Chapter 7

Stream Flow

The earth is shaken in various ways, and wonderful effects are produced; ... sometimes great masses of earth are heaped up, and rivers forced out, sometimes even flame and hot springs, and at others the course of rivers is turned. (Pliny the Elder, Natural History, Chap. 82)

Abstract Changes in stream discharge after earthquakes are among the most interesting hydrologic responses because they are visible at Earth's surface and can be dramatic. Here we focus on changes that persist for extended periods but have no obvious source. Such increases have been documented for a long time but their origins are still under debate. We first review some general characteristics of streamflow responses to earthquakes; we then discuss several mechanisms that have been proposed to explain these responses and the source of the extra water. The different hypotheses imply different crustal processes and different water–rock interactions during the earthquake cycle. In most instances, these hypotheses are under-constrained. We suggest that multiple mechanisms may be activated by an earthquake.

7.1 Introduction

It has been known for millennia that earthquakes can induce a variety of hydrological responses. The introductory quotation from Pliny (ca. AD 77–79) describes new flows that appeared after earthquakes almost 2000 years ago. Other examples include the damming of valleys by landslides and rockfalls to form lakes and decrease downstream discharge, creation of waterfalls by the earthquake faulting, and increases of discharge in regions of high relief caused by the avalanche of large quantities of

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snow to lower elevations that increases the supply of melt water. In addition, changes in stream discharge after earthquakes are among the most interesting hydrologic responses partly because they can be directly observed and may be large enough to be visually compelling. Fig. 7.1 shows the destruction of the famed waterfall in Jiuzhaigou, western China, after the 2017 Mw7.0 Jiuzhaigou earthquake, which recovered a few months later.

Such changes only redistribute the surface discharge budget, with excess and deficit flows compensating each other. More interesting and less well understood is a type of discharge change that follows earthquakes and persists for an extended period (commonly several weeks to months) but has no obvious source. Such increases of streamflow are more than curiosities because understanding their origin can provide insight into the interactions between hydrogeologic and tectonic processes at spatial and temporal scales that are otherwise difficult to study. These changes have been quantitatively documented for a long time. For example, extensive networks of stream gauges in the western United States were established by the US Geological Survey

Fig. 7.1 Jiuzhaigou waterfall, China, before (top) and after (bottom) the 2017 M7 Jiuzhaigou earthquake. (from <https://www.sohu.com>)

(USGS) in the early twentieth century, and long and continuous gauging measurements have been collected. Such USGS data, along with measurements made globally, record streamflow changes following earthquakes. In the following sections, we first discuss some general characteristics of streamflow responses to earthquakes that have a subsurface origin. We then discuss several mechanisms that have been proposed to explain these responses and the source of the extra water. Following this, we discuss the available observational constraints on these hypotheses and how different models perform when tested against constraints. Finally, we discuss the occurrence of streamflow responses in special geologic settings, such as that in geothermal areas.

7.2 Observations

7.2.1 *Measurement with Flow Meter and Tape*

Most creeks are too small to have a permanent stream gauge installed to measure their discharge. The discharge of these creeks is more often determined manually by using measuring tapes and flow-meters (Fig. 7.2). The cross-section of the creek is subdivided into several subsections and the discharge across each subsection is determined by measuring its average depth and the depth averaged flow velocity (e.g., Fetter 2001). The discharge across each subsection is then calculated; the total discharge of the stream is then the sum of the discharges across all the sub-sections.

After the 2014 Mw6.0 South Napa earthquake, which occurred during a prolonged drought in California when most creeks in central California were nearly dry, many creeks within about 50 km of the epicenter (Fig. 7.3a) showed increased discharge by a factor of more than an order of magnitude (Fig. 7.3b, d, e). Except Sonoma Creek where discharge was measured automatically by a USGS stream gauge (Fig. 7.3b), the discharge in all the other creeks (Fig. 7.3d, e) was measured manually (Wang and Manga 2015).

7.2.2 *Measurement with Stream Gauges*

Many streams with appreciable discharge are monitored with stream gauges. For example, within the USA, at stream gauges maintained by the US Geological Survey, the elevation of the water surface (stage) at a given location is monitored and converted to discharge using a stage-discharge rating curve constructed for that section (e.g., Fetter 2001). The data are collected at 15-min interval and relayed to USGS offices via satellite and are available for viewing within minutes of arrival.

As far as can be determined from the stream-gauge records (e.g., Fig. 7.3b), the onset of streamflow changes can be coseismic. The change can, however, continue

Fig. 7.2 Measuring discharge by using a flow meter and tape. Two hydrologists from the Sonoma Ecology Center measure the discharge in a small stream in the Sonoma County, California

for a few days or weeks to reach a maximum, and then gradually declines to reach the pre-earthquake level after several months. Also noticeable in Fig. 7.3 are the sudden increases in streamflow in response to precipitation. Precipitation can easily obscure the earthquake-induced streamflow response when it occurs at the time of an earthquake and makes the analysis of the latter difficult or impossible.

The majority of coseismic streamflow responses documented this way show increased discharge (e.g., Rojstaczer and Wolf 1992; Muir-Wood and King 1993; Rojstaczer et al. 1995; Sato et al. 2000; Montgomery et al. 2003; Wang et al. 2004a; Wang and Manga 2015; Mohr et al. 2015, 2017). But coseismic decrease of stream discharge has also been reported. An interesting example is the response of the Shira River to the 2016 Mw7.0 Kumamoto earthquake in Japan, documented by three stream gauges (see Fig. 7.4a for gauge locations; Hosono et al. 2019). The uppermost gauge, Gauge A (Fig. 7.4b), documented a coseismic increase of stream discharge, which was followed by a second larger increase ~ 10 h later, while the lower Gauges B and C (Fig. c and d) both showed coseismic decreases of discharge, followed by large increases ~ 10 h later. Ichiiyanagi et al. (2020) estimated from these records that an amount of approximately $900,000 \text{ m}^3$ of river water was lost between Gauge A and Gauge C in a 12-h period, consistent with the estimate of Hosono et al. (2019) from the change of groundwater level in the area (Chap. 6).

Koizumi et al. (2019) studied the stream discharge in central Kyushu at eight monitored stations (Fig. 7.5a) before and immediately after the 2016 Kumamoto

◀**Fig. 7.3** **a** Map showing sampling locations of the streams and a spring (Spencer Spring) that increased flow after the 2014 Mw6.0 South Napa earthquake, monitored sites on the perennial streams, and the USGS stream gauges. Red lines show the West Napa fault zone; thick red line shows the ruptured fault. The focal mechanism of the earthquake, shown by a ‘beach ball’ symbol, divides the surrounding region into quadrants of static compression and dilatation, bounded approximately by the thick black lines on the map. Areas extending from the white sections of the ‘beach ball’ are in static dilatation; areas extending from the black sections are in static compression. Notice that while the Sonoma Valley is located in a compressional regime, the Napa Valley is located in a tensional regime. **b** Daily averaged discharge of Sonoma Creek documented by a USGS gauge at Agua Caliente before and after the South Napa earthquake, showing changes in stream discharge after the South Napa earthquake. Measurement errors are similar to the size of symbols. Short duration increases of discharge, indicated by thick arrows, were due to precipitation both inside and outside of the studied area, and do not necessarily correspond to the precipitation in the valley **c**. Thin arrow shows the time of the earthquake. Curves show simulated stream discharge based on the model of coseismic vertical recharge (Sect. 7.4.2). Two simulations were made for the Sonoma Creek discharge; the first simulation (black line) is based on data for the first 17 days after the earthquake (open squares) to exclude the first incursion of precipitation and the second simulation (red line) is based on all data before significant precipitation in the valley (60 days after the earthquake). The similarity between the two simulated results and data suggests robustness of the model. **d** Discharges as a function of time in streams and Spencer Spring in Napa County. Discharges in different streams and spring are shown by different colored symbols; measurement errors are shown as error bars except where the error bars are smaller than the symbols. Symbols in brackets show conditions reported by local residents and one discharge data converted from early depth measurements, with depth-to-discharge conversion calibrated during subsequent surveys. Colored curves show simulated stream discharge based on 60 days of data after the earthquake using the coseismic vertical recharge model (Sect. 7.4.2); dashed curves show extrapolations from the simulated discharges. Measurement at Spencer Spring started 21 days after the earthquake and was not simulated. **e** Discharges and simulated discharges as a function of time in new streams in Sonoma County. Upward arrow indicates that measured discharges were off scale (modified from Wang and Manga 2015)

earthquake. They showed that, while some rivers exhibited coseismic increase of discharge (Fig. 7.5b), most increases occurred during heavy rainfall that can obscure a coseismic increase.

7.3 Proposed Mechanisms

In the absence of recent precipitation or snowmelt, an increase in stream discharge implies either an increase of the hydraulic gradient created by a new source, or an increase of the hydraulic conductivity of the aquifer along the flow path, or both, see Eq. (2.3). Several mechanisms have been proposed to explain the changes in stream-flow following earthquakes, which may be generally separated into ‘new source’ and the ‘enhanced conductivity’ categories. Mechanisms involving new sources include: (1) the expulsion of deep crustal fluids resulting from coseismic elastic strain (e.g., Muir-Wood and King 1993), (2) coseismic consolidation and liquefaction of shallow sediments (Wang et al. 2001; Manga et al. 2003), (3) coseismic release of water from mountains (Fleeger et al. 1999; Wang et al. 2004a), and (4) shaking water out

Fig. 7.4 a Map of area around Central Kyushu showing the epicenter of the 2016 Kumamoto earthquake (yellow star), stream gauge locations (triangles) and pre-existing active faults (red lines). Red triangles show gauges with coseismic increases of stream discharge and blue triangles show gauges with coseismic decreases (modified from Ichiyanagi et al. 2020). **b–d** Time series of relative river water level at gauges A, B and C (locations shown in panel a). Hourly records of precipitation within the catchment area are shown with the bar graphs (modified from Hosono et al. 2019). Notice that Gauge A showed a coseismic increase with the main shock, while Gauges B and C showed coseismic decreases; all were followed by a large increase ~10 h later. The foreshock did not cause significant coseismic change of water level (from Honoso et al. 2019)

of unsaturated soil (Mohr et al. 2015). Mechanisms involving enhanced hydraulic conductivity: (1) the enhancement of the horizontal permeability (Rojstaczer et al. 1995), and (2) the enhancement of the vertical permeability (Wang et al. 2004a). Differences between these different hypotheses are nontrivial because they imply different hydrologic processes during and after earthquakes, and have implications for the nature of groundwater flow paths. In the following we summarize the basic elements of the hypotheses and discuss some implications and problems of each. We then evaluate the proposed hypotheses with the existing data.

7.3.1 Static Elastic Strain

Muir-Wood and King (1993) applied the coseismic elastic strain model proposed by Wakita (1975) to explain the increased stream discharge after the 1959 M7.5

Fig. 7.5 **a** Map of central Kyushu showing locations of streamflow stations (black squares), studied springs (black circle), weather stations (black triangles), major rivers (light blue) and ruptured fault (dark blue). Red areas indicate strong ground motion during the Kumamoto earthquake. **b** Temporal changes in the accumulated flow rate (after the removal of the effect of the average accumulated precipitation) and relative precipitation from 2001 to 2017. σ is the standard deviation from 2001 to 2015. Vertical red and blue broken lines indicate, respectively, the occurrence of the main shock of the 2016 Kumamoto earthquake and of the heavy rainfall 2 months after (modified from Koizumi et al. 2019)

Fig. 7.6 Simplified model for the interseismic accumulation and coseismic release of strain in extensional and compressional tectonic environments. For extensional faulting, the interseismic period is associated with crack opening and increase of effective porosity. At the time of the earthquake, cracks close and water is expelled. For compressional faulting, the interseismic period is associated with crack closure and the expulsion of water. At the time of the earthquake, cracks will open and water will be drawn in. In the case of normal faulting, water can be expelled to the surface at the time of an earthquake and thus immediately affect river flow (shown schematically as surface fountains). For reverse faulting, cracks must be filled from the water table, a slower process that may not be observed in river flow rates (from Muir-Wood and King 1993)

Hebgen Lake earthquake and the 1983 M7.3 Borah Peak earthquake. They argued that changes in the static elastic strain in the crust produced by earthquake faulting cause rocks to dilate or contract and thus saturated cracks in rocks to open or close, resulting in a decrease or increase in the groundwater discharge into streams (Fig. 7.6).

7.3.2 Consolidation and Liquefaction

Consolidation of loose, saturated sediments may expel pore water in a ‘drained’ process if sufficient time is available (Sect. 3.3; Fig. 3.5). During an earthquake, the amount of time available is too short for the pore water to drain; thus, the process

is ‘undrained’ (Sect. 3.4). As numerous field observations and laboratory experiments have shown (Sect. 11.2), undrained consolidation of saturated loose sediments causes pore pressure to increase and eventually the sediments can liquefy. Wang et al. (2001) suggested that the coseismic undrained consolidation of the loose sediments on the Choshui River flood plain during the Chi-Chi earthquake caused the coseismic increase of water level (Fig. 6.2a). Manga et al. (2003) suggested that coseismic liquefaction of loose sediments on floodplains may provide the water for the increases in stream discharge following earthquakes. The occurrence of liquefaction in areas that experienced increased stream flow is suggestive, but direct evidence that associates liquefaction to the increased discharge has not been found (Montgomery and Manga 2003).

7.3.3 *Water Released from Mountains*

Within hours after the 1998 M5.2 Pymatuning earthquake in northwestern Pennsylvania, local residents reported that many wells on a local ridge becoming dry, while other wells in the valleys started to flow. Fleeger et al. (1999) reported the observed changes in groundwater level and suggested that the earthquake increased the vertical hydraulic conductivity of shales beneath the ridge, allowing groundwater to drain from the hilltops. They also used numerical modeling to show that an increase of the vertical permeability by 10–60 times from the pre-seismic values would be needed to reproduce the earthquake effects on groundwater beneath the ridge (Fig. 7.7).

After the 1999 Mw7.6 Chi-Chi earthquake in Taiwan, several stream systems showed coseismic increases in stream discharge, many wells in the foothills above the thrust fault experienced a significant decline in water level, and a tunnel beneath the foothills experienced sudden downpours right after the earthquake (Lin 2000; Yan 2001). Field mapping after the Chi-Chi earthquake also showed numerous new fractures in the hanging wall of the thrust fault (Angelier et al. 2000; Lee et al. 2000, 2002). Wang et al. (2004a) attributed these hydrologic events to the coseismic release of groundwater from mountains through subvertical fractures, which recharges aquifers in the valley, that in turn recharge the local streams. They further provided an analytical model of vertical recharge by groundwater from mountains to simulate the observed changes (Fig. 7.11). More discussion of this model is provided in Sect. 7.4.2.

After the 2016 Mw7.0 Kumamoto earthquake in Japan, several authors (e.g., Hosono et al. 2019; Ichiyanagi et al. 2020; Kagabu et al. 2020) showed that their data are consistent with the model of coseismic release of groundwater from the nearby caldera rim mountains.

7.3.4 *Water Released from Unsaturated Soils*

Following the 2010 M8.8 Maule earthquake, Mohr et al. (2015) reported increased streamflow in the Chilean coastal range (Fig. 7.8) and proposed that water was

Fig. 7.7 Groundwater levels (in feet) over a ridge near Greenville, Pennsylvania, before (left) the 1998 M5.2 Pymatuning earthquake, and changes of groundwater level after the earthquake (right). Contours were reconstructed from values estimated from historical measurements and reports from homeowners and drillers (modified from Fleeger et al. 1999)

Fig. 7.8 Observed (red line) and modeled (blue line) streamflow in a catchment in the Chilean Coastal Range for periods before and after the Maule earthquake. Gray bars show potential evapotranspiration rates (in mm/h). Dashed red line indicates maximum streamflow rates assuming negligible nightly evapotranspiration. Vertical dashed black line shows time of the Maule earthquake (from Mohr et al. 2015)

released from unsaturated soils. In this model the aquifer is recharged by the coseismically released water from the unsaturated zone. The observed and the simulated discharge, subjected to evapotranspiration, show good agreement (Fig. 7.8).

Mohr et al. (2015) estimated that a threshold seismic energy density of 10^2 J/m^3 is required, which is three orders of magnitude greater than the threshold seismic energy density to initiate undrained consolidation (see Sect. 7.4.6). Thus, the mechanism of releasing pore water from the unsaturated zone may be significant only in the near field.

7.3.5 Enhanced Permeability

Following the 1989 M6.9 Loma Prieta earthquake in central California, sudden increases of stream discharge occurred in nearby drainage basins. Rojstaczer and Wolf (1992) and Rojstaczer et al. (1995) proposed that the increased stream discharge was due to coseismic enhancement of the hydraulic diffusivity of the aquifer, with flow governed by

$$\frac{\partial h}{\partial t} = D \frac{\partial^2 h}{\partial x^2}, \quad (7.1)$$

where h is the hydraulic head, D is the horizontal hydraulic diffusivity of the aquifer, x is the horizontal position, L is the length of the aquifer, and t is time. By using this equation, Rojstaczer et al. (1995) has assumed the horizontal diffusivity and its change to explain the increase of stream discharge.

Similar models were applied to the 1995 Kobe earthquake in Japan to explain observed hydrological changes (Tokunaga 1999; Sato et al. 2000). The model of enhanced permeability was also invoked to explain the increased electrical conductivity of water discharged after an earthquake (Charmoille et al. 2005) and to explain the coseismic increases of phase shifts in the water-level response to tidal strain in southern California (Elkhoury et al. 2006).

7.3.6 Enhanced Vertical Permeability

A lively debate ensued about the enhanced permeability model after the finding of Manga (2001) that no significant change of the baseflow recession occurred after the 1989 Loma Prieta earthquake and other earthquakes. We discuss this finding after a brief introduction to the concept of baseflow recession. The discharge of streams following recharge, known as baseflow recession, often shows an approximately linear relationship between the logarithm of the stream discharge (Q) and time, i.e.,

$$\log Q = a - c t \quad (7.2)$$

where a and c are the empirical constants for the linear fit, and t is the time since recharge. A minus sign is placed in front of c , known as the baseflow recession constant, so that c itself is positive; its inverse, i.e., $\tau \equiv 1/c$, is the characteristic time of the stream response to the recharge. As shown in Sect. 7.4.2, the recession constant (c) is related to the hydraulic diffusivity (D) and the characteristic length of the aquifer (L) by

$$c \approx -\frac{\pi^2 D}{4L^2}. \tag{7.3}$$

Manga (2001) analyzed the hydrographs of a number of streams, including some that responded to the 1989 Mw6.9 Loma Prieta earthquake (Fig. 7.9a). Figure 7.9b plots the recession constant c determined from the hydrographs as a function of time; no significant change in baseflow recession was found before and after the earthquake, even though discharge increased by an order of magnitude after the earthquake. Given that the length of the aquifer is not likely to change during an earthquake and the

Fig. 7.9 **a** Hydrograph of the San Lorenzo River, CA, showing postseismic response to the 1989 Mw6.9 Loma Prieta earthquake. The vertical line indicates the time of the earthquake. The postseismic period of baseflow recession is shown by the bold sloping line. **b** The baseflow recession constant for periods of baseflow before and after the earthquake shows that even though discharge increased by an order of magnitude after the earthquake there was no significant change in baseflow recession. Figure made with US Geological Survey stream gauge data (from Manga 2001)

Fig. 7.10 Stream discharge (logarithm of Q , in m^3/s , daily averages) documented by stream gauge H032, located on a stream in the mountains (see Fig. 7.13 for gauge location). Note the surge in discharge right after the 1999 Chi-Chi earthquake as indicated by the downward pointed arrow. Precipitation within the study area is shown with the bar graphs. Notice that there was little precipitation many months after the Chi-Chi earthquake. Straight lines are the best fits to the baseflow recessions before and after the Chi-Chi earthquake and show similar slopes (modified from Wang et al. 2004a)

similarity in the recession constants before and after the earthquake, the finding of no significant change in baseflow recession after the earthquake (Manga 2001) implies that the horizontal diffusivity was not significantly enhanced by the earthquake, in contradiction to the conclusion of Rojstaczer et al. (1995).

Similar observations were made later. For example, after the 1999 Chi-Chi earthquake (Wang et al. 2004a), the hydrograph of a stream in the foothills of western central Taiwan (gauge H032, see Fig. 7.13 for location) shows nearly identical recession as that before the earthquake (Fig. 7.10).

After the 2010 Mw8.8 Maule earthquake in Chile, Mohr et al. (2017) analyzed the baseflow recession of eighty streams that experienced increased discharge. Their result (Fig. 7.11) again showed no clear change of baseflow recession after the earthquake and thus does not support the hypothesis of a seismic enhanced horizontal permeability as the mechanism for the observed streamflow anomalies.

Wang et al. (2004a) resolved this apparent dilemma by invoking hydraulic anisotropy; they suggested that the Chi-Chi earthquake enhanced the *vertical* permeability in the nearby mountains that allowed coseismic release of water to recharge the aquifer (Sect. 7.3.1) without significantly affecting the baseflow recession, since the latter is controlled by the horizontal permeability (Eq. 7.3). Based on this conceptual model Wang et al. (2004a) proposed an analytical model (Sect. 7.4.2) and used it to simulate the post-seismic stream discharge in Taiwan after the 1999 Chi-Chi earthquake (e.g., Fig. 7.15). This model has since been applied to simulate the increased streamflow after other earthquakes, such as that after the 2014 South Napa earthquake in central California (Wang and Manga 2015) and that after the 2016 M5.8 Pawnee earthquake in Oklahoma (Manga et al. 2016). The mechanism of enhanced vertical permeability also received direct support from the study of water level changes and tidal response after the 1999 Chi-Chi earthquake in several clustered wells on an alluvial fan near the epicenter (Wang 2007; Wang et al. 2016).

Fig. 7.11 The recession constants after the Maule earthquake plotted against the calculated recession constants before the earthquake, as daily values (m^3/day) over one year each for catchments with observed streamflow response. Error bars are $\pm 1\sigma$ of daily recession constants (from Mohr et al. 2017)

7.4 Model Constraints

7.4.1 Constraints from Earthquake Mechanism

Manga et al. (2003) took advantage of the long record of stream discharge data in the United States collected by the USGS, together with the relatively high rate of seismicity in southern California, to characterize the response of Sespe Creek, California, to several earthquakes (streamflow records go back to 1928). Figure 7.12 shows the location of the stream together with the epicenters and the focal mechanisms of several large earthquakes. Manga et al. (2003) found that the streamflow in the Sespe Creek basin always increased regardless of whether the earthquake-induced static strain in the basin was contraction or expansion. This finding rules out the static strain hypothesis as a viable mechanism for the coseismic increases of streamflow, at least for this basin.

Fig. 7.12 Map showing the Sespe Creek basin in southern California and the location of stream gauge, together with the epicenters and focal mechanisms of several large earthquakes (grey—streamflow increase, grey dots—possible increase, black—no change). Inset shows the region in the dashed box (from Manga et al. 2003)

Fig. 7.13 Map shows the three stream systems and stream gauges (in black triangles, each labeled by its gauge number) near the epicenter of the Chi-Chi earthquake. Choshui alluvial fan is on west side and the foothills are on east side. Open circles with crosses show well locations. Stream systems are labeled as Choshui S. for Choshui Stream, etc. Tributaries are not labeled. AB marks the location of hydrogeologic cross-section shown in Fig. 6.4d (modified from Wang et al. 2004a)

7.4.2 Constraints from Recession Analysis

The post-seismic baseflow recession of a stream not only contains information about the hydraulic properties of the aquifers immediately after an earthquake, but can also be used to estimate the amount of extra water released by an earthquake. In this

section we show, with an example, how these parameters may be estimated from streamflow data.

Seventeen stream gauges were installed on three stream systems near the Chi-Chi earthquake epicenter (Fig. 7.13). During and after the Chi-Chi earthquake, many of these gauges registered large increases in stream discharge (Water Resource Bureau 2000; Wang et al. 2004a) and are used here as examples to illustrate how to analyze the postseismic recession and to estimate of the amount of streamflow increase.

The values for c and τ for a number of streams are obtained by fitting the stream hydrographs with Eq. 7.2. The results are listed in Table 7.1. Although these values are entirely empirical, they are closely related to the geometry and the physical properties of the aquifer that recharges the stream.

Since the aquifers are approximately horizontal with a length scale much greater than their thickness, Wang et al. (2004a) approximated them with a one-dimensional aquifer that extends from a local water divide (at $x = 0$) to a local discharge (at $x = L$), as shown in Fig. 7.14b. High-angle fractures formed during the earthquake facilitate the coseismic release of water from mountains (Fig. 7.14a, d) to recharge the aquifer below, which in turn recharges the local stream.

The baseflow recession in this case is determined by the time-dependent discharge of the aquifer at $x = L$. This may be determined by solving the groundwater flow equation under appropriate boundary and initial conditions,

$$S_s \frac{\partial h}{\partial t} = K \frac{\partial^2 h}{\partial x^2} + A(x, t) \quad (7.4)$$

where S_s and K are, respectively, the specific storage and the hydraulic conductivity of the aquifer, and are related to D (diffusivity) in Eq. 7.1 by $D = K/S_s$. Wang et al. (2004a) assumed that the enhanced vertical permeability is high such that the rate of vertical recharge to the aquifer per unit volume, i.e., $A(x, t)$ in Eq. 7.3, occurs coseismically, i.e., $A(x, t) = Q_o(x)$ at $t = 0$. Even though this model is highly simplified, several studies (e.g., Roeloffs 1998; Manga 2001; Manga et al. 2003, 2016; Brodsky et al. 2003; Wang et al. 2004a; Wang and Manga 2015) have demonstrated that such approximations are useful for characterizing the catchment-scale response

Table 7.1 Recession constant c and characteristic time τ from recession analysis of some stream gauge data after the Chi-Chi earthquake

Streams	Stream gauges	$c(s^{-1})$	$\tau(s)$
<i>Wushi system</i>			
Wushi	H025	5.2×10^{-7}	1.9×10^6
Beikang	H032	5.0×10^{-7}	2.0×10^6
Nankang	H037	5.0×10^{-7}	2.0×10^6
Wushi	H042	4.7×10^{-7}	2.1×10^6
<i>Choshui system</i>			
Choshui	H057	3.5×10^{-7}	2.9×10^6
Choshui	H058	6.7×10^{-7}	1.5×10^6
Choshui	H063	5.0×10^{-7}	2.0×10^6

Fig. 7.14 Conceptual model for the coseismic release of groundwater from mountains. **a** Cartoon showing coseismic groundwater release from mountains to recharge an underlying aquifer (modified from Wang and Manga 2015). **b** The model aquifer between a local water divide located at $x = 0$ and a local discharge located at $x = L$. **c** The boundary conditions: the gradient of the groundwater head (i.e., dh/dx) is zero at the local water divide ($x = 0$), and $h = 0$ at the local discharge ($x = L$). **d** The initial condition: recharge to the aquifer at $t = 0$ is Q_0 for $x \leq L'$ and zero for $x > L'$ (modified from Wang et al. 2004a)

Fig. 7.15 Logarithm of the post-seismic excess discharge in m^3/s (dots) plotted against time after the Chi-Chi earthquake at stream gauge H032 adjusted to a reference of $q_{ex} = 0$ before the earthquake, compared with the predicted excess post-seismic discharge (curve) using the model of coseismic vertical recharge (from Wang et al. 2004a)

of hydrological systems to earthquakes. Equation (7.4) is also the linearized form of the differential equation that governs the groundwater level in unconfined aquifers, but with S_s replaced by S_y/b where S_y is the specific yield and b the saturated thickness of the unconfined aquifer. Because these equations are linear, the head change due to the earthquake may be superimposed on the background hydraulic head.

For boundary conditions, we adopt a no-flow boundary condition at $x = 0$ (i.e., a local water divide) and $h = 0$ at $x = L$ (i.e., a local discharge to a stream) (Fig. 7.14b). Taking the background head as the reference value, we have the initial condition $h = 0$ at $t = 0$. The solution for Eq. (7.4) under these boundary conditions was derived

in Sect. 6.4.4 and is given below,

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

where $Q_o(x)$ is the coseismic vertical recharge distribution at $t = 0$. Differentiating (6.13) with respect to t we have

$$\frac{\partial h}{\partial t} = \frac{1}{LS_s} \sum_{n=1}^{\infty} \left(\cos \frac{n\pi x}{2L} \right) \left(-\frac{Dn^2\pi^2}{4L^2} \right) \exp \left(-\frac{Dn^2\pi^2}{4L^2} t \right) \times \int_{-L}^L Q_o(x') \cos \frac{n\pi x'}{2L} dx'. \quad (7.5)$$

For sufficiently long times after the earthquake, Eqs. (6.13) and (7.5) are dominated by the first term ($n = 1$) of the series expansion, i.e.,

$$h(x, t) \approx \frac{1}{LS_s} \cos \frac{\pi x}{2L} \exp \left[-\frac{D\pi^2}{4L^2} t \right] \int_{-L}^L Q_o(x') \cos \frac{\pi x'}{2L} dx' \quad (7.6)$$

and

$$\frac{\partial h}{\partial t} \approx \frac{1}{LS_s} \left(\cos \frac{\pi x}{2L} \right) \left(-\frac{D\pi^2}{4L^2} \right) \exp \left(-\frac{D\pi^2}{4L^2} t \right) \times \int_{-L}^L Q_o(x') \cos \frac{\pi x'}{2L} dx' = \left(-\frac{D\pi^2}{4L^2} \right) h \quad (7.7)$$

Dividing (7.7) by (7.6) we have Eq. (7.8), i.e.,

$$c \equiv \frac{\partial \ln h}{\partial t} \approx -\frac{D\pi^2}{4L^2}. \quad (7.8)$$

This model of coseismic recharge of streams by groundwater released from mountains has been used to simulate the increased stream discharge after the 1999 Chi-Chi earthquake in Taiwan (Fig. 7.15) and the 2014 South Napa earthquake in California (Fig. 7.3b, e).

The model also makes it possible to compute the amount of excess discharge in a stream after an earthquake. To make this estimate, Wang et al. (2004a) simplified the function $Q_o(x)$ further to that shown in Fig. 7.14d, i.e., $Q_o(x) = Q_o$ for $0 \leq x \leq L'$, otherwise $Q_o(x) = 0$. Equation (7.6) then reduces to

$$h(x, t) \approx \frac{Q_o}{LS_s} \cos \frac{\pi x}{2L} \left(\frac{2L}{\pi} \sin \frac{\pi L'}{2L} \right) \exp \left(-\frac{D\pi^2}{4L^2} t \right) \quad (7.9)$$

Differentiating (7.9) with respect to x and evaluating the derivative at the stream (i.e., $x = L$) we have the excess discharge

$$q_{ex} = -KA \left. \frac{\partial h}{\partial x} \right|_{x=L} \approx \frac{KAQ_o}{LS_s} \sin \frac{\pi L'}{2L} \exp \left(-\frac{D\pi^2}{4L^2} t \right), \quad (7.10)$$

where A is the cross-sectional area of the aquifer. Representing the amount of recharge to the aquifer by $Q_o V$ where $V = AL'$ and using $D = K/S_s$, we may rewrite (7.9) as

$$q_{ex} = \frac{DVQ_o}{L^2(L'/L)} \sin \frac{\pi L'}{2L} \exp \left(-\frac{D\pi^2}{4L^2} t \right) \quad (7.11)$$

Assuming that the coseismic recharge is eventually discharged as excess stream flow, the amount of excess discharge may then be obtained from the amount of coseismic recharge, i.e., VQ_o in (7.11). Notice that the parameter D/L^2 in the above equation may be calculated from the post-seismic baseflow recession constant listed in Table 7.1. We may then use Eq. (7.11) to fit the streamflow data, with VQ_o and the ratio L'/L being the unknown fitting parameters. Given $D/L^2 = 2.4 \times 10^{-7} s^{-1}$ (Table 7.1) for the post-seismic discharge at stream gauge H032 (see Fig. 7.13 for location), an excellent fit to the data is obtained with $VQ_o = 0.14 \text{ km}^3$ and $L'/L = 0.8$. The latter is consistent with the fact that the stream gauge H032 is located in the foothills where the flood plain is narrow and the station is close to the water divide (i.e., $x = 0$). Using the values of c or τ from Table 7.1 and fitting the stream flow data, Wang et al. (2004a) obtained the amount of excess flow at each stream gauge, as listed in Table 7.2. Summing the excess discharges in the two stream systems (H025 and H058), they estimated a total excess discharge of 0.7–0.8 km^3 from the west-central Taiwan foothills after the Chi-Chi earthquake.

Figure 7.16 shows another case where Sespe Creek, southern California, responded to the 1952 M7.5 Kern County earthquake located 63 km away from the center of the drainage basin. Again, the vertical recharge model Eq. (7.11) predicts an excess discharge that fits the observed postseismic discharge well (baseflow has been added back to the calculated excess discharge). Here the peak discharge occurs 9–10 days after the earthquake, even though the discharge began to increase coseismically. The difference in rise time from that in Fig. 7.15 (~2 days) reflects the differences in the distance between the stream gauge and the location of the coseismic recharge as well as the aquifer diffusivity.

Table 7.2 Estimated excess discharge in some streams after the Chi-Chi earthquake

Streams	Stream gauges	Total excess discharge (km ³)
<i>Wushi system</i>		
Wushi	H025	0.21
Baikang	H032	0.14
Nankang	H037	0.10
Wushi	H042	0.23
<i>Choshui system</i>		
Choshui	H057	0.56
Choshui	H058	0.55
Choshui	H063	0.44

Fig. 7.16 Response of Sespe Creek, CA to the 1952 M7.5 Kern County earthquake. Daily discharge measurements collected and provided by the US Geological Survey are shown with circles. Curve is solution for the excess flow with $L'/L = 0.4$ added to the baseflow, to recover the entire hydrograph. Vertical line shows the time of the earthquake. There was no precipitation during the entire time interval shown in this graph (modified from Manga et al. 2003)

The 2010 M8.8 Maule earthquake in Chile triggered regional streamflow responses across Chile's diverse topographic and hydro-climatic gradients. Mohr et al. (2017) analyzed the stream response and reported that out of 85 responding streams, 78 showed increased flow. Using the methods discussed in this section, they estimated the total amount of excess discharge to be $\sim 1.1 \text{ km}^3$, which is the largest reported to date. Other estimates include $0.7\text{--}0.8 \text{ km}^3$ after the 1999 M7.5 Chi-Chi earthquake (Wang et al. 2004a), 0.5 km^3 after the 1959 M7.5 Hebgen Lake earthquake (Muir-Wood and King 1993), 0.3 km^3 after the 1983 M7.3 Borah Peak earthquake (Muir-Wood and King 1993), 0.01 km^3 after the 1989 M6.9 Loma Prieta earthquake (Rojstaczer et al. 1995), and 10^6 m^3 after the 2014 M6.0 South Napa earthquake (Wang and Manga 2015).

7.4.3 Constraints From Multiple Stream Gauges

The extensive network of stream gauges near the epicenter of the Chi-Chi earthquake (Fig. 7.13) provides another constraint to test suggested hypotheses. Among the three gauged stream systems, two (Choshui Stream and the Wushi Stream) have many tributaries in the mountains, but the third (Peikang Stream) originates on the western edge of the frontal thrust (Fig. 7.13) and does not have any mountain tributaries.

After the Chi-Chi earthquake, all the tributaries in the mountains showed large postseismic streamflow increases (Table 7.2). On the alluvial fan, the Choshui Stream and the Wushi Stream, both with tributaries in the mountains, also showed large increases in streamflow, but comparison between the excess discharge documented at Gauge H057 and H058 (Table 7.2) shows that the discharge increase in the proximal area (H057) of the Choshui alluvial fan was the same as that in the distal area (H058) of the fan, suggesting that there was relatively little contribution of water from undrained consolidation or liquefaction of the sediments on the fan. In contrast, the Peikang Stream system, which does not have tributaries in the mountainous area, did not show any noticeable postseismic streamflow increases. We thus conclude that the excess discharge after the Chi-Chi earthquake originated mostly from the mountains where groundwater stored at high elevations was released by earthquake-enhanced vertical permeability, and any contribution from coseismic consolidation and liquefaction in the floodplain (alluvial fan) must have been volumetrically insignificant.

7.4.4 Constraints From the Threshold Seismic Energy

As for the case of coseismic change of groundwater level, most of the coseismic changes of stream discharge have been documented together with the earthquake magnitude and the epicentral distances. Figure 7.17 shows a compilation of the occurrences of coseismic change of stream discharge, plotted on a distance versus magnitude diagram. Also plotted as a metric are lines of constant seismic energy density (Eq. 6.10). Figure 7.17 shows that the seismic energy density of 0.1 J/m^3 , which concurs with the liquefaction limit (Chap. 8), also delimits the occurrence of coseismic changes of stream discharge.

Figure 7.17 shows the relationship between earthquake magnitude and distance between the epicenter and the center of the gauged basin for streams that responded to earthquakes. Also shown for reference is the liquefaction limit suggested by Papadopoulos and Lefkopoulos (1993), i.e., the maximum distance over which liquefaction was then reported. The coincidence of the liquefaction limit suggested by Papadopoulos and Lefkopoulos (1993) and the limit for the occurrence of coseismic streamflow increase (Fig. 7.17) is suggestive, though the empirical bound proposed by Papadopoulos and Lefkopoulos (1993) has been outdated by more recent compilation of liquefaction occurrences (Fig. 11.8). Furthermore, Montgomery et al. (2003)

Fig. 7.17 Seismically triggered streamflow changes (circles and squares) as a function of earthquake magnitude and distance from epicenter, plotted together with contours of constant seismic energy density (grey lines; Eq. 6.10) (modified from Mohr et al. 2017). Brown circles are data from a compilation of global data (Wang and Manga 2010); solid red squares are data from the 2014 Mw6.0 South Napa earthquake; open red square is data for the Mw5.8 Pawnee earthquake in Oklahoma; black circles are data from the Maule earthquake, and the M7.1 2011 Araucania after-shock in Chilean headwater catchments; blue circles are data from Chile in response to the Maule earthquake. Brown dashed line is an empirical bound for observed liquefaction (Papadopoulos and Lefkopoulou 1993). Inset shows the histogram of responded streams plotted as a function of the estimated seismic energy density in the Chilean catchments after the Maule earthquake (Mohr et al. 2017) and the energy domains for liquefaction (Wang and Manga 2010) and for the release of vadose zone water in nearly saturated sandy soils (Mohr et al. 2015)

searched for a field association between liquefaction and increased streamflow after the 2001 M6.8 Nisqually, WA, earthquake, but found none.

7.4.5 Constraints from Laboratory Experiment

As discussed in Sect. 6.4.1, Breen et al. (2020) carried out laboratory experiments to test the models of consolidation and of water released from unsaturated soils for the coseismic increase in groundwater level and stream discharge. The result of the experiments showed that both mechanisms can explain the observation. These mechanisms are particularly useful to explain an increase in discharge in flat areas away from mountains, such as that after the 2016 Mw5.8 Pawnee earthquake, Oklahoma (Manga et al. 2016), where no other sources for the extra water are apparent. This point serves as a reminder that earthquakes may activate multiple mechanisms, but often only the dominant mechanism is revealed by observation and analysis. Thus, the mechanisms of undrained consolidation and releasing water from the vadose

zone may become important in flat areas like Oklahoma where other competing mechanisms are absent.

7.4.6 Constraints from Chemical Composition of the Excess Flow

Rojstaczer et al. (1995) argued that, for the elastic strain model of Muir-Wood and King (1993) to explain the increased discharge after earthquakes, a large portion of the deep crust needs to be involved in order to account for the extra water in the increased streamflow. The process would require not only a characteristic time far exceeding that observed in the earthquake-induced stream discharge but also would impart a distinct chemical signature in the discharged water from the deep crust. Following the 1989 Loma Prieta earthquake, Rojstaczer and Wolf (1992) collected water samples from streams near the epicenter and analyzed their chemical composition. They found that, while the stream chemistry showed a marked post-seismic increase in overall ionic strength, the overall proportions of the major ions were nearly the same as those before the earthquake; they argued that these changes were derived from groundwater released from the surrounding highlands instead of from the deep crust.

The hypothesis may also be constrained from the isotopic composition of the post-seismic increased flows. This is because the isotopic composition of rocks is significantly different from that of meteoric water; deep water–rock reactions would thus impart a distinct isotopic signature in the released groundwater. This prediction has been contradicted by several studies of the changes in groundwater composition after earthquakes (e.g., Claesson et al. 2004, 2007; Manga and Rowland 2009; Wang and Manga 2015; Hosono et al. 2020). Detailed discussion on this topic, however, is deferred to Chap. 9 where we focus on the earthquake-induced changes of groundwater composition.

7.5 Streamflow Changes in Hydrothermal Areas

Within 15 min. of the 22 December 2003 M6.5 San Simeon earthquake in central California, two stream gauges registered increased stream discharge, one along the Salinas River near the town of Paso Robles and the other along the Lopez Creek near the town of Arroyo Grande (Fig. 7.18), both known for their hot springs. As explained next, these streamflow increases can be explained by the coseismic recharge model introduced earlier, but apparently driven by the excess pore pressure in a geothermal reservoir, and are thus entirely different from those discussed earlier which were driven by gravitational potential.

Some background information about the local geology and climate may be required to better understand the different responses of these streams. Active tectonics

Fig. 7.18 Map showing the intensity of ground shaking in the 22 December 2003 M 6.5 San Simeon earthquake. Focal mechanism of the earthquake is taken from Harvard CMT Catalog. Circles show locations of stream gauges and triangles show locations of seismometers. Bold line shows the ruptured fault. The Salinas River flows NW through town of Paso Robles and the Salinas Valley (from USGS website)

since the late Tertiary has repeatedly faulted and uplifted the Coast Ranges of California. The climate of the area is semiarid, with most of the annual 250–330 mm precipitation occurring during the winter. A growing population and increased urbanization and agriculture has caused basin-wide decline of the groundwater level during the past several decades. As a result, the streambed of the Salinas River, with a flood plain ~100 m wide through the Paso Robles Basin, is usually dry except during rainy season, and was dry before the San Simeon earthquake. Drilling at Paso Robles encountered a hydrothermal reservoir at a depth of ~100 m. On the other hand, no hot springs are known in the nearby valleys of the San Antonio River or the Nacimiento River.

The epicenter of the San Simeon earthquake occurred 11 km NE of the town of San Simeon and 39 km WNW of Paso Robles. Rupture during the earthquake shows a strong ESE directivity (Fig. 7.18). Four new hot springs appeared after the earthquake on the two sides of the Salinas River (Fig. 7.19) near the town of Paso

Fig. 7.19 Map of Paso Robles showing locations of the stream gauge and four new hot springs that formed after the earthquake. Note that the four new hot springs lie along a straight line that is parallel to the ruptured fault shown in Fig. 7.17. This line intersects Salinas River ~ 1 km upstream of the stream gauge. The location of a hot spring well, established long before the earthquake, is marked by a black triangle (from Wang et al. 2004b)

Robles. These new hot springs occurred along a straight line striking WNW, parallel to the earthquake rupture (Fig. 7.19) and crossing the Salinas River ~ 1 km upstream of the local stream gauge (Fig. 7.19). The well-head pressure at a hot spring well (Fig. 7.19) in Paso Robles was steady before the earthquake, but decreased from 0.33 to ~ 0.2 MPa within 2 days after the earthquake.

Recession analysis of the postseismic stream discharge in the Salinas River and Lopez Creek yields a characteristic time of ~ 40 min, suggesting that the sources of the extra water were close to the surface. However, there was no surface water source in the Paso Robles Basin and any surface water would have to be supplied from distant mountains, which is contradicted by the short characteristic time from the recession analysis. Thus, as suggested by the appearance of new hot springs in the area (Fig. 7.19), Wang et al. (2004b) proposed that the source for the coseismic increase of discharge in the two streams was a subsurface hydrothermal reservoir (Fig. 7.20b) that was sealed above by an impermeable layer (Fig. 7.20c). The seal was ruptured by the earthquake (Fig. 7.20d) and the hydrothermal water erupted to the surface to form the new hot springs and to recharge the stream. An ideal test of this hypothesis would have been a chemical analysis of the increased flow. Unfortunately, the duration of the extra discharge was short and precipitation in the area started one day after the earthquake, which made such analysis unattainable. Wang et al. (2004b) supported the conceptual model with a simulation. The excess discharge based on this model may be simulated using Eq. (7.11). The simulated result, shown with the curve

Fig. 7.20 **a** Normalized stream gauge data (triangles and diamonds) for the streamflow changes following the 2003 M 6.5 San Simeon earthquake, California. Curve shows the model simulation of the observed hydrographs. **b** Cartoon of the model proposed to explain the hour-long increase in streamflow. Rupturing of the seal of hydrothermal reservoir leads to expulsion of fluid into fracture zone. **c** Enlarged cartoon showing the seal with cracks over the geothermal reservoir. **d** Clogged crack and cleared crack; clearing of a clogged crack during the earthquake significantly increases its permeability and effective length (modified from Wang et al. 2004b)

in Fig. 7.20a, fits the observed postseismic discharge (triangles and diamonds). The estimated excess discharge after this earthquake ranged from 10^2 to 10^3 m³, orders of magnitude smaller than the examples mentioned in Sect. 7.4.2.

Abrupt increases in streamflow and hot spring discharge after earthquakes were also reported in other hydrothermal areas such as in the Long Valley, California (Sorey and Clark 1981), in the Napa Valley, California (Wang and Manga 2015), and in Japan (Mogi et al. 1989), suggesting that this type of hydrologic response may be common in hydrothermal areas. Such discharge may also cause changes in the temperature and the chemical composition of the streams and hot springs, as reported by Mogi et al. (1989) and Hosono et al. (2018), among others, and are discussed further in Chaps. 8 and 9.

In convergent tectonic regions, large volumes of pore water may be locked in the subducted sediments (Townend 1997) or beneath volcanic areas (Hartmann 2006). Sealing may be enacted partly by the presence of low-permeability mud, partly by precipitation of minerals in fractures and pores, and partly by the prevailing compressional stresses in such tectonic settings (Sibson and Rowland 2003). Earthquakes may rupture the seals and allow pressurized pore water to erupt to the surface and recharge streams. Husen and Kissling (2001) suggest that postseismic changes in the ratio of P- and S-wave velocities above the subducting Nazca Plate reflect fluid migration into the overlying plate following the rupture of permeability barriers. This process may explain the time variations in submarine fluid discharge at convergent margins (Carson and Screaton 1998). Episodes of high discharge are correlated with seismic activity having features similar to tremor and are not correlated with large regional earthquakes (Brown et al. 2005).

7.6 Concluding Remarks

The different hypotheses discussed in this chapter imply different crustal processes and different water–rock interactions during an earthquake cycle. In most instances, these hypotheses are under-constrained. A reasonable approach is to test the different hypotheses against cases in which abundant and accurate data are documented such as the 1999 Chi-Chi earthquake in Taiwan and the 2016 Kumamoto earthquake in Japan. We may note that a single explanation need not apply to all cases of coseismic increased streamflow, and that multiple mechanisms may be activated by an earthquake.

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