Chapter 6

Groundwater Level

Abstract Groundwater level has long been known to respond to earthquakes; several types of response have been documented. Advances in the last decade were made largely through the studies of water-level response to Earth tides and barometric pressure. These studies have demonstrated that the hydraulic properties of groundwater systems are dynamic and change with time in response to disturbances such as earthquakes. This approach has been applied to estimate the permeability of several drilled active fault zones, to identify leakage from deep aquifers used for the storage of hazardous wastewater, and to reveal the potential importance of soil water and capillary tension in the unsaturated zone. Enhanced permeability is the most cited mechanism for the sustained changes of groundwater level in the intermediate and far fields, while undrained consolidation remains the most cited mechanism for the step-like coseismic changes in the near field. A new mechanism has emerged that suggests that coseismic release of pore water from unsaturated soils may also cause step-like increases of water level. Laboratory experiments show that both the undrained consolidation and the release of water from unsaturated zone may occur to explain the step-like water-level changes in the near field.

6.1 Introduction

Coseismic changes of groundwater level have been documented since the time of antiquity (e.g., Institute of Geophysics—CAS 1976). Since the late twentieth century, instrumental records of groundwater level have become widely available and advanced our understanding of earthquake-induced groundwater changes (e.g., Waller et al. 1965; Wakita 1975; Whitehead et al. 1984; Rojstaczer and Wolf 1992; Quilty and Roeloffs 1997; Roeloffs 1998; Wang et al. 2001; Chia et al. 2001; Brodsky et al. 2003; Matsumoto et al. 2003; Montgomery and Manga 2003; Sato et al. 2004; Kitagawa et al. 2006; Sil and Freymueller 2006; Wang and Chia 2008; Mohr et al. 2015, 2017; Xue et al. 2013, 2016; Yan et al. 2014; Wang et al. 2016, 2018; Zhang et al. 2017; Liao and Wang 2018; Barbour et al. 2017, 2019; Zhang et al. 2019a, b; Zhu and Wang 2020). Furthermore, a broad range of coseismic responses of water

level to earthquakes have been discovered. The oscillatory response of water level was discussed in the last section of Chap. 5. Here we focus on the non-oscillatory responses (Fig. 6.1). The rise time of a non-oscillatory response ranges from ‘step-like’ to several days or even weeks. However, whether a coseismic change appears ‘step-like’ or ‘gradual’ depends on the time resolution of the recording system. For example, a ‘step-like’ change recorded at a rate of every hour may appear ‘gradual’ if recorded every minute or second. Here we qualitatively define a coseismic signal as step-like if the coseismic change between the last data point before the earthquake and the first data point after the earthquake is the predominant change in the data record (Fig. 6.1a, b), and as ‘sustained’ if the coseismic change is part of a continuous trend (Fig. 6.1c, d).

Earthquakes produce static and dynamic strains in the crust. Static strains are permanent crustal deformations produced by fault slip, and dynamic strains are oscillatory crustal deformations caused by the passage of seismic waves. Both can produce compression and extension in the crust and thus change pore pressure. Local hydraulic gradients force groundwater to flow through pores and fractures, to mobilize fine particles and to change permeability, storage and compressibility (e.g., Brodsky et al. 2003).

As noted in the previous chapter, measurements of the groundwater level response are made in wells. Two general approaches have been followed to study the water-level responses to earthquakes. First is to analyze the response of water level in a

Fig. 6.1 **a** Step-like, positive water-level change in the Yuanlin I well during the 1999 Mw7.5 Chi-Chi earthquake. The well is ~25 km from the hypocenter, thus in the near field, and ~13 km from the surface rupture of the causative fault. The step-like coseismic water-level increase is +6.55 m. **b** Step-like, negative water-level change in the Liyu II well during the 1999 Mw7.5 Chi-Chi earthquake in Taiwan. The well is ~20 km from the hypocenter, thus in the near field, but is only ~5 km from the surface rupture of the causative fault. The step-like coseismic water-level decrease is -5.94 m. **c** Sustained change of water level in USGS well 364,821 during the 6 November Mw 5.0 Cushing earthquake. Note that the coseismic change of water level was gradual, instead of step-like, and the post-seismic water level was sustained, instead of decaying exponentially with time. **d** Sustained change of water level in USGS well 364,831 during the 6 November Mw 5.0 Cushing earthquake (modified from Wang et al. 2001, 2017)

single well to many earthquakes. Second is to compare and contrast the response of water levels in many wells in a region to a specific earthquake. The advantage of the first approach is that the well itself is often carefully calibrated, thus the various non-seismic influences on the groundwater-level records can often be eliminated and very small changes in the groundwater level may be detected at great distances from the earthquake (e.g., Roeloffs 1998). Furthermore, since the geology of the well site likely does not change between different earthquakes, the complications introduced from geological and hydrological heterogeneities at different well sites—a problem for the use of multiple wells—are eliminated and the groundwater records can often be used to effectively discriminate different models of the causal mechanisms of groundwater-level changes.

By using many wells in a region to study the groundwater response to a particular earthquake, it is possible to examine the regional pattern of the coseismic responses and thus the influence of geological and hydrogeological properties. For example, after the 2016 Mw7.0 Kumamoto earthquake, Japan, a dense network of monitoring wells in central Kyushu, Japan, captured dramatic spatiotemporal changes in groundwater levels (Fig. 6.2; Hosono et al. 2019). Water-level dropped over an area of 160 km² along crustal ruptures immediately after the main shock (within 35 min) in both confined and unconfined aquifers. A maximum drop of ~5 m was documented in a confined aquifer. Water level in the unconfined aquifer largely recovered within 45 days toward the background level (Fig. 6.2e). Water level in the confined aquifer continued to rise and reached ~5 m higher than that before the earthquake, one-year after the earthquake (Fig. 6.2h).

We may also note that currently two types of wells have been used in studying groundwater responses. Most data for studying groundwater responses to earthquakes have been collected from groundwater wells installed in unconsolidated sediments. On the other hand, a large number of wells, including those for monitoring earthquakes in China and those for oil exploration in the USA, have been installed in consolidated rocks, which have been increasingly used to study groundwater response to earthquakes. The lithologic contrast between these two types of wells may cause significant differences in their response to earthquakes. For instance, some mechanisms for groundwater response to earthquakes, such as undrained consolidation and liquefaction, may not occur in consolidated rocks. On the other hand, other mechanisms, such as enhanced permeability, may occur in both consolidated and unconsolidated rocks. Hence lithological differences among different wells need to be accounted for when comparing observations from different areas and for understanding the groundwater response to earthquakes.

As discussed in earlier chapters, studies of induced seismicity in the mid-continental U.S.A. since 2009 significantly advanced our understanding of the relationship between water injection and earthquakes (Chap. 4) and studies of the tidal and barometric responses of groundwater level (Chap. 5) have significantly improved our understanding of earthquake hydrology. In this chapter, we focus on the effects of natural earthquakes on the changes of groundwater level. We first discuss the relevant observations, followed by a discussion on the proposed hypotheses and relevant

Fig. 6.2 Spatiotemporal changes of water levels following the 2016 Mw7.0 Kumamoto earthquake in Japan. The western rim of the Aso Mountain is visible on the upper right of each diagram. **(a to h)** Relative water level changes comparing water levels before the Kumamoto earthquake sequence at 35 min, 7 days, 45 days, and 365 days after the main shock for unconfined and confined aquifers. The area enclosed by a dashed curve shows the region affected by extensive faults. Earthquake epicenters and fault systems are marked by stars and red lines, respectively. Numbers 1–10 in **(g)** refer to wells discussed in Chap. 8, Fig. 8.3 (from Honoso et al. 2019)

constraints. We end the chapter with two short sections on estimating the hydraulic properties of continental fault zones and in the oceanic crust.

6.2 Observations

Based on the records from a single well in central California, Roeloffs (1998) showed three categories of groundwater-level response. In the near field, groundwater level shows step-like increases (Fig. 6.1a). In the intermediate field, groundwater-level changes are more gradual and can persist for days or weeks (Fig. 6.1c, d). At even greater distances (the far field), transient oscillations of the water-level occur (Fig. 5.39).

Based on the data from a network of monitoring wells in central Taiwan responding to the 1999 Chi-Chi earthquake, Wang et al. (2001) showed abrupt decreases of groundwater level in the immediate vicinity (<10 km) of the ruptured fault—(Fig. 6.1b) that were followed by either an exponential increase or an exponential decrease with time. Other examples of using large networks of wells to examine the response of groundwater level to earthquakes include the study by Yan et al. (2014) on the response of groundwater level on the Chinese mainland to the 2011 M9.0 Tohoku earthquake, Japan, and the study of by Hosono et al. (2019) on groundwater level changes in Japan after the 2016 Kumamoto earthquake (Fig. 6.2).

6.2.1 Coseismic Step-like Changes of Groundwater Level

In the near field, i.e., the area around the hypocenter within a distance of ~ 1 ruptured fault length, groundwater level often shows step-like changes during earthquakes (Wakita 1975; Quilty and Roeloffs 1997; Chia et al. 2001; Wang et al. 2001) and changes in excess of 10 m in amplitude are not uncommon. In some cases (e.g., Jonsson et al. 2003; Akita and Matsumoto 2004), coseismic changes of water level in wells show a characteristic pattern of positive and negative changes correlated with the coseismic static change of volumetric strain (Fig. 6.3). For this reason, the coseismic static change of volumetric strain has often been taken as the mechanism for coseismic step-like water-level changes.

However, such correlation was not found in other cases with distributed wells (e.g., Wang et al. 2001; Yan et al. 2014; Mohr et al. 2017; Hosono et al. 2019). For example, a dense network of monitoring wells was installed on an alluvial fan (the Choshui River fan) near the epicenter of the 1999 Mw7.6 Chi-Chi earthquake in Taiwan (Fig. 6.4) prior to the earthquake. About 200 wells captured the groundwater level changes during and after the earthquake. Water levels in these wells are recorded by digital piezometers and are logged at 1 h intervals. Some wells are equipped with data loggers operating up to 1 Hz. Piezometer readings in the well are converted to groundwater level with an accuracy of 1 cm. The resolution of the reading, on

Fig. 6.3 Calculated pore-pressure change based on the calculated coseismic volumetric strain and theory of poroelasticity (color map) and observed coseismic water-level changes (circles) following a 2000 M 6.5 strike-slip earthquake in Iceland. Black and white circles indicate water-level increases and decreases, respectively. The white line shows the mapped surface rupture (from Jonsson et al. 2003)

the other hand, is finer by an order of magnitude. In addition, several rain gauges installed around the fan provide continuous records of precipitation in the area. The close proximity to a large earthquake and the dense network of hydrological stations made this dataset one of the most comprehensive and systematic to study the spatial distribution of the hydrologic response in the near field. For the sake of better interpreting the observations, a brief summary of the subsurface hydrogeology of the Choshui River alluvial fan is needed. Figure 6.4d shows a simplified hydrogeological cross-section across the alluvial fan. It shows that the fan consists of subhorizontal layers of unconsolidated Holocene and Pleistocene sediments. Three distinct aquifers may be distinguished: Aquifer I, the topmost aquifer, is partly confined and partly unconfined, while aquifers II and III are confined. To the east of the alluvial fan is the Western Foothills of the Taiwan fold-and-thrust belt, which consists of pervasively faulted and fractured Pleistocene sedimentary rocks.

The step-like coseismic rise of groundwater level, such as that shown in Fig. 6.1a, was commonly observed on the Choshui River fan during the Chi-Chi earthquake (Wang et al. 2001; Chia et al. 2001). Figure 6.4 shows the coseismic changes in groundwater level in the three aquifers in the Choshui River fan during the Chi-Chi earthquake. In the uppermost partially confined aquifer (Fig. 6.4a), the coseismic changes are generally small except in an area on the northeastern edge of the alluvial fan where positive changes occurred. This area of positive water-level change is closely associated with the occurrence of liquefaction on the fan during the Chi-Chi earthquake (Chap. 10). The distribution of coseismic changes of water level in the two lower confined aquifers (Fig. 6.4b, c), on the other hand, showed coseismic groundwater level changes that are not associated with the occurrence of liquefaction on the surface. Instead, the water-level rise in these aquifers showed an increase with distance away from the ruptured fault, reaching a maximum at distances of 20–30 km from the fault, and then decreased at greater distances (Chia et al. 2001; Wang et al. 2001).

Fig. 6.4 Distribution of observed coseismic changes in groundwater level (**a**, **b**, **c**) and calculated volumetric strain (**d**) in the Choshui River fan during the Chi-Chi earthquake. **a** Contours (m) of groundwater-level change in the uppermost aquifer (Aquifer I). **b** Contours of groundwater-level changes in a confined aquifer (Aquifer II). **c** Contours of groundwater-level changes in a lower confined aquifer (Aquifer III). Groundwater monitoring stations are shown in open circles, epicenter of Chi-Chi earthquake in star, and the ruptured fault in discontinuous red traces. (from Wang et al. 2004). **d** Simplified hydrogeological cross-section along the cross-section marked by AB in Fig. 7.4, showing the Choshui River fan and the fold-and-thrust Foothills. The surface trace of the ruptured Chelungpu fault is marked by the letter C. The enlarged inset shows three aquifers in the Choshui River alluvial fan. Massive gravel beds (marked with greyish oblate ellipsoids) occur in proximal area; away from proximal area, gravel beds decrease in thickness, while coarse sands (highlighted yellow) and fine sands (highlighted blue) increase in proportion and interfinger with gravel beds; further away, silty sands and silty clays (highlighted brown) increase in proportion and eventually dominate the distal margin of fan. Three aquifers can be identified from top (Aquifer I) to bottom (Aquifer III). Dashed vertical lines show boreholes. Numbers on vertical axes give elevation relative to the mean sea-level. (from Wang et al. 2005) **(e)** Distribution of coseismic volumetric strain changes calculated from a dislocation model for the Chi-Chi earthquake. Positive and negative values indicate dilatation and contraction, respectively. Black dots are the locations of observation wells (from Koizumi et al. 2004)

Lowered groundwater levels were also reported in the vicinity of ruptured faults, including near the ruptured fault in the 1999 Mw7.5 Chi-Chi earthquake in Taiwan (Fig. 6.1c, d, Wang et al. 2001), or in areas of tectonic extension, such as the valley floor cut by extensional faults after the 2016 Mw7.0 Kumamoto earthquake in Japan (Fig. 6.2b, Hosono et al. 2019). In the second case, significant drawdown of groundwater over an area of 160 km² along crustal ruptures and ~10⁶ m³ of surface water disappeared within 35 min after the main shock. The Duoqing Co lake, located in

a rift valley in southern Tibet with an average surface area of 57 km², dried up suddenly following a nearby M3.7 earthquake in 2016. Geologic examination of the dry lake floor showed numerous tensional fractures oriented subparallel to the general direction of the rift valley (Wu et al. 2018).

6.2.2 *Sustained Changes*

Sustained groundwater-level changes are characterized by a gradual onset and can last for days or weeks (Fig. 6.1c, d). They are normally documented in the intermediate field of an earthquake, defined here as a distance from an earthquake source greater than 1 ruptured fault length but within a radius of 10 ruptured fault lengths. Although these changes are probably also common in the near field, their relatively small amplitude (<1 m) makes them easily obscured by the step-like changes of relatively large amplitude at such distances (see Sect. 6.2.1). Thus, they are clearly revealed only in the intermediate field where the abrupt changes become sufficiently small or absent.

Roeloffs (1998) showed that the water level changes in a well (BV) in California showed gradual responses to earthquakes and may take a few days to a few weeks to reach their peak. Furthermore, the coseismic change does not exponentially decay to the pre-seismic state (Fig. 6.1c, d). Based on observations Roeloffs (1998) showed that the occurrences of such changes are bounded by an empirical relation $M = -3.91 + 1.82 \log r$, where M is earthquake magnitude and r the maximum epicentral distance in meters for a specific M , beyond which water level response is not expected. Different relations were proposed by King et al. (1999) and Matsumoto et al. (2003) to describe the threshold distance for other data sets. The differences among these relations most likely reflect the relatively small number of the data used in the analyses and/or the different geology among the well sites where the data were documented, or both. Wang and Manga (2010) collected published data for the coseismic water-level changes (Roeloffs 1998; King et al. 1999; Roeloffs et al. 2003; Brodsky et al. 2003; Matsumoto et al. 2003; Sato et al. 2004; Kitagawa et al. 2006; Sil and Freymueller 2006) and referred to it as the 'global dataset' (Fig. 6.5). Weingarten and Ge (2014), using 24 years of data documented at Devils Hole, Nevada, significantly expanded the 'global dataset' (Fig. 6.5); new data from Sun et al. (2018) and Zhang et al. (2019b) are added to this compilation.

6.2.3 *Breached Confinement*

Taiwan is a north-south elongated island arc formed by the oblique collision between the Luzon volcanic arc on the Philippine Sea plate and the continental margin of China beginning in the late Cenozoic (Teng 1990). The Choshui River Alluvial Fan (Fig. 6.4) is part of the Coastal Plain that lies along the central western coast of the island and

Fig. 6.5 Global coseismic occurrences of groundwater level change and liquefaction plotted on a diagram of logarithm of epicentral distance ($\log r$) versus earthquake magnitude (M). Also plotted are the contours of constant seismic energy density, based on Eq. (6.10). The thick purple line, associated with a seismic energy density of 0.1 J/m^3 , marks the upper bound of the occurrence of liquefaction and is known as the liquefaction limit (see also Chap. 10) (modified from Zhang et al. 2019)

is covered by unconsolidated sediments of Neogene and Quaternary age, floored by a faulted basement (Fig. 6.4d). The Western Foothills that lie immediately to the east of the Coastal Plain, on the other hand, is a fold-and-thrust belt of consolidated sedimentary rocks (Fig. 6.4d; Ho 1988). The 1999 Chi-Chi ($M_w = 7.5$) earthquake, the largest to hit Taiwan in the last century, ruptured the Western Foothills along a $\sim 80 \text{ km}$ fault on the east of the Choshui River fan (Fig. 6.4a–c).

Wang et al. (2007) showed that the 1999 Chi-Chi earthquake breached the confinement of some aquifers in the Choshui River alluvial fan (Fig. 6.4d) near the epicenter. Monitoring stations with clustered wells installed on this alluvial fan (Fig. 6.4a–c) revealed distinct water levels in different aquifers at the same station before the earthquake (Fig. 6.6) showing that good confinement was present. After the earthquakes (marked by $t = 0$), however, some stations showed converged water levels in different aquifers to the same level and stayed converged for one to several days, before they

Fig. 6.6 Water levels at two stations with clustered wells before and after the Chi-Chi earthquake. Water level is referenced to the mean sea level, with positive values above sea level and negative values below sea level. **a** Hourly data for water level in four wells at the Chuanhsin Station documented by piezometers located, respectively, at depths of 8 m (aquifer 1), 111 m (aquifer 2), 181 m (aquifer 3), and 245 m (aquifer 4). Time of the earthquake is marked by $t = 0$. The wells are open to aquifers separated by aquitards 40–50 m thick. **b** Hourly data for water level in three wells at the Louchin Station documented by piezometers located, respectively, at depths of 12 m (aquifer 1), 110 m (aquifer 2), and 199 m (aquifer 3). Time of earthquake is marked by $t = 0$. The wells are open to aquifers separated by an aquitard ~60 m thick. Notice that the water levels at each station were distinct before the Chi-Chi earthquake, but converged rapidly to the same level after the earthquake. **c, d** Daily average of water levels at the same two stations (from Wang et al. 2016)

gradually diverged and returned to the pre-seismic levels. Such convergence of water levels is unlikely to have occurred by chance, because the probability for the earthquake to generate pore pressure changes in different aquifers that reach the same level, and stay at the same level for an extended time, is extremely small. It is more likely that the earthquake created or opened hydraulically conductive paths (e.g., cracks) that allowed pore pressures in the different aquifers to communicate and to equilibrate. The different amount of time for water levels at different stations to reach equilibrium may reflect the uneven distribution of the hydraulically conductive paths created by the earthquake.

The suggestion that the confinement of the aquifer was breached during the Chi-Chi earthquake was further demonstrated by using the change of the tidal response of the aquifer before and after the earthquake (Wang et al. 2016), as further explained in Sect. 6.4.3. Loss of confinement may commonly occur, even for deeply buried aquifers such as the Arbuckle aquifer in Oklahoma, as discussed in Chap. 5 (see also Wang et al. 2018).

6.3 Models and Constraints

Many hypotheses have been proposed to explain the observed water level changes during earthquakes. In this section we briefly describe in separate sub-sections the frequently invoked hypotheses, which include coseismic static strain, undrained consolidation and liquefaction, enhanced permeability, and shaking water out of the unsaturated zone.

6.3.1 Coseismic Static Strain

The coseismic static strain model (Wakita 1975) proposes that coseismic changes of volumetric strain cause the coseismic changes of groundwater level; it thus predicts water-level rises in areas of coseismic contraction and falls in areas of coseismic dilation. In support of this hypothesis, several authors reported a correlation between the distribution of coseismic groundwater level changes after some earthquakes and the calculated coseismic changes of static strain from a dislocation model in support of this hypothesis (e.g., Wakita 1975; Igarashi and Wakita 1991; Roeloffs 1996; Quilty and Roeloffs 1997; Ge and Stover 2000; Jonsson et al. 2003; Akita and Matsumoto 2004). As an example, Fig. 6.3 shows the spatial correlation in the pattern of groundwater level changes observed following a 2000 M 6.5 strike-slip earthquake in Iceland and that calculated from a dislocation model (Jonsson et al. 2003).

The correlation between the observed pattern of coseismic water-level change and that calculated from a dislocation model, however, was not found in some areas with a dense distribution of monitoring wells were installed and the coseismic groundwater level changes were documented; these include the responses to the 1999 Mw7.6 Chi-Chi earthquake in western Taiwan (Chia et al. 2001; Wang et al. 2001), to the Darfield earthquake sequence in eastern New Zealand (Rutter et al. 2016), to the 2010 Maule earthquake in Chile (Mohr et al. 2017), and to the 2016 Mw7.0 Kumamoto earthquake in southern Japan (Hosono et al. 2019). Figure 6.4a, b, c show that the coseismic changes of water level in western Taiwan during the 1999 Mw 7.6 Chi-Chi earthquake were mostly positive in three aquifers. This pattern of the coseismic changes of water level is opposite to that expected from the static strain hypothesis because the thrust mechanism of the Chi-Chi earthquake predicts a positive change (extension) of coseismic strain in front of the ruptured thrust fault, and thus predicts a coseismic decrease of water level over the Choshui River fan (Wang et al. 2001). This argument was further supported by the result of a numerical simulation by Koizumi et al. (2004) (Fig. 6.4e), in contradiction to the prediction of the coseismic strain hypothesis.

Wang and Barbour (2017) showed that most measured coseismic volumetric strains differ substantially from the predicted strains based on the static dislocation model, sometimes with opposite signs. The disagreement suggests that some

Fig. 6.7 Time histories of the measured normalized volumetric strain (black), volumetric strain converted from the measured pore pressure (red), and volumetric strain converted from calculated pore pressure based on the Roeloffs (1998) model of diffusion from a localized coseismic source (blue) during the Tohoku earthquake with an assumed hydraulic diffusivity of $10^{-3} \text{ m}^2/\text{s}$ (modified from Zhang et al. 2017)

processes affecting the coseismic volumetric strain are missing from the dislocation model. Roeloffs (1998) also found that some sustained changes of water level in wells have opposite signs to that expected from the static volumetric strain, and that the groundwater level always rises in some wells but falls in other wells, regardless of the locations or focal mechanisms of the earthquakes, in contradiction to the static strain model. Kitagawa et al. (2006) showed that only half of the measured coseismic strains and water level changes in Japan during the 2004 M9 Sumatra earthquake, more than 5000 km away, were consistent with the static strain hypothesis. Finally, the static strains at such large distances are too small to explain many of the observed water-level responses (e.g., Igarashi and Wakita 1991; Itaba and Koizumi 2007; Wang and Barbour 2017).

Figure 6.7 shows the water level and volumetric strain in the Fuxin well, NW China, during the 2011 Mw9.1 Tohoku earthquake, Japan, which was >1500 km from the well. At such a large epicentral distance, the static elastic strain due to the earthquake rupture is negligible and cannot produce any measurable coseismic change of pore pressure. However, significant changes of both pore pressure and volumetric strain occurred in this well and, furthermore, the observed coseismic increase of pore pressure (Fig. 6.7, black dashed curve) is in the opposite direction from that expected from the hypothesis that the observed volumetric strain was the causal mechanism (Fig. 6.7, red dashed curve, converted to the equivalent water level, extension positive). Zhang et al (2017) suggested that, instead, the increase of volumetric strain was caused by a coseismic increase of pore pressure. They converted the volumetric strain to pore pressure and compared the measured time-series of pore pressure with the predicted pore pressure from the model of pressure diffusion from a local coseismic source (Roeloff 1998). The good agreement among the three curves (Fig. 6.7, blue curve) supports the hypothesis that the coseismic increase

of volumetric strain was due to increased pore pressure produced by earthquake-enhanced permeability and diffusion between the well and a pre-existing, local crustal heterogeneity in pore pressure.

6.3.2 *Undrained Consolidation and Liquefaction*

The undrained consolidation and liquefaction model was based on the results of a great number of laboratory experiments in the past four decades by geotechnical engineers to study earthquake-induced liquefaction (e.g., Seed and Lee 1966; Dobry et al. 1982; Vucetic 1994; Hsu and Vucetic 2004) and was used to explain the coseismic increase of groundwater level following the Chi-Chi earthquake (Wang et al. 2001). These experiments show that loose sediments consolidate under cyclic shear stress, which causes pore-pressure to increase, the effective stress to decrease, and eventually liquefaction of sediments. Examples of the experimental results are shown in Chap. 3 (Figs. 3.8 and 3.10–3.12). More discussions are given in Chap. 11.

After the Chi-Chi earthquake, an interesting correspondence was found between the area of liquefaction occurrence (Fig. 3.10) and the occurrence of coseismic increase of water level in the topmost aquifer (Fig. 6.4a; Wang et al. 2006). Also interesting is the absence of an association between liquefaction occurrence and water-level increase in the lower aquifers (Fig. 6.4b, c). Such association between liquefaction occurrence with pore-pressure increases in the topmost aquifer and the lack of association with pore pressure change in the lower aquifers supports the common assumption in earthquake engineering that liquefaction occurs mostly in the upper 15 m of sediments.

6.3.3 *Enhanced Permeability*

Permeability controls groundwater flow; thus, any change of permeability would lead to changes in groundwater flow and groundwater level if the hydraulic gradient is fixed. One model of seismically enhanced permeability assumes that seismic shaking increases the mobility of colloidal particles and air bubbles and removes them from flow channels such as microcracks and pore throats, which may increase the permeability of rocks and sediments (e.g., Mogi et al. 1989; Roeloffs 1998; Brodsky et al. 2003; Roberts and Abdel-Fattah 2009; Manga et al. 2012).

Another model assumes that permeability may change due to poroelastic opening and closing of micro-fractures in response to the transient pore pressure (Faoro et al. 2012). If the external stress remains constant, the aperture of a fracture would change in response to transient pore pressure in the fracture, and the evolving permeability would scale with the change in aperture Δb as $k/k_o = (1 + \Delta b/b_o)^3$, where b_o is the initial width of the aperture.

Brodsky et al. (2003) found that the water level recorded in a well in Southern Oregon did not respond significantly to the first 10 cycles of the ~ 20 s Rayleigh waves from the 1999 Mw7.4 Oaxaca earthquake, but responded with large amplifications to the Rayleigh waves that arrived later (Fig. 5.39a). These observations are more consistent with mobilization of trapped colloidal particles or air bubbles, which are progressively mobilized by the oscillatory pressure beyond some threshold, rather than with poroelastic opening or closing of fractures (Manga et al. 2012). The common observation of increased turbidity in wells after earthquakes (e.g., Sneed et al. 2003) is also consistent with the model of mobilization of colloidal particles in pores and fractures of aquifers by earthquakes.

Elkhoury et al. (2011) used Berea sandstone samples fractured in a triaxial deformation apparatus to investigate the influence of pore pressure oscillations on permeability. A servo-control system was applied at the inlet and outlet to establish pore pressures. After the fluid flow reached a steady state, sinusoidal oscillations of pore pressure, with 20 s periods, were applied at the inlet while keeping the outlet pressure constant. The application of the pore pressure oscillation leads to an immediate increase of permeability, with magnitude increasing exponentially with the amplitude of the pressure oscillations (Fig. 6.8). Following the dynamic oscillations, permeability recovered as the inverse square root of time. This recovery demonstrates that the observed flow rate change was due to changes of permeability, but not to the change of poroelastic storage (Elkhoury et al. 2011). It also led the authors to favor the mechanism of clogging and unclogging of the fracture flow paths. Additional experiments were performed on samples fractured outside the apparatus and

Fig. 6.8 Permeability changes in sandstone samples measured in laboratory as a function of imposed pore pressure oscillation amplitude. Permeability changes are normalized by permeability before the oscillations; pressure amplitudes are normalized by the pore pressure difference driving the flow. Different symbols and colors correspond to different rock samples (modified from Elkhoury et al. 2011)

then re-assembled for the flow through experiments. In these instances no change in permeability was observed after stimulation.

Roberts (4) studied the influence of axial stress oscillations on permeability in intact sandstone cores. After steady state had been reached, continuous dynamic stressing with frequency of 50 Hz and amplitude of 0.3 MPa was applied, resulting in no observable effects on permeability. The amplitude of the continuous stress cycling was increased to 0.6 MPa; after which permeability increased 15%. A further increase in the amplitude of stress cycling to 0.9 MPa added another 5% increase in permeability. Permeability returned to the original value after the stress oscillations terminated.

Liu and Manga (2009) performed similar experiments on fractured sandstone samples saturated with de-ionized water. Permeability was first measured with steady flow, followed by transient stresses by oscillating the axial displacement to achieve strain amplitudes of 10^{-4} at frequencies from 0.3 to 2.5 Hz. In general, permeability decreased after each set of oscillations. Additional experiments were performed with natural silt particles injected into the fractures. Fractures with added silt showed the largest decrease in permeability in response to the oscillatory stresses. No recovery of permeability was documented within ~10 min of the stimulation.

Collectively, these experiments show that rock permeability does change with oscillatory stresses and pore pressure. The observed differences in these experiments may be due to differences in the sample preparation, in the type of applied stresses (oscillation of pore pressure or axial strain/stress), and in the differences in the applied oscillation frequency (Manga et al. 2012). Also noticeable is that the transient strains in all these experiments are at least an order of magnitude greater than those that cause permeability to change in natural systems ($\sim 10^{-6}$).

6.3.4 Shaking Water Out of Unsaturated Soil

Mohr et al. (2015) observed that the stream flows in some headwater catchments in the Chilean Coast Range increased following the 2010 Mw8.8 Maule (Chile) earthquake. They attributed this increase to the release of water from the unsaturated soil to recharge the local groundwater. This observation will be discussed further in the next chapter on stream flow. The proposed mechanism of water lease from the unsaturated zone is tested in a recent laboratory study (Breen et al., 2020) and is discussed in the next section.

6.4 Constraints

6.4.1 Constraints from Laboratory Experiments

Both the mechanism of water release from the unsaturated zone (Mohr et al. 2015; Sect. 6.3.4) and that of undrained consolidation (Wang et al. 2001; 6.3.2) predict step-like increases of pore pressure in the affected aquifer. Breen et al. (2020) used laboratory experiments to test and distinguish between these mechanisms. The experiments were designed to study the response of unconsolidated sediments during “seismic” shaking and were carried out in a sand column (Fig. 6.9a) subjected to

Fig. 6.9 a Schematic drawing of the sand column in the experiment. b Pore pressure relative to hydrostatic in the experiments with the water level above the sand surface. The solid blue line shows the pore pressure during the third impact, and the dotted blue line shows the pore pressure after the last (60th) impact. Time is relative to impact. Note that the duration of the pressure increase far exceeds the duration of shaking (acceleration). c Measured pore pressure in three sets of experiment with an unsaturated zone above the water table. In each experiment, the water level was set at a distinct height above the base of the sand column, as shown in the figure legend. Time of impact is marked by $t = 0$ (from Breen et al. 2020)

“seismic” shaking of controlled energy. In one set of experiments, the water level was set above the sand surface; thus, there was no capillary effect. All the experiments showed that, under “seismic” shaking, pore pressure suddenly increased with a rise time of ~ 1 s, and then declined slowly with time (Fig. 6.9b). Since there was no capillary effect, the increase of pore pressure must be due to the in situ volumetric contraction of the sediment matrix, confirming the undrained consolidation hypothesis. If the water level was set below the sand surface, leaving a thick unsaturated zone above the water table, pore pressure also suddenly increased with shaking, and then declined exponentially with time (Fig. 6.9c). Breen et al. (2020) attributed the rise to the disruption of capillary tension, which caused pore pressure to suddenly increase, supporting the hypothesis that capillary forces were altered. The rise time of pore pressure in the experiments with the unsaturated zone is much shorter (< 0.1 s) than that in the experiments with the undrained consolidation mechanism. Hence, in principle, these two mechanisms can be distinguished from their different rise times. However, most field experiments have so far been carried out at a recording rate much lower than 1 Hz, and hence do not have the time resolution to distinguish between the two mechanisms. For this reason, both mechanisms are acceptable for the moment to explain the coseismic step-like water-level changes.

6.4.2 Constraints from Field Observations

Direct association of field-scale observations to a specific mechanism is challenging because the subsurface cannot be easily monitored at the scales required for making such connections (Manga et al. 2012). On the other hand, indirect inferences may be drawn from field observations and laboratory experiments to provide useful constraints on the various mechanisms.

Wang and Chia (2008) showed in Fig. 6.10a that, for a global dataset, the sign and magnitude of sustained coseismic water level changes were randomly distributed with the epicentral distance. The enhanced permeability model requires connection of the well to a nearby source (or sink) that can occur either up-gradient or down-gradient of the well; thus, either positive or negative change of water-level may be expected. If a sufficiently large number of observations is available, the enhanced permeability model would predict a statistically random occurrence in the sign and the magnitude of the water-level changes, consistent with the global data presented in Fig. 6.10a.

Figure 6.10b shows the data from a dense network of monitoring wells in central Taiwan near the epicenter of the 1999 Mw7.6 Chi-Chi earthquake (Fig. 6.4). The data provides a nearly continuous sequence of the changes of sign and amplitude of water-level from the vicinity of the ruptured fault to a distance of 160 km.

In the immediate neighborhood of the ruptured fault during the Chi-Chi earthquake, marked by the downward arrow, large decreases of groundwater level occurred. This was attributed to downward flow through dilatant fractures formed during strong seismic vibrations (Chap. 3, Sect. 3.4). Similarly, large decreases of

Fig. 6.10 Amplitude and sign of water-level changes during earthquakes plotted as a function of the epicentral distance. **a** Water-level changes in a global dataset (Wang and Chia 2008) updated with responses to the 2011 Tohoku earthquake in wells on the Chinese mainland (Yan et al. 2014). **b** Water-level changes during the 1999 Chi-Chi earthquake. The upward-pointing arrow shows the distance equal to one ruptured fault length. The downward-pointing arrow shows the location of the ruptured fault. Note that nine wells near the downward-pointing arrow documented abrupt decreases of water level as illustrated in Fig. 6.1b. These wells are all located within 5 km of the surface rupture of the causative fault. **c** Water-level changes during the 2006 Hengchun earthquake. The upward-pointing arrow shows the distance equal to one ruptured fault length. Note that at distances beyond one ruptured fault length, the sign of water-level changes is random (from Wang and Chia 2008)

groundwater level were also reported near the ruptured faults during the 1989 Mw6.9 Loma Prieta earthquake, California (Rojstaczer and Wolf 1992; Rojstaczer et al. 1995) and during the 1995 Mw6.9 Kobe earthquake, Japan (Tokunaga 1999).

At epicentral distances further from the ruptured fault but still within one ruptured fault length (~85 km) during the Chi-Chi earthquake, the groundwater-level changes were predominantly positive (Fig. 6.10b), consistent with the undrained consolidation hypothesis. At epicentral distances greater than one ruptured fault length, marked by the upward arrow, the signs of the water level changes become random and the magnitude of the changes became relatively small.

Figure 6.10c shows the water-level changes in a large number of monitoring wells in southern Taiwan, documented during the 2006 Hengchun earthquake off the southern coast of Taiwan. Most wells were at distances beyond one ruptured fault length, marked by the upward arrow, and showed random distribution of signs, with both positive and negative changes at further distances, consistent with the enhanced permeability model. A few wells at closer distances show positive changes, consistent with the occurrence of undrained consolidation.

Overall, the observations from wells installed in unconsolidated sediments are consistent with the model that the dominant mechanism for the coseismic change of groundwater level in the near field is undrained consolidation of saturated sediments and/or release of water from the unsaturated zone. In the intermediate and far fields, the dominant mechanism may be earthquake-enhanced permeability. These mechanisms provide simple explanation for why the water-level changes are step-like in the near field, but are more gradual and sustained in the intermediate and far fields. Furthermore, in the intermediate and far fields, since the same source (or sink) and the same passageway may be activated during different earthquakes, some wells may show consistently positive or consistently negative water-level changes during different earthquakes (e.g., Roeloffs 1998; Matsumoto et al. 2003).

Finally, we note that enhanced permeability may occur both in the near field and in the intermediate field. In the near field, the change of groundwater level due to undrained dilatation or consolidation may be so large that the sustained groundwater level changes, which are usually of smaller amplitude (<1 m), are obscured. Sustained changes can thus be clearly detected only in the intermediate and far fields where undrained consolidation is no longer important (Wang and Chia 2008). We also note that the post-seismic recovery of the enhanced permeability would proceed with re-clogging of the passageways by various hydrological and geochemical processes that take time. Thus, the model predicts a gradual recovery of the water level, in contrast to the exponential recovery that characterizes pressure diffusion following the step-like increases of water level in the near field.

6.4.3 Constraint from Tidal Analysis

The tidal response of groundwater was discussed in Chap. 5. Here we discuss the application of this response to the study of the interaction between water and earthquakes. Elkhoury et al. (2006) first applied this method to study earthquake-induced changes in groundwater systems by analyzing 20 years of water-level response using data from two wells in southern California. They found repeated step-like changes in phase at the time of earthquakes (Fig. 6.11), each followed by a gradual recovery of the phase to the pre-earthquake values. They also noted that all earthquakes produce decreases in the phase lag, implying an increase in aquifer permeability (Hsieh et al. 1987), regardless of the sign of the earthquake-induced static strain at the well sites, which shows that the change in permeability was not caused by the coseismic static strain. They further showed that the amount of increase in permeability appears to be linearly proportional to the magnitude of the peak ground velocity (PGV) of ground shaking, with a maximum increase of a factor of 2–6 (Fig. 6.12).

Since the study by Elkhoury et al. (2006), the tidal method has been widely applied to study earthquake effects on groundwater systems. These studies have revealed that earthquakes can not only enhance aquifer permeability but also breach the confinement between aquifers and the surface. Wang et al. (2016) analyzed the tidal response of groundwater level in two wells discussed in Sect. 6.2.3 (the Chuanhsin station); their results before and after the Chi-Chi earthquake are plotted in Fig. 6.13a, b on amplitude versus phase shift diagrams. On such a diagram, data points before the Chi-Chi earthquake (blue dots) in each well show a trend consistent with that of a confined aquifer model (dotted lines). After the earthquake, however, data points (red dots) deviate significantly from that of a confined aquifer. Each data point represents a 3-day average of the tidal response. The arrows in both diagrams connect

Fig. 6.11 Phase of the semi-diurnal tides for the water levels in two wells in southern California relative to the tidal strain. Transient changes of the phase are clearly evident at the time of earthquakes, as shown by the vertical lines (from Elkhoury et al. 2006)

Fig. 6.12 Increased permeability of the aquifers at the two wells plotted against the peak ground velocity (PGV) of ground motion during earthquakes. The maximum increases correspond to a factor of 5–6 increase in aquifer permeability (from Elkhoury et al. 2006)

the last tidal data before the Chi-Chi earthquake to the first data after the earthquake; these also deviate significantly from the trend defining a confined aquifer (dashed lines). The numbers attached to the data points show the number of days since the Chi-Chi earthquake. Thus, after the earthquake, the tidal response in each well exhibits a time-dependent excursion from the trend defining a confined aquifer, and the excursions in the two wells are closely similar, suggesting that the two aquifers behaved hydraulically the same after the earthquake. In other words, the two aquifers were effectively connected by hydraulically conductive fractures after the earthquake and they behaved hydraulically in the same way. The excursions in both wells terminated ~60 days after the earthquake after which the tidal responses fall back onto the trend for a confined aquifer, as indicated by the dotted lines. Note also that the duration of the excursion of the tidal response in each well from that of a confined aquifer lasted ~60 days, an order of magnitude longer than the duration suggested by the change of groundwater level in Fig. 6.6a, c. This difference suggests a greater sensitivity of the tidal response to the occurrence of aquifer leakage. Quantitative interpretation of the tidal response, however, was not made for the present case because the data for ocean tides off western Taiwan, which load on the aquifers and cause the groundwater to oscillate, was not measured before or after the Chi-Chi earthquake.

Numerous other applications of the tidal method have been made to study earthquake-affected aquifer properties including evaluating the permeability of the ruptured fault in the 2008 Mw7.9 Wenchuan earthquake (Xue et al. 2013; Sect. 6.5), the long-term or irreversible changes of aquifer permeability after the Wenchuan and the 2011 Mw9.0 Tohoku earthquakes (Liao et al. 2015; Zhang et al. 2019b; Fig. 5.37), the leaking of the Arbuckle aquifer in Oklahoma, a target for a great amount of wastewater disposal produced by hydrocarbon extraction (Wang et al. 2018; Zhu and Wang 2020; Sect. 5.4.3), and capillary effects on the tidal response of unconfined aquifers (Wang et al. 2019; Sect. 5.4.4).

Fig. 6.13 Amplitude versus phase shift for water-level response to the M₂ tide in two wells at the Chuanhsin Station. In each well, each point represents the phase shift and amplitude calculated for a data window. Data points in both wells before the earthquake are shown by blue dots and after the earthquake by red dots. Before the earthquake the data fall along a trend for a confined aquifer model (dashed line) (Hsieh et al. 1987). During the earthquake, the coseismic changes in both wells are given by the arrows that connect the last pre-seismic data point to the first post-seismic data point. Data points during the first 60 days after the earthquake are numbered in sequence to show the time-dependent evolution. Both the arrows and the post-seismic data points during the first ~60 days deviate significantly from the trend defining a confined aquifer (dashed line). Numbering refers to the number of days after the earthquake (from Wang et al. 2016)

6.4.4 Constraints from Threshold Seismic Energy

In the past four decades earthquake engineers have performed numerous laboratory experiments to study consolidation and liquefaction of saturated sediments under cyclic shearing (e.g., National Research Council 1985, 2016; Ishihara 1996). The results of these experiments show that, when sediments are subjected to cyclic shearing, they begin to consolidate if the shear strain magnitude exceeds a threshold of $\sim 10^{-4}$ (Fig. 3.9; Dobry et al. 1982; Vucetic 1994). Since deformation during earthquakes is undrained, pore pressure increases when the shear strain exceeds 10^{-4} and the effective stress decreases. This would eventually lead to liquefaction if deformation continues. Wang et al. (2001) suggested that undrained consolidation explains the coseismic water-level rises on the Choshui River fan during the Chi-Chi earthquake. If the amplitude of shear deformation exceeds some critical threshold, however, cracks and fractures may form and cause pore pressure to decrease (Luong 1980), which may explain the water-level drop immediately adjacent to the ruptured fault during the Chi-Chi earthquake (Wang et al. 2001).

The geotechnical laboratory data may be used to calculate the amount of dissipated energy required to cause undrained consolidation; the magnitude of this energy may then be compared with the seismic wave energy in the field at different epicentral distances to constrain the mechanism. In cyclic loading, the dissipated energy density required to initiate undrained consolidation in saturated sediments may be estimated from the experimental time histories of shear stress τ and shear strain γ by performing the following integration:

$$e_d(t) = \int_0^t \tau d\gamma, \quad (6.1)$$

where the integration extends from the beginning of the cyclic loading to the onset of pore-pressure increase. Since both the shear strain and the dissipation are small and the stress and strain relation is nearly linear in this case, we may express the cyclic experimental stress and strain in the form $\tau = \tau_o \sin\theta$ and $\gamma = \gamma_o \sin(\theta + \varphi)$, respectively, where τ_o and γ_o are the corresponding amplitudes, and φ is the phase angle between stress and strain. Integrating Eq. (6.1) we obtain, for small φ ,

$$e_d = \frac{N\pi}{2} \tau_o \gamma_o \sin \varphi \approx \frac{N\pi}{2} \mu \gamma_o^2 \varphi \quad (6.2)$$

where $N \sim 10$ is the usual number of cycles adopted in experimental studies, μ is the shear modulus measured at strain amplitude of 10^{-4} and φ is also known as the damping ratio (Ishihara 1996).

Table 6.1 lists some experimental data for the shear modulus and the damping ratio of sands and gravels collected from different sources, prepared with different procedures, and subjected to different confining pressures. These experimental data are used with Eq. (6.2) to calculate the dissipated energy density e_d required to initiate

Table 6.1 Shear moduli and damping ratios of sediments, determined under laboratory conditions and cyclically sheared to strain amplitude (γ_o) of 10^{-4}

Samples	Shear modulus (MPa)	Damping ratio	Dissipated energy (J/m^3)
Undisturbed Fujisawa sand	100	0.07	1
Disturbed Fujisawa sand	40	0.02	0.1
Undisturbed Tokyo gravel, at a confining pressure of 300 kPa	300	0.04	2
Undisturbed Tokyo gravel, at a confining pressure of 500 kPa	600	0.05	5
Reconstituted Tokyo gravel, at a confining pressure of 300 kPa	200	0.07	2
Reconstituted Tokyo gravel, at a confining pressure of 500 kPa	400	0.08	5

Dissipated energy density is calculated using Eq. (6.2) with $N = 10$. Experimental values for the shear modulus and the damping ratio were extracted from figures on p. 143 and 145 in Ishihara (1996)

undrained consolidation of the sediments. The calculated dissipated energy density required to initiate undrained consolidation, e_d , range from ~ 0.1 to $\sim 5 \text{ J/m}^3$. The broad range of energy density required to initiate undrained consolidation highlights that different sediments have different sensitivity to cyclic loading and that the sediments were subjected to different confining pressures during the measurements (e.g., Ishihara 1996). Irrespective of the limited laboratory experiments, the existing data suggests that a threshold energy density of 0.1 J/m^3 may be taken as the threshold e_d to initiate undrained consolidation in saturated sands.

Wang et al. (2006) argued that the observed bounds on the epicentral distance for a hydrologic response to earthquakes may reflect the threshold seismic energy required to initiate the hydrologic response. Here we first derive an empirical relation among the seismic energy density, epicentral distance, and earthquake magnitude; we then use this relation to associate the threshold energy density to epicentral distance and earthquake magnitude. Using $\sim 30,000$ strong-motion records for southern California earthquakes, Cua (2004) showed that the peak ground velocity (PGV) for sediment sites attenuates with the epicentral distance as

$$\text{PGV} \sim A/r^{1.5} \quad (6.3)$$

where A is an empirical parameter for southern California. The seismic energy density e at a site during ground shaking may be evaluated from the time histories of particle velocity of the ground motion as recorded by strong-motion seismometers (Lay and Wallace 1995):

$$e(r) = \frac{1}{2} \sum_i \frac{\rho}{T_i} \int v_i(t)^2 dt, \quad (6.4)$$

where the summation is taken over all the relevant modes of the ground vibrations, ρ is density, and T_i and v_i are, respectively, the period and the velocity of the i th mode. Since most energy in the ground motion resides in the peak ground velocity, Wang et al. (2006) simplifies (6.4) to

$$e(r) \sim \text{PGV}^2 \quad (6.5)$$

and showed that this relation is consistent with ground motion records in California. It follows from Eqs. (6.3) and (6.5) that the seismic energy density declines with the epicentral distance according to

$$e(r) \sim A/r^3. \quad (6.6)$$

Note that this relation does not include the effect of source dimension or rupture directivity that may become important for the distribution of seismic energy, and thus can only be taken as a first-order point-source approximation. The constant A in (6.6) may be evaluated by noting that the total seismic energy of an earthquake from a point source is related to the energy density by

$$E = \frac{4\pi}{3} r^3 e(r). \quad (6.7)$$

Thus, at $r = 1$ m we have

$$E \sim \frac{4\pi}{3} A. \quad (6.8)$$

Inserting this A into Eq. (6.6) we have

$$e(r) = \left(\frac{3E}{4\pi} \right) r^{-3} \quad (6.9)$$

Replacing E in Eq. (6.9) by the Båth's empirical relation $\log E = 5.24 + 1.44 M$ (Båth 1966, Eq. 11.9), where the unit of energy is converted to Joule, we obtain a relation among e , r and M (Wang 2007)

$$\log_{10} e = -3 \log_{10} r + 1.44 M - 4.62, \quad (6.10)$$

where r is in km. Equation (6.10) shows that contours of constant seismic energy density e plot as straight lines on a diagram of $\log r$ versus M (Fig. 6.5). Note that this relation is entirely empirical and is based on the field data from southern California and is thus strictly valid only for southern California. This relation is expected to show significant differences from region to region. However, due to the lack of sufficient seismic data in different regions, the southern California attenuation relation has often been applied to other regions without validation.

Figure 6.5 shows that the coseismic change of groundwater level may occur at a threshold seismic energy density as low as 10^{-6} J/m^3 , many orders of magnitude lower than the threshold seismic energy density of 0.1 J/m^3 required for the initiation of undrained consolidation of saturated sand and the occurrence of liquefaction under laboratory condition (Ishihara 1996). Thus, there are large differences in the minimum energy required to initiate different earthquake-induced hydrological responses. Regardless of whether the mechanism for hydrological responses is directly connected to seismic energy density, these differences imply very different sensitivities to transient stresses. While part of this difference may be due to incomplete data and different geological conditions, some data sets, such as those for groundwater level and liquefaction, are large and the differences in the threshold seismic energy density should be robust. These differences also indicate that during earthquakes there may be more than one mechanism at work and responsible for the observed hydrological responses.

Hazirbaba and Rathje (2004) measured the threshold strain required to initiate undrained consolidation in the laboratory and showed that it is the same as that in the field. Wang and Chia (2008) showed that this threshold agrees well with the field-observed liquefaction limit but is many orders of magnitude greater than that required for the coseismic changes of groundwater level (see also Fig. 6.5). Thus, assuming that the laboratory data may be compared with field observations, other mechanisms must be evoked to explain the coseismic occurrence of groundwater level change beyond the near field.

The threshold energy for triggering hydrological responses may depend on the type of seismic wave. Wang et al. (2006) observed that liquefaction documented during underground explosions is characterized by a threshold energy two orders of magnitude greater than that for liquefaction during earthquakes, even though a similar attenuation relation exists between ground-motion energy density and distance. They interpret the observation to be consistent with the understanding that the seismic energy generated by explosion occurs mostly in compression, with much less shear energy than that in natural earthquakes of equivalent magnitude, and shearing is more effective than compression in triggering liquefaction.

6.4.5 Post-seismic Recession of Groundwater Level

The coseismic disturbances of the groundwater level in the near field are often followed by an exponential decline of the water level towards an equilibrium state

Fig. 6.14 **a** Schematic drawing of model geometry, with water divide at $x = 0$ and local discharge at $x = L$. **b** Schematic drawing of hydraulic head, with zero head gradient at the local water divide and zero head at the local discharge. **c** Schematic drawing of coseismic vertical recharge A to the aquifer. **d** Time-history of post-seismic groundwater level documented at Yuanlin I well (Wang et al. 2001). Black dots show data points. Colored curves are model predictions with several values of x/L (see text for explanation). Excellent fit between data and curve occurs for $x/L = 0.5$ (modified from Wang et al. 2004)

(Fig. 6.1a, b). This post-seismic recession of water level contains information about aquifer properties immediately after the earthquake (Wang et al. 2001), that may provide useful constraints on the proposed mechanisms. We discuss here the post-seismic recession and how to extract information about the aquifer properties from post-seismic recession data.

Using an analytical model (Fig. 6.14), Wang et al. (2004) simulated the post-seismic water level recession often observed in water wells in the near field. The model assumes a subhorizontal aquifer with length much greater than its thickness, and may thus be approximated by a one-dimensional model extending from a local groundwater divide ($x = 0$) to a local discharge or recharge area ($x = L$). The time-dependent change of the hydraulic head h is controlled by the governing flow equation (Chap. 2):

$$S_s \frac{\partial h}{\partial t} = K \frac{\partial^2 h}{\partial x^2} + A \quad (6.11)$$

where K is the horizontal hydraulic conductivity, S_s is the specific storage, and $A(x, t)$ is the rate of water released per unit volume (Fig. 6.14c) from the coseismic consolidation in the saturated zone or water released from the unsaturated zone due to the disruption of the capillary force during seismic shaking, or a coseismic sink of water due to earthquake-induced porosity or fractures. Even though this model may appear highly simplified, many studies have demonstrated that the procedure is useful for characterizing the catchment-scale response of hydrological systems to earthquakes (e.g., Roeloffs 1998; Manga 2001; Manga et al. 2003; Brodsky et al.

2003; Montgomery et al. 2003; Manga et al. 2016; Wang et al. 2017; Mohr et al. 2017).

Taking the background head as the reference, the initial condition before the earthquake is $h = 0$ at $t = 0$. For boundary conditions, a no-flow boundary condition is applied at the local water divide (i.e., $\partial h / \partial x = 0$ at $x = 0$) and $h = 0$ at the local discharge ($x = L$). Since the time duration for the coseismic release of water is very much shorter than the time duration for the post-seismic recession of groundwater level, we may consider that the coseismic release of water occurs instantaneously; i.e.,

$$A(x, t) = A_o(x)\delta(t = 0). \quad (6.12)$$

The solution for this problem is (derivation is given in the Appendix)

$$h(x, t) = \frac{1}{LS_s} \sum_{n=1}^{\infty} \cos \frac{n\pi x}{2L} \exp\left[-\frac{Dn^2\pi^2 t}{4L^2}\right] \int_{-L}^L Q_o(x') \cos \frac{n\pi x'}{2L} dx' \quad (6.13)$$

where L is the length of the aquifer, $D \equiv K/S_s$ and $Q_o(x)$ is the integration of (6.12) with time and is a function of x only. The terms in (6.13) decrease rapidly with increasing n and time; thus Eq. (6.13) is dominated by the first term ($n = 1$) of the series expansion for sufficiently long times after the earthquake, i.e.,

$$h(x, t) \approx \frac{1}{LS_s} \cos \frac{\pi x}{2L} \exp\left[-\frac{D\pi^2 t}{4L^2}\right] \int_{-L}^L Q_o(x') \cos \frac{\pi x'}{2L} dx'. \quad (6.14)$$

Differentiating Eq. (6.14) with respect to time we have

$$\frac{\partial \log h}{\partial t} \approx -\frac{\pi^2 D}{4L^2} \equiv -b \equiv -\frac{1}{\tau}. \quad (6.15)$$

We use the letter b here to denote the post-seismic recession of groundwater level to distinguish it from the post-seismic recession of stream discharge c in Chap. 7.

Notice that Eq. (6.15) is independent of the location of measurement; hence τ (or its inverse, $b = 1/\tau$, the recession constant), thus D/L^2 , of a responding well may be determined from the field measurement of water level versus time during the post-seismic recession. Various environmental factors such as barometric pressure, precipitation, tides, and human activities such as the withdrawal or injection of groundwater, may affect the temporal groundwater-level record, which are not considered in this formulation. Such environmental disturbances thus need to be corrected before the water level records may be used for estimating the characteristic time or the recession constant of an aquifer.

Table 6.2 lists the values for b , τ and the square of the correlation coefficient, R^2 , determined from least square fitting of the data for the postseismic recession

Table 6.2 Values for the post-seismic recession constant, b , the characteristic time, τ , and the square of the correlation coefficient, R^2 , determined by least square fitting of the data for several aquifers on the Choshui River alluvial fan, documented at several monitoring stations after the 1999 Chi-Chi earthquake (from Wang et al. 2004)

		Aquifer I	Aquifer II	Aquifer III	R^2
Chiulung	b (s^{-1}) τ (s)			4.0×10^{-7} 2.5×10^6	0.99
Hsihu	b (s^{-1}) τ (s)			3.6×10^{-7} 2.8×10^6	0.99
Huatang	b (s^{-1}) τ (s)		3.8×10^{-7} 2.6×10^6		0.99
Kuoshen	b (s^{-1}) τ (s)		9.7×10^{-7} 1.0×10^6		0.98
Yuanlin	b (s^{-1}) τ (s)	8.8×10^{-7} 1.1×10^6	9.3×10^{-7} 1.1×10^6		1.00
Chukou	b (s^{-1}) τ (s)	1.7×10^{-6} 5.9×10^5	2.0×10^{-6} 5.0×10^5		0.98
Liyu	b (s^{-1}) τ (s)	4.4×10^{-7} 2.2×10^6	6.0×10^{-7} 1.7×10^6		0.99
Hsichou	b (s^{-1}) τ (s)		1.0×10^{-5} 1.0×10^5		0.92
Chushan	b (s^{-1}) τ (s)	1.4×10^{-5} 0.7×10^5			0.99
Hsinkuang	b (s^{-1}) τ (s)	1.0×10^{-5} 1.0×10^5			0.99
Pingting	b (s^{-1}) τ (s)	1.4×10^{-5} 0.7×10^5			0.98

documented at several monitoring stations on the Choshui River alluvial fan after the 1999 Chi-Chi earthquake. It shows that the characteristic time for the postseismic dissipation of the hydraulic head is about 3×10^6 s for most aquifers. But the last three stations (Chushan, Hsinkuang and Pingting) show characteristic times an order-of-magnitude shorter, i.e., $\sim 10^5$ s. These wells are all located close to the ruptured Chelungpu fault and the smaller characteristic times may reflect a relatively high post-seismic hydraulic diffusivity (D), a relatively small post-seismic characteristic length (L), or both, at these stations.

Given the characteristic time τ in Table 6.2 determined from the recession analysis and a specific storage of 10^{-4} m^{-1} from well tests (Tyan et al. 1996), we may compare the model prediction with the post-seismic time history of the groundwater level change. As an example, we compare in Fig. 6.14d the post-seismic time history of the groundwater level change at the well Yuanlin I against the model predictions (Eq. 6.14) for different values of x/L . A constant value of $Q_o = 3.4 \times 10^{-4} \text{ m}^3/\text{m}^3$ was used to match the amplitude of h at $t = 0$. The curve for $x/L = 0.5$ shows an excellent fit to the field data (Fig. 6.14d), suggesting that the Yuanlin station may be

situated roughly at the mid-point of an aquifer between a local groundwater divide and the local discharge location.

The transmissivity (T) and the storativity (S) of the aquifers determined from well tests on the Choshui River fans (Lee and Wu 1996; Kester and Ouyang 1996; Tyan et al. 1996), yielded an average diffusivity $D = T/S \sim 10 \text{ m}^2/\text{s}$. Combining this D with an average value of $\tau \sim 10^6 \text{ s}$ (Table 6.2) and Eq. (6.15), we obtain $L \sim 5000 \text{ m}$ for the confined aquifers. It is interesting to note that this characteristic length for the confined aquifers is considerably shorter than that shown on the geologic cross-sections reconstructed from the hydrological well logs (Water Resource Bureau 1999). This difference might be expected because the actual geologic structure of aquifers in the alluvial fan may be more complex than that as shown on an idealized geologic cross-section.

A simple picture of the plumbing system in the Choshui River fan immediately after the Chi-Chi earthquake emerged from the analysis for post-seismic data. The stepwise rise of groundwater level, as documented by most wells in the unconsolidated sediments of the Choshui River fan, was dissipated by groundwater flow through subhorizontal aquifers with a typical length of 5 km. Aquifers with stepwise decreases of groundwater level, as documented by wells drilled at relatively high elevation near the ruptured fault, were mostly recharged locally by surface runoff, but may also discharge to aquifers at lower hydraulic potential after the earthquake. The aquifers that facilitated this discharge also have a characteristic length of $\sim 5 \text{ km}$, similar to that of the aquifers that facilitated the dissipation of the stepwise rise in the alluvial fan, hinting that similar aquifers in the Choshui River Alluvial fan may be involved in the post-seismic recovery to equilibrium.

6.5 Pore Pressure and Permeability of Continental Faults

Permeability in fault zones controls fluid flow and hence the in situ effective stress and seismic hazard, and is thus a time-honored topic of research. Earlier discussions were mostly based on inferences from geological and geophysical structures and the material properties of fault zones (e.g., Aydin and Johnson 1978; Wang 1984; Wang et al. 1978, 1986; Chester and Logan 1986; Scholz 1990; Sibson 1996; Caine et al. 1996; Schulz and Evans 2000; Bense et al. 2013). These studies converged to a basic model that fault zones consist of a narrow low-permeability, clay-rich core surrounded by a broad zone of highly fractured and damaged rocks (the ‘damage zone’). This model suggests that fault zones are hydrologically anisotropic and may serve as an effective hydraulic conduit for flow along the fault but is an effective barrier for flow across the fault (e.g., Scibek et al. 2016). More recently, instrumental measurements have provided quantitative in situ information about the hydraulic properties of fault zones (e.g., Zoback et al. 2010; Xue et al. 2013, 2016; Brixel et al. 2020a, b). Here we focus on the latter developments.

Drilling into active fault zones has been undertaken in several settings, with a central objective to determine the fault zone properties and pore pressure; these include the Nojima Fault Zone Probe in Japan (Kitagawa et al. 1999), the Chelungpu Fault Drilling Project in Taiwan (Wu et al. 2007), the San Andreas fault near Parkfield in central California (SAFOD; Zoback et al. 2010), and the Wenchuan Fault Scientific Drilling Project in China (Li et al. 2013).

The SAFOD project is probably the best documented and reported in the English language (e.g., Zoback et al. 2010). Drilling was initially vertical on the SW side of the San Andreas fault to a depth of 1.5 km, and then steered 60° from vertical to intersect the active fault zone. After passing through a variety of wall rocks with different lithology, drilling passed through a damage zone ~200 m wide and encountered three actively slipping, clay-rich gouge zones, each 2–3 m wide. From the pore-pressure measurements, Zoback et al. (2010) concluded that the active fault core showed no evidence of high pore pressure. Wang (2011) showed, however, that the very low permeability of the fault core material ($<10^{-20}$ m²; Morrow et al. 2014) implies that the amount of time available during drilling may be much too short for pore pressure to reach equilibrium and thus the measured pore pressure may not represent the equilibrium values.

While it may be difficult to show that the measured pore pressure deep in active faults is at equilibrium, several measurements of fault zone permeability have been made based on the analysis of the tidal response of pore pressure on faults. For example, Xue et al. (2013) analyzed the tidal response of groundwater deep in the active fault that ruptured during the 2008 Wenshuan earthquake in China. Xue et al. (2016) also monitored the tidal response of pore pressure in the vicinity of the San Andreas fault. Here we discuss the former study as an example.

After the 2008 Mw7.9 Wenchuan earthquake (Fig. 6.15a) in Sichuan, China, a drilling project drilled a series of boreholes that penetrated the main rupture zone at a depth of ~1.2 km (Li et al. 2013). Figure 6.15b shows that the borehole intersects a major slip zone at depths >600 m. Pore pressure in the borehole was monitored from January 2010 to August 2011 with a sampling interval of 2 min and a resolution of 6 mm (Fig. 6.16a). Xue et al. (2013) showed that the tidal response of groundwater in the penetrated fault zone had a phase shift from -20 to -30° and an amplitude response from 5.5×10^{-7} to 6.3×10^{-7} m⁻¹. The authors interpreted the negative phase lag as suggesting that the fault zone aquifer was confined and inverted the phase and amplitude responses with the analytical solution of Hsieh et al. (1987) for a confined, isotropic, homogeneous and laterally extensive aquifer. They obtained an average transmissivity of 5×10^{-6} m²s⁻¹ (Fig. 6.16b) and an average storage coefficient of $S = 2.2 \times 10^{-4}$ (Fig. 6.16c). The transmissivity may be converted to an average permeability of $k = 1.4 \times 10^{-15}$ m² by using the identity $k = \mu T / \rho g d$, where $d = 400$ m is the thickness of the fault zone aquifer. The use of the entire open interval of the borehole for the fault zone may have led to an estimated permeability that represents a lower bound on the fault zone permeability (Xue et al. 2013).

The inverted permeability for the ruptured fault measured during the Wenchuan drilling project shows a continuous decline during most of the study period and discontinuous increases during far-field earthquakes (Fig. 6.16b). The continuous

Fig. 6.15 **a** Location map of the studied WFSD-1 site. Red lines in the inset indicate the main rupture zone; the red star is the epicenter of the Wenchuan earthquake. In the sketch **b** the black line is the fault core, which is surrounded by the damage zone. The borehole is 1201 m deep, and 800–1201 m is the open interval where water can flow into the hole from the formation (white arrows). The fault that was most likely active during the Wenchuan earthquake is the major lithological boundary between the Precambrian complex and the Triassic sediments at 590 m (from Xue et al. 2013)

decline of permeability was interpreted to represent post-seismic healing of the fractures generated by the Wenchuan earthquake (Xue et al. 2013) that was interrupted intermittently by seismic waves from remote earthquakes.

The permeability of the Chelungpu fault that ruptured in the 1999 Chi-Chi earthquake was also determined in a drilling project (10^{-18} – 10^{-16} m²; Doan et al. 2006). The experiment was designed to determine the permeability along the ruptured fault, with cross-hole pumping experiments between two boreholes, separated by ~40 m, drilled across the ruptured fault (Fig. 6.17). Because the flow during the pump tests

Fig. 6.16 **a** Water levels from the borehole recorded from 1 January 2010 to 6 August 2011. The oscillations in the inset are generated by Earth tides. The precision of the water level measurement is 6 mm. The measured ‘water level’ is the height of water above the pressure transducer. **b** Permeability and transmissivity and **c** storage coefficient. Values were inverted from the phase and amplitude of each 29.6-day segment based on the analytical model of Hsieh et al. (1987). The black dots denote unconstrained inversion; the red dots are the results of inversion with the storage coefficient fixed to a single value. Because the two separate inversions have identical results for transmissivity, the red dots cover the black dots in (b). The vertical dashed lines show the time of the selected teleseismic events, which correspond to sudden increases in permeability. The best-fit linear trends between each set of permeability increases are shown as light gray dashed lines (from Xue et al. 2013)

would preferentially move along the high permeability damage zone along the fault, rather than along the low permeability fault core, the permeability so measured would represent that of the damage zone. This may explain the large difference between the permeability of the recovered fault cores from the SAFOD project (Morrow et al. 2014) and the in situ permeabilities from the two drilling projects. The difference between the permeabilities determined in the two in situ experiments is relatively small, and may be due to the different methods used in the two field experiments, the different fault zone lithologies in the two field sites (a damaged zone in a clay-rich Quaternary shale of the Chelungpu drilling site versus the damaged zone in a consolidated Triassic sedimentary sequence in the Wenchuan drilling site), and the different spatial scales involved in the two measurements (e.g., Ingebritsen et al. 2006; Kinoshita and Saffer 2018).

Detailed measurement of fault permeability based on boreholes drilled into granitic rocks from an underground rock laboratory in the Swiss Alps (e.g., Brixel et al. 2020a, b) showed that the permeability of fault zones in these rocks may fall sharply from 10^{-13} to 10^{-21} m² within 1–5 m from the fault. These studies suggest that fault permeability may be strongly anisotropic and the flow patterns near faults may be complex. At the same time, we should note that the thicknesses of these faults are orders of magnitudes smaller than those of the major faults, such as those in the San Andreas fault zone in central California and the Wenchuan fault zone in China. As a result, the physical properties and architecture of the faults may also be fundamentally different.

Fig. 6.17 Cross-hole experiment along the ruptured Chelungpu fault (red) in the 1999 Chi-Chi earthquake (from Doan et al. 2006)

6.6 Pore Pressure and Permeability on the Ocean Floor

Research into pore pressure within the oceanic crust began in the late last century (e.g., Davis et al. 2001, 2006, 2009; Vinas 2013; Akmal 2013; Hornbach and Manga 2014; Kinoshita et al. 2018), largely as a part of the Ocean Drilling Program (ODP). Here we review two representative studies in different tectonic settings, first in a subduction zone (Kinoshita et al. 2018), and next near an oceanic spreading ridge (Davis et al. 2001).

6.6.1 Pore Pressure and Permeability in an Accretionary Prism

The example of pore pressure and permeability measurements in accretionary prisms comes from the Nankai Trough (Fig. 6.18; Kinoshita et al. 2018), offshore of SW Honshu, Japan. A splay fault was penetrated by the C0010 borehole to a depth of 407 m beneath the seafloor; pore pressure was monitored for 5.3 years, and showed a sequence of small changes during earthquakes with implications for the poroelastic properties of the fault.

Kinoshita et al. (2018) measured pore pressure in the fault zone and the oceanic tidal loading on the seafloor. The two sets of measurements show no phase lag (Fig. 6.19a) and are linearly correlated with a mean loading efficiency (slope of pore pressure versus load, Eq. 3.81) of ~ 0.74 (Fig. 6.19b). The lack of a phase delay between the pore pressure response and the reference loading implies a high hydraulic diffusivity of the aquifer connecting the formation to the borehole and allows the authors to set a lower bound of $6.4 \times 10^{-13} \text{ m}^2$ on the fault zone permeability that

Fig. 6.18 Location of drillsites in the Nankai Trough Seismogenic Zone Experiment (Kinoshita et al. 2018). **a** Map of the drilling sites (circles). The borehole installed at Site C0010 is shown by the red circle, which penetrates a shallow fault at 407 m below seafloor. **b** Schematic image of the cross section along the dashed line in (a); red vertical line shows the location of the C0010 borehole (from Kinoshita et al. 2018)

is broadly consistent with reported fault zone permeabilities from other subduction zones (e.g., Fisher and Zwart 1997; Screaton et al. 2000; Saffer 2015). The measured loading efficiency implies a formation compressibility ~ 10 times smaller than that measured on the retrieved core sample. Kinoshita et al. (2018) interpreted this lower loading efficiency to imply a small amount of dissolved gas in the interstitial fluids.

Figure 6.19a shows that the measured pressure was dominated by oceanic tidal signals. The latter were removed by using the model of loading efficiency, i.e., $P_{\text{corr}} = P_{\text{form}} - \gamma P_{\text{ref}}$, where γ is the loading efficiency (Sect. 3.2.5), and the subscripts for

Fig. 6.19 **a** and **b** Time series of pore pressure and loading efficiency measured in the fault zone at Site C0010. Vertical lines mark the time of earthquakes that produced detectable changes of pore pressure and loading efficiency. **c** and **d** Coseismic changes of pore pressure and loading efficiency at Site C0010 plotted on diagrams of earthquake magnitude versus the epicentral distance, together with the seismic energy density relation determined on land (Wang 2007). The blue and red circles show events that produced coseismic changes in pore pressure and loading efficiency, respectively. The open circles show events that did not produce detectable coseismic changes. The gray solid and dashed lines represent the contours for constant seismic energy densities of 10^{-3} and 10^{-1} J/m^3 , respectively (modified from Kinoshita et al. 2018)

pressure (P) refer, respectively to the corrected pressure, the formation pressure and the reference pressure on the seafloor. The time series of the corrected pore pressure (Fig. 6.19c) and of the loading efficiency (Fig. 6.19d) are both affected by numerous coseismic changes during the studied period, marked by the vertical lines in the figure. While most of the changes show coseismic increases of pore pressure and decreases of the loading efficiency, the largest of these changes during the March 2011 Tohoku earthquake showed coseismic decreases of both pore pressure and loading efficiency. Also noticeable is that most postseismic recoveries of the loading efficiency have longer recovery time than the post-seismic recovery of pore pressure, suggesting different recovery mechanisms. Kinoshita et al. (2018) suggested that the coseismic decreases of loading efficiency were due to the exsolution of the dissolved gas in the pore fluid in response to dynamic shaking. Because the compressibility of free gas is much larger than that of dissolved gas, this phenomenon would produce coseismic decreases in loading efficiency. At the same time, the redissolution of this gas back into the pore water takes time, resulting in protracted recovery of the loading efficiency. The authors also supported the suggested mechanism by the volume of gas in the recovered pore fluid.

On a diagram of earthquake magnitude versus epicentral distance, the coseismic changes of pore pressure (Fig. 6.19e) and loading efficiency (Fig. 6.19f) follow a systematic linear trend similar to that defined by land-based observations of hydrological response to earthquakes (Wang 2007) and are bounded by a seismic energy density of $\sim 10^{-3} \text{ J/m}^3$. This result suggests that the coseismic responses of pore pressure beneath the seafloor may involve similar mechanisms as those on land.

Pore pressure in accretionary prisms also change during slow seismic events in which slip occurs over weeks to months. Figure 6.20 shows the occurrences of several pore pressure transients in the accretionary prism of the Nankai Trough over a period of six years from 2011 to 2016 in two boreholes (C0100 and C0002) separated by 11 km in the dip direction (Araki et al. 2017). Increases or decreases of pore pressure and occurred during slow seismic events, but the largest two changes occurred right after the Tohoku (March, 2011) and the 2016 Kumamoto (April, 2016) earthquakes, respectively, and showed decreases of pore pressure in both boreholes.

6.6.2 Pore Pressure Changes Near an Ocean Ridge

Davis et al. (2001) investigated pore pressure near the actively spreading Juan de Fuca oceanic ridge (Fig. 6.21a) during an earthquake swarm that began on June 8, 1999. The earthquake swarm lasted more than two months (Fig. 6.21e) and caused pore-pressure transients (Fig. 6.21d) in several boreholes of the Ocean Drilling Program on the eastern flank of the ridge, 25–100 km from the epicenter (Fig. 6.21b). Also recorded are the pore-pressure responses to the tidal loading of the seafloor (Fig. 6.21c). The transient responses to the first earthquake are characterized by a

Fig. 6.20 Summary of pore pressure changes measured in boreholes C0010 (red) and C0002 (blue) (see Fig. 6.17 for locations). Solid circles show pore pressure increases, and open squares show pore pressure decreases. The red lines and arrows in the schematic cross-sections show the location, the amount and the direction of slip that are required to interpret the pore pressure measured in both boreholes assuming a dislocation model (modified from Araki et al. 2017)

rapid coseismic rise in pressure, followed by a continuing slower rise to a peak, and then a much slower decay (Fig. 6.21d), similar to the sustained groundwater responses on land described in Sect. 6.4. As noted by Davis et al. (2001), the pore-pressure transients occurred only during the first earthquake, but not during the latter earthquakes (vertical lines in Fig. 6.21e), even though several of the latter shocks were of greater magnitude than the first one. Davis et al. (2001) accounted for the differences among the pore-pressure responses by suggesting that the pore-pressure change was associated with a much larger tectonic event at the spreading center, most of which occurred aseismically; thus the earthquake was merely the seismic expression of a much larger tectonic event, not the cause of the pore-pressure transients.

Here we offer an alternative interpretation of the non-responsiveness of pore pressure to the latter earthquakes. According to the enhanced permeability model, the first earthquake opened some permeable pathways between the ODP site and local high-pressure sources, which caused the observed increase in pore pressure at the borehole. In order for the second earthquake to cause a second transient increase in pore pressure, sufficient time must pass between the two earthquakes to allow the fluid pathways to seal and the high-pressure sources to re-pressurize. According to this alternative hypothesis, there may simply be insufficient time between the first and the subsequent earthquakes during the two-month span to allow the permeable channels to re-seal and the local sources to re-pressurize. Hence, after the first pore pressure transient induced by the first earthquake, no further pore-pressure transients were possible during the remaining time span of the swarm.

Fig. 6.21 **a** Location map of study area near the Juan de Fuca spreading ridge (thick lines). **b** Cross section of the primary lithology (basement: black; sediment cover: grey) at each of well site. **c** Raw formation pressure record from ODP Site 1024C at the time of the June 1999 earthquake swarm along the Endeavour ridge segment, showing strong tidal signals in response to the loading of the ocean tides. **d** Pore-pressure records from this site and several other ODP Sites after the removal of the responses to tidal, barometric, and oceanic loading. **e** Histogram of the number of events detected; vertical lines show earthquakes recorded at onshore seismic stations (from Davis et al. 2001)

6.7 Concluding Remarks

It is reassuring that some earlier findings on the groundwater response to earthquakes have largely stood the test of time and more observations. The coseismic response of water level in wells in the near field is dominated by step-like changes, while in the intermediate and far fields sustained changes and groundwater oscillations dominate. Enhanced permeability remains a plausible explanation for the sustained changes of groundwater level during earthquakes. Undrained consolidation remains the most cited explanation for the step-like coseismic changes, but a new mechanism has emerged that suggests that earthquakes may release pore water from unsaturated soils to cause step-like increases of water level in the near field. Laboratory experiments have verified that both mechanisms may explain the step-like coseismic changes. The

minimum seismic energy density for coseismic changes of groundwater level has been extended from 10^{-4}J/m^3 down to 10^{-6}J/m^3 .

New advances in the last decade in understanding the groundwater response to earthquakes were made mostly by using the groundwater response to tidal and barometric forcing to monitor changes in hydrogeological properties. These studies have demonstrated that the hydraulic properties of groundwater systems are dynamic and may change with time in response to disturbances by natural and induced earthquakes. These methods have been applied broadly, including to estimate the permeability of several drilled active fault zones, to identify leakage from deep aquifers used for the storage of hazardous wastewater, and to reveal the potential importance of soil water and capillary tension in the unsaturated zone. On the other hand, it should be noted that the tidal and barometric responses of water level in wells represent local responses and, when applied in a regional context, they need to be considered with hydrogeological investigations and information from deep drilling.

Appendix: Derivation of Eq. 6.13

The governing differential equation is

$$S_s \frac{\partial h}{\partial t} = K \frac{\partial^2 h}{\partial x^2} + A. \quad (6.11)$$

Taking the background head as the reference, the initial condition before the earthquake is $h = 0$ at $t = 0$. A no-flow boundary condition is applied at the local water divide (i.e., $\partial h / \partial x = 0$ at $x = 0$) and $h = 0$ at the local discharge ($x = L$). If A is a function of x only, the solution is given by (Carslaw and Jaeger, 1959, p. 132):

$$h(x, t) = \frac{4L}{\pi^2 K} \sum_{n=1}^{\infty} \frac{1}{n^2} \left[1 - \exp\left(-\frac{Dn^2\pi^2 t}{4L^2}\right) \right] \times \cos \frac{n\pi x}{2L} \int_{-L}^L A(x) \cos \frac{n\pi x'}{2L} dx' \quad (A6.1)$$

For the present problem, however, A is a function of both x and t . In this case, we apply the Duhamel's principle (Carslaw and Jaeger 1959, p. 32) to (A6.1) to obtain

$$h(x, t) = \frac{1}{LS_s} \sum_{n=1}^{\infty} \cos \frac{n\pi x}{2L} \times \int_{-L}^L \int_0^t \exp\left(-\frac{Dn^2\pi^2(t-\lambda)}{4L^2}\right) A(x, \lambda) \cos \frac{n\pi x'}{2L} d\lambda dx' \quad (A6.2)$$

Since the coseismic release of water takes much less time than that for the post-seismic recession of groundwater level, we assume that the coseismic release of water occurs instantaneously; i.e., $A(x, t) = A_o(x)\delta(t = 0)$, where $\delta(t = 0)$ equals to 1 when $t = 0$ and equals to 0 when $t > 0$. Eq. A6.2 then reduces to Eq. 6.13.

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