Supplementary Materials

Explanation of Methods

1. Temperature calculation

Temperature is calculated by the integration of equation (1) together with the flow velocity q_z of groundwater. To evaluate q_z we follow Manga and Rowland (2009) by approximating q_x in equation (3) with an averaged head gradient between the fault zone and the discharge zone:

$$q_x = -K_H \frac{dh}{dx} \approx -K_H \frac{h_o - h}{D},$$
(S1)

where K_H is the horizontal conductivity of the aquifer outside the fracture zone, h_o the head at the discharge end of the aquifer and D the horizontal distance to the discharge end. Replacing q_x in (3) by (S1) we have

$$\frac{S_y}{L} \frac{\partial h}{\partial t} = K_V \frac{\partial^2 h}{\partial z^2} + K_H \frac{h_o - h}{wD}.$$
(S2)

This equation is identical to equation (2) in Manga and Roland (2009) with the specific storage S_s replaced by S_y/L. The solution for h and the corresponding vertical discharge q_z, under the boundary conditions of $h = 0$ at $z = 0$ and $\partial h/\partial z = 0$ at $z = L$, are given by equations (6) and (7) in Manga and Roland (2009). The solutions for q_z at depth intervals of 1 m and time intervals of 1 hour are inserted into a finite-difference scheme (S3) to integrate (1) for T, starting from the temperature profile before the earthquake.

$$T(z_i, t) = T(z_i, t\text{-}dt) + dt * \kappa * [T(z_{i+1}, t\text{-}dt) - 2.0 * T(z_i, t\text{-}dt) + T(z_{i\text{-}1}, t\text{-}dt)]/(dz*dz)$$
$$- 0.5*dt*\gamma *q_z(z_i, t)* [T(z_{i+1}, t\text{-}dt) - T(z_{i\text{-}1}, t\text{-}dt)]/dz.$$
(S3)

2. Estimate modeling parameters

The effective horizontal permeability for the aquifers at the upper rim of the alluvial fan, composed of gravels and coarse sands intercalated with marine mud, is given by

$$K_H = \frac{1}{L} \int_0^L K(z)\, dz$$
(S4)

where L is the saturated thickness of the layered sequence. It is controlled by the layer with the highest conductivity and an average value of $\sim 10^{-4}$ m/s is estimated from well tests (Tyan et al., 1996). Baseflow recession analysis of stream discharge (Wang et al., 2004a) shows that K_H did not change after the earthquake.

On the other hand, the effective vertical conductivity (K_V) of the sequence changed after the earthquake. Before the earthquake, K_V is given by

$$K_{Vi} = \left[\frac{1}{h} \int_0^L \frac{dz}{K(z)} \right]^{-1} \tag{S5}$$

and is controlled by the layer with the lowest conductivity (i.e., aquitards); an average value of $\sim 10^{-8}$ m/s is estimated from lithological logs and tabulated conductivity measurements. After the earthquake, the aquitards were breached by vertical fractures; the new effective vertical conductivity, K_{Vf}, may be estimated from the post-seismic 'recession' of the water level in wells (Wang et al., 2004b), with an average value of $\sim 10^{-5}$ m/s. A specific yield $S_y \sim 0.2$ is estimated for the sandy and gravelly aquifer (Fetter, 2001, Table 3.5). Finally, we approximate the width of the fracture zone $w \sim 10$ km from the distribution of wells that showed a coseismic drop of water level (Wang et al., 2001). The distance to the discharge zone $D \sim 10$ km is based on postseismic recession analysis (Wang et al., 2004b). The basement depth L is likely to vary from well to well; we use $L = 300$ m to include all wells listed in Table S1. The parameters so estimated constitute the "reference model" and are listed in Table 1.

References

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