earthquakes on subsurface waste isolation may need to consider whether confining layers could be compromised by the passage of seismic waves. The possibility that seismic waves can damage grain contacts at depths where earthquakes nucleate, thereby reducing rock stiffness and strength, may help explain observed interactions between earthquakes. In areas of magmatic activity, increased vertical permeability could explain how seismic waves stimulate hydrothermal activity, perhaps even helping to clear pathways for hydrothermal explosions.

Additional field and laboratory research is needed to identify the mechanisms by which seismic ground motion changes aquifer properties. In particular, deployment of an accelerometer and broadband seismometer at the Christiansburg well, and increasing the water-level sampling rate to 1 sps or faster would allow us to differentiate between the seismic wave phases responsible for water-level oscillations and those that lead to persistent water-level changes. Video inspection and flowmeter logging of a subset of the affected wells could ascertain the distribution of heads in contributing fractures, helping to assess the role of near-wellbore fracture-transmissivity changes. Earthquake-induced aquifer-property changes are potentially of practical importance and warrant further, more quantitative, investigation.

APPENDIX. ONLINE WELL SITE INFORMATION AND GROUNDWATER-LEVEL DATA

Site maps for every well are available online at http://nwis.waterdata.usgs.gov/nwis/nwismap.

Enter the site number listed in the first column of Table 1 and choose the "Show sites on a map" output format option. Topographic contours can be added as a layer to the map.

Groundwater-level data for every well can be obtained starting at the same website. Enter the site number from the first column of Table 1 and choose the "Table of sites..." output option. On the next screen, click the site number, which will bring up a location map. At the top of the map is a blue bar labeled "Available data for this site." Change the selection from "Location Map" to "Time series: current/historical observations." This brings up a page on which starting and ending times can be entered to obtain a graph of the data or a table for download.

Intensity data were obtained from the U.S. Geological Survey "Did You Feel It?" website for the Mineral earthquake (http://earthquake.usgs.gov/earthquakes/dyfi/events/se/082311a/us/index.html).

Atmospheric pressure data were obtained from the National Oceanic and Atmospheric Administration (NOAA) Satellite and Information Service National Climatic Data Center website (http://www7.ncdc.noaa.gov/CDO/cdopoemain.cmd?datasetabbv=DS3505&countryabbv=&georegionabbv=&resolution=40; Hourly data - 3505).

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ACKNOWLEDGMENTS

The facilities of the Incorporated Research Institutions for Seismology (IRIS) Data Management System (DMS), and specifically the IRIS Data Management Center, were used for access to seismic waveform and metadata required in this study. The IRIS DMS is funded through the National Science Foundation and specifically the Directorate for Geosciences through the Instrumentation and Facilities Program of the National Science Foundation under Cooperative Agreement EAR-1063471.

We thank William Cunningham [U.S. Geological Survey (USGS) Office of Groundwater], Mark Kozar (USGS Washington Water Science Center), Dennis Risser (USGS Pennsylvania Water Science Center), John Williams (USGS New York Water Science Center), and Cian Dawson (USGS Office of Groundwater) for identifying observation wells with water-level changes and gathering data after the earthquake. Insightful reviews by John Lane, David Lockner, Jamison Steidl, and Chi-Yuen Wang allowed us to improve the manuscript significantly.

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Manuscript Accepted by the Society 6 June 2014
Manuscript Published Online 22 October 2014

Printed in the USA

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Widespread groundwater-level offsets caused by the M_W 5.8 Mineral, Virginia, earthquake of 23 August 2011

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Geological Society of America Special Papers 2015;509; 117-136 , originally published onlineOctober 22, 2014
doi:10.1130/2014.2509(07)

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