U.S. DEPARTMENT OF THE INTERIOR

U.S. GEOLOGICAL SURVEY

DOG "BITES" EARTHQUAKE
A HANDOUT FOR THE VALLEY/MOUNTAIN SHAKING MODEL

by

John R. Evans[1] and John M. Coakley[1]

Open-file Report 95-22

[1]USGS, Menlo Park, California

Say What?!

Some time ago we got a very nice letter from one of our most helpful and interested friends—one of the kind souls who allows us to put seismic equipment on her land for the benefit of all. She had purchased a toy bone for her ranch dogs, a bone that *barks* when the dog picks it up. It senses the dog's movement with a simple wobble switch, and plays back the sound of a bark recorded on a small circuit chip.

Our friend reported that when she left the bone on the patio it did not bark during strong winds, but it did bark at other times that the dogs were not playing with it. She thought it might be responding to the earthquakes she sometimes feel at her ranch. Indeed, the bones almost certainly will bark in many earthquakes large enough for people to feel (though, it appears, also at a few random moments). We thought this would be a fun way for younger folks to learn what makes some earthquakes easy to feel and others not so easy to feel.

This handout is designed so you can read the basics or get into the details, as you desire. Feel free to skip over sections called "**Details**" *and, in one case,* "**Even More Details**".

Did You Feel It?

Quite a few things work together to let you feel an earthquake or to keep you from feeling it. These are the main ones:

- **Earthquakes come in many sizes, called "magnitudes". The magnitude of an earthquake is how strong it is** *where it begins*, **deep in the Earth. This place is the earthquake "focus" or "hypocenter", where the rocks break and rumble past each other along the deep "fault plane". The shaking you actually feel travels out from there and arrives where you are somewhat later.**

- **"Magnitude" is often confused with "intensity" but is not at all the same thing. "Intensity" is a measure of the actual** *effect* **of the shaking (on buildings and such)** *in some particular place* **on the Earth's surface. "Intensity" is usually given on a scale of I to XII—Roman numerals 1 to 12. As you will see, there is only a partial relationship between the "magnitude" of an earthquake and its "intensity" where you happen to be at the time.**

Details: Magnitude (often called "Richter magnitude") can be any number at all—the scale is "open ended" and always has been. But the largest earthquakes on Earth are about magnitude 10—rocks are just not strong enough to make bigger ones. The largest events yet recorded on modern seismic instruments were the 1960 Chilean earthquake (magnitude 9.5) and the 1964 Alaskan earthquake (magnitude 9.2). In contrast, the gigantic meteor impact that may have led to the extinction of dinosaurs 65 million years ago (the Chicxulub, Mexico, impact) yielded energy roughly equivalent to a magnitude **14** earthquake, if such could exist! This rough estimate, made by Doctors Chael and Boslough of Sandia National Laboratories, is for a 6-mile diameter lump of rock hitting the Earth at a speed of about 45,000 mph (ouch!).

The small end of the magnitude scale is populated by many *very* small types of tremors—even your footsteps cause tiny "earthquakes". For example, a 50-pound child jumping from a one-foot high step and landing on his or her feet would create an "earthquake" of about magnitude −2 (*minus* two). The smallest natural earthquakes we normally record with our seismographs are about magnitude +1, but still smaller ones are happening all the time. The smallest earthquakes you can feel in most places are usually about magnitude +3.

Even More Details: The magnitude scale is "logarithmic" or "geometric", which means that *adding* one unit to the magnitude *multiplies* the shaking by some amount, no matter where on the scale one is looking. The shaking that affects small buildings (like many homes) roughly doubles for each magnitude unit. The shaking that affects high rises (which sway more slowly) increases *about three times* per magnitude unit (Figure 1). It turns out that the *energy* radiated by the earthquake is about *30 times* bigger for each increase of one magnitude unit. In larger earthquakes, much of this increased energy comes as *longer* shaking, rather than harder shaking.

The intent of magnitude scales (there are many flavors) has always been to provide a single, reliable estimate of an earthquake's true size on the fault plane. These "flavors" are the many different ways the scales have been defined over the years. The original definition of magnitude by Professor Charles Richter was very limited, based on the record made by a *particular type* of seismograph at a *particular distance* from the earthquake, and then only for southern California.

Each of the definitions of earthquake magnitude has good points and bad points. The most uniform definition now used is called the "moment magnitude" (the Chilean and Alaskan values we gave are their moment magnitudes; Figure 1 uses moment magnitudes too). Moment magnitude is based, to no one's surprise, on the earthquake's "moment"—the distance that the rocks slip past each other across the fault plane *times* the size of the fault plane patch that slips *times* the strength of the rocks. The moment magnitude of the 1906 San Francisco, California, earthquake was 7.7; that of the 1989 Loma Prieta, California, earthquake was 6.9 (Figure 2 shows the epicenters of the California earthquakes mentioned in this handout). The disadvantage of this measure of magnitude is that it is harder to compute.

Other forms of the magnitude scale tend to "saturate" at larger values, to stop getting higher as the earthquakes get bigger than some size. Because of this "saturation" of most types of magnitude scales, great earthquakes may be bigger than first reports say.

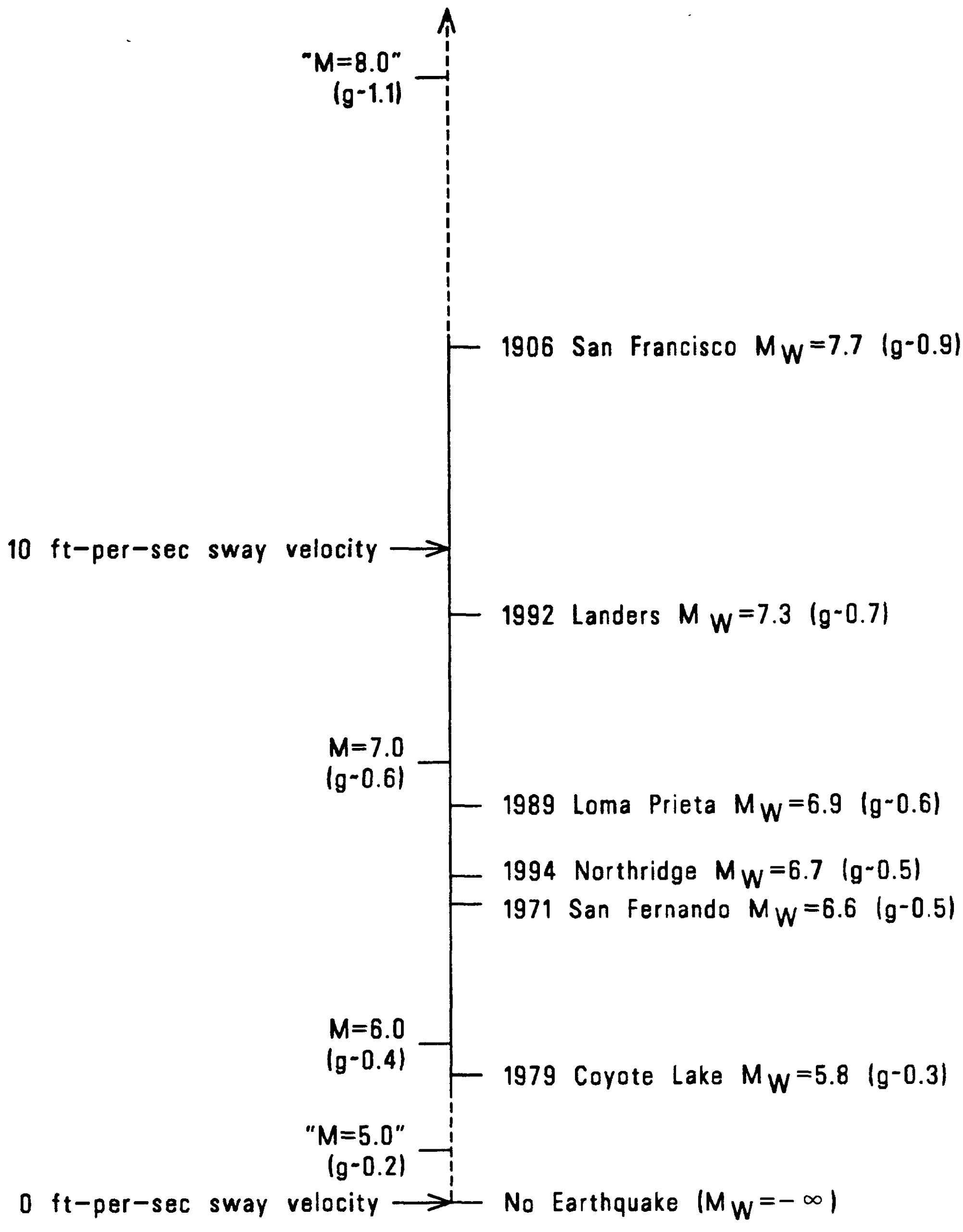

Figure 1. An illustration of the abrupt increase in shaking strength with increasing earthquake magnitude. The length of the line is proportional to the sway speed of a simple 10-story high building near the epicenter. (For larger and smaller earthquakes, shaking is not predicted accurately by the equations used here, so those parts of the line are dashed (Wm. Joyner, personal communication, 1994).) The typical accelerations under the same conditions—hence, the forces you feel—are shown in parentheses. "One g" is the force of Earth's gravity pulling down on us. This Figure is based on the work of Boore et al. *(1993) for buildings built on a soft soil ("type C"), as is much of the San Francisco Bay area.*

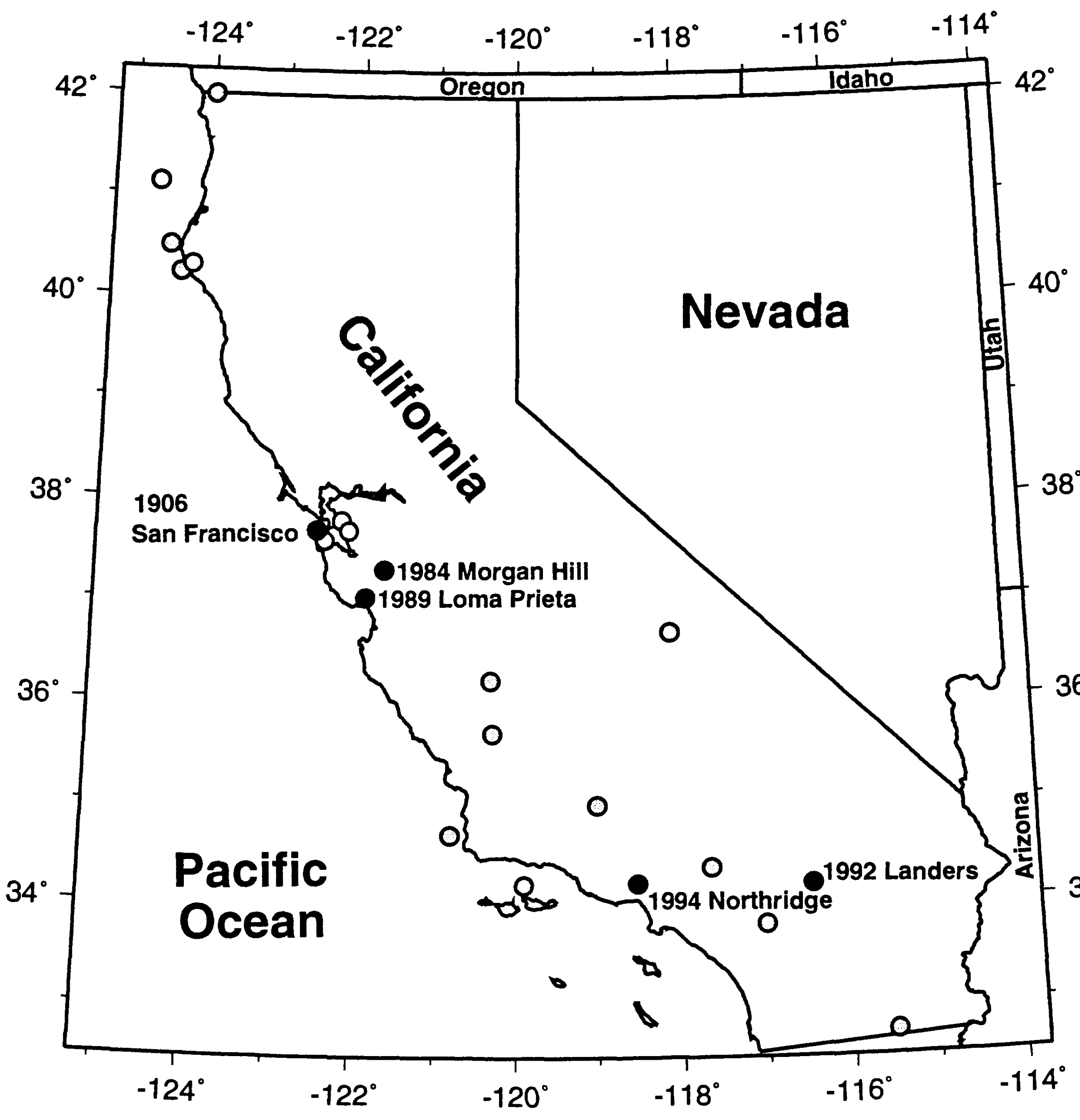

Figure 2. Map showing the California earthquakes mentioned in this handout (black dots) and other large earthquakes happening from 1812 through 1989 (gray dots). These epicenters are the spots on the Earth's surface just above the starting point of each fault-rupture. This figure was plotted with the "GMT" software of Wessel and Smith (1991) using data from Ellsworth (1990) and other sources.

- **Earthquake waves spread out and dissipate as they move away from the earthquake's "focus". So shaking also depends on *how far away* you are from that focus.**

Details: Spreading out of the waves just thins their strength, spreading the same amount of shaking energy over a wider area. The dissipation (fading out) is called "attenuation", and happens because the earthquake waves slowly turn into a small amount of heat in the rocks. The rocks are not perfect "springs" and just slowly consume the energy of the earthquake waves. (Any spring eventually stops shaking for the same reason.)

Attenuation is lower in the eastern United States than in California. Even though large earthquakes happen less often in the east, they are felt over even greater distances than in California.

Also, since most California faults usually slip at several miles depth, your distance from the earthquake "focus" generally is larger than your distance from the fault at the Earth's surface. In rare cases, your distance from the focus can be *less* than your distance from the fault on the Earth's surface. (Most folks think of "the fault" as the place where the fault surface reaches the Earth's surface. But seismologists think of faults as a surface within the Earth, sometimes a complexly shaped surface. Indeed, not all faults reach the surface at all, as was probably the case with 1994 Northridge, California, earthquake (moment magnitude 6.7).) Larger earthquakes break larger patches of the fault surface, often right up to the Earth's surface where we can see the broken ground. Your distance from the *nearest* broken part may control the shaking you actually feel.

- **The type of ground you are on makes a big difference too. Soft ground, like valley sediments—the thick layer of dirt most of us live on—shake more strongly than rocks. The model shows part of the reason for this difference in ground shaking.**

  With the help of a grownup, make an earthquake by pulling the pin out with the cord. Watch the "valley" (sediments) shake more strongly than the "mountains" (hard rocks). Instructions are on page 11.

Details: Soft ground is like a weak spring and rocks are like stronger springs. In fact the ground and rocks are "elastic" just like springs, though they don't squeeze, stretch, or twist nearly as easily as most springs you see. Earthquake waves are just elastic vibrations in the rocks and ground, similar to the vibrations that can happen in springs and similar to the sound waves that you talk with.

The model uses weak springs to imitate the thick layer of sediment beneath most flatlands, like the valleys of the San Francisco Bay Area and southern California. The strong springs represent rocks in the neighboring mountains. The youngest sediments are usually the weakest ones and usually are found in the lowest part of the flatlands (under water, along shorelines, along streams or rivers, and so on).

- **What you are *doing* when an earthquake comes has a *big* effect on whether you feel it.**

If you are:	driving	walking	sitting	lying down
Then you will be:	least aware	→	→	most aware

Details: When you are in a car, even strong shaking sometimes can be mistaken for gusts of wind or other problems. Because of the car's suspension, you may not feel shaking as strongly inside a car, while the normal vibrations of driving can dull your sense of the earthquake shaking. (This does not mean that cars are safe in strong shaking.)

On the other hand, if you are resting in bed you can often feel *very* small earthquakes, even below magnitude 3.

And if the earthquake focus is very far away, but the magnitude is large, you may not sense shaking at all but instead just feel sea sick. A slow rolling motion can affect people at great distances from the earthquake focus.

A short, useful lesson in seismology: Two main groups of earthquake waves result from nearby fault breaks. They are called "*P*" waves and "*S*" waves. "*P*" means "primary", because they travel faster and arrive sooner. "*S*" means "secondary", but *they* carry about **90%** of the shaking energy. (For smaller earthquakes, people sometimes feel only the *S* wave because the *P* wave is so much weaker.) So, if the *P* waves are strong enough to be scary, you usually have a few seconds to find a safe place to hide (less if you are close to the fault break, more if further out). The best thing to do, if you are inside, is to **get under a strong table** and hold on tight. **Please, do *not* run outside—that can be a very dangerous thing to do.** But if you are *already* outside, move into the open, away from things that could fall on you or bounce into you. Earthquakes do not hurt people—things falling or being thrown around are what can do that.

- **Other things can affect how strong the shaking is. These include: ☐ what type of structure you are in or on, ☐ what direction the fault-break travels along the fault, ☐ where strong spots on the fault are (they can break more sharply), ☐ the *shape* of the bedrock beneath the flatland sediment, ☐ the size and shape of hills, ☐ the size and shape of "geologic units" (regions made of different types of rocks), and ☐ even the reflection of waves back from deeper parts of the Earth's crust.**

Details: The natural swaying rate of the structure you are in (or on, in the case of a bridge), along with other aspects of the structure's design, combine with the many factors affecting the amount of earthquake shaking of different rates (frequencies) at your location to amplify or mute that shaking. The actual shaking of structures is described approximately by a "response spectrum"—the response of simple, lightly damped structures to the actual ground motion. (Your car is "lightly damped" when it badly needs new shock absorbers.) The large swaying distances common for tall buildings, and no doubt one's distance from the ground, can amplify the *perception* of shaking, over and above real structural-amplification effects.

The rocks on either side of a fault rub past each other for a short distance during an earthquake (by inches to as much as yards). But the *place* where this breaking happens usually starts in one spot on the fault and moves out from there, sometimes for miles along the fault surface (Figures 3 and 4). It can happen that the fault breaks near one end first so that the break moves out along the fault mostly in one direction. This happened in the 1992 Landers, California, earthquake, which broke from south to north (Figure 4) and sent the strongest shaking north toward open desert (to the benefit of the large cities lying to the south and west). This pointing effect also happened in the Morgan Hill earthquake of April 24, 1984, which broke from northwest to southeast. Folks living around Morgan Hill, near the southeast end of the fault break, got the worst of it.

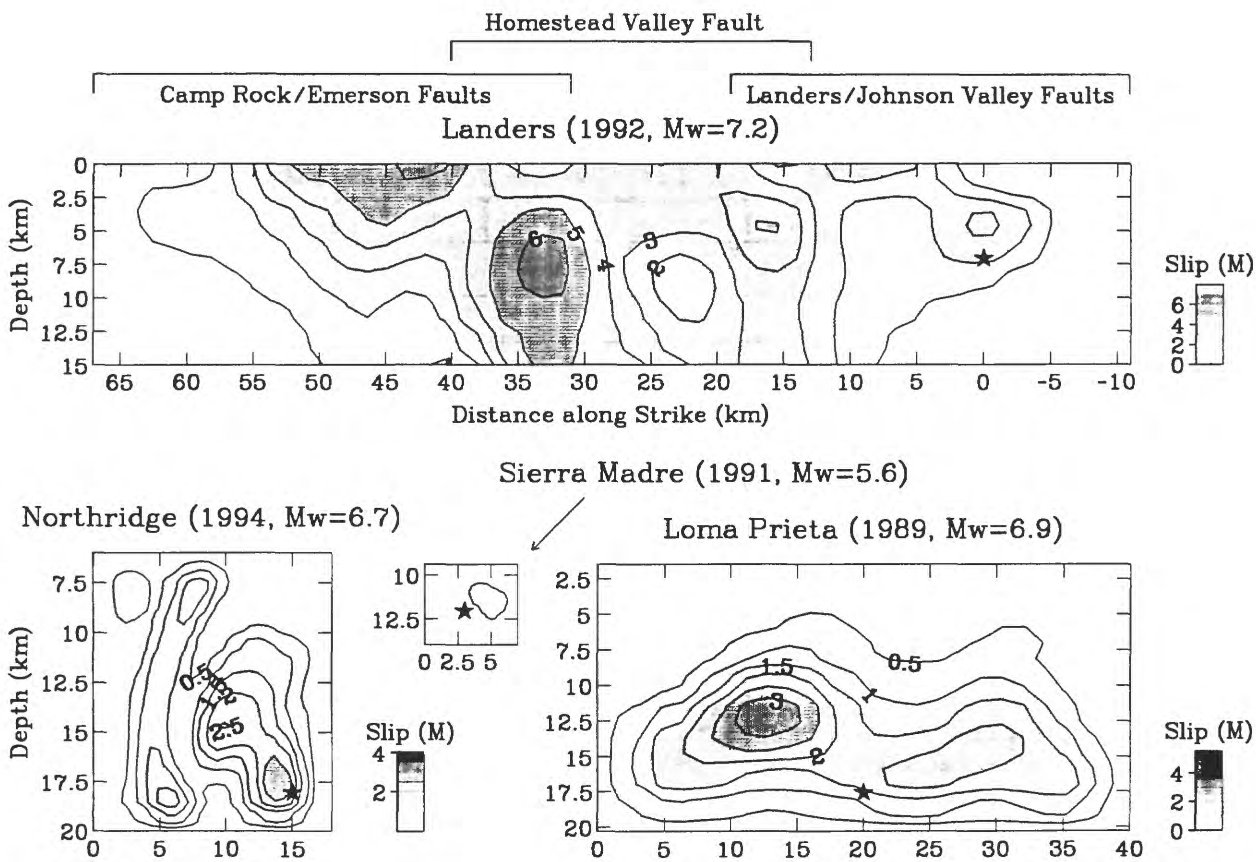

Figure 3. Maps of the fault surfaces that ruptured in the Earth to cause each of four earthquakes. M_W *is the "moment magnitude" of each earthquake. (Observant readers will note the 0.1 unit discrepancy with Figure 1 for Landers—typical of measurement errors even for a well studied event like this one.) The shades of gray and contours show how far the fault slipped at each point. The contour interval is one-meter of slip (~39 inchs) for Landers and 0.5 m for the three smaller events. The Landers earthquake was actually on a complex fault surface composed of the three adjacent, overlapping fault planes named at the top (each of these with its own complexities). The fault rupture began at the south (the right) and jumped from one fault to the next (Figure 4). Modeling and Figure by Wald* et al. *(1991), Wald (1992), and Wald and Heaton (1994*a,b*)*

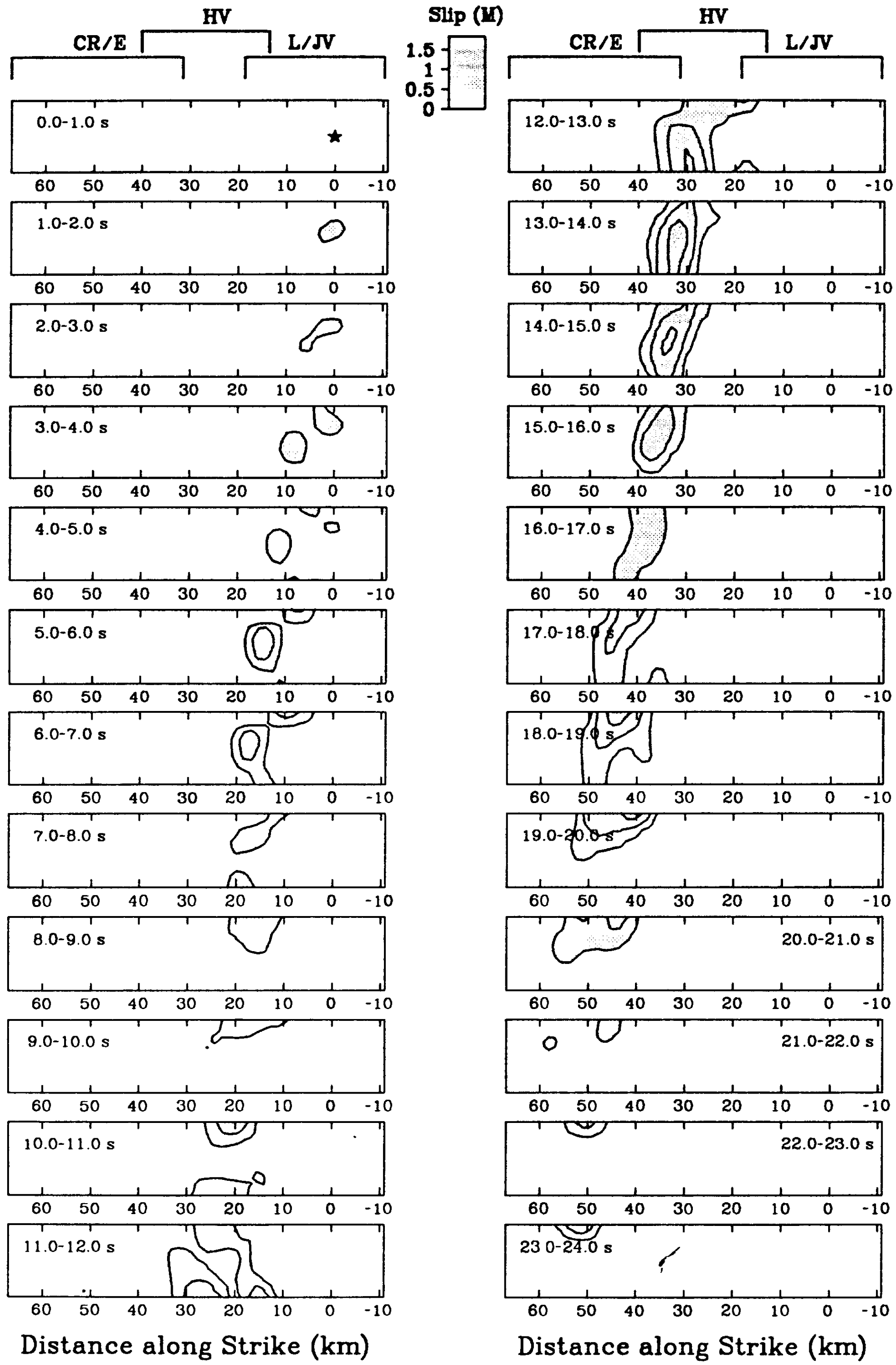

Figure 4. This is a movie lasting 24 seconds of the fault surface rupturing within the Earth (1992 Landers earthquake). At zero seconds (top left frame) the rupture starts at the hypocenter, shown by the star, and propagates north, hesitating a bit where it has to jump from one fault plane to the next. Modeling and Figure by Wald and Heaton (1994).

The shape of bedrock beneath a valley (Figure 5) changes how the earthquake waves echo back and forth in the soft sediment and can concentrate shaking in some places. This effect is very complicated and also depends on where the fault is and how the break moves along the fault surface. This "echoing thunder" can give one neighborhood a very bad ride while mostly sparing nearby areas. Similar things can happen because of the shape of hills or because of boundaries between different rock types. Hilltops with soft rocks may help explain the damage in parts of the Santa Cruz Mountains during the 1989 Loma Prieta earthquake.

According to several recent studies, the reflection of waves back from deeper layers of the Earth's crust added to the shaking in San Francisco and Oakland during the 1989 Loma Prieta earthquake. The *type* of ground in the Marina District and beneath the Cypress Structure (very soft sediment, in some places artificial) clearly also had a major effect.

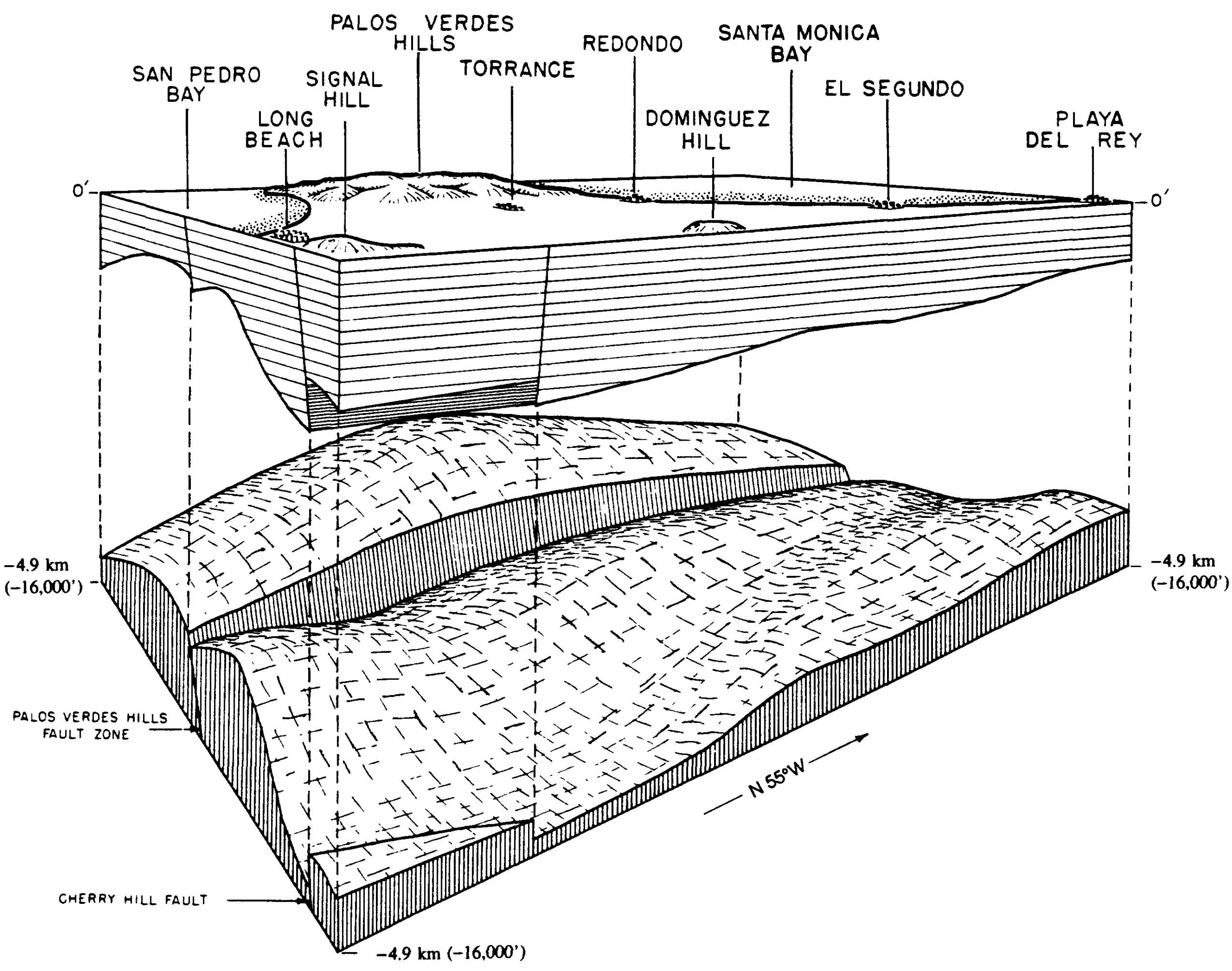

Figure 5. The shape of bedrock beneath western Los Angeles basin, showing the great depth of sediments and various fault surfaces. The stack of sediments (as thick as 4300 meters (14,000 feet) in places) has been lifted up in this Figure to reveal the bedrock. The view is to the west, from land toward sea. Modified from Woodford et al. (1954).

Make Some Earthquakes!

The model shows what happens to earthquake waves when they get into soft ground, like the sediment filling many valleys and lowlands.

The thin springs on the left are weak like the weaker sediment, while the thick springs on the right are stronger like the rocks in nearby hills (and beneath the sediment). The platform on the left, supported by the weak springs, will shake more than the one on thick, strong springs. Shaking in valleys is likely to be stronger than in neighboring hills, if other things are equal (hills do have other problems, like landslides).

There are two ways to use this model. (I) gives you a pair of seismograms to take home; (II) makes the bones bark. (I) works every time, but (II) is inconsistent. **Instructions:**

I. (1) Push back the board with the lead weights on it and put in the **longer** of the two pins (the one to your right). The cord end of the pin—purple—fits into a small groove on the back of the angle iron that holds the green spring. *Be careful not to get it into the larger hole just below that groove!* The other end of the pin—red—fits into the dimple in the steel plate on the front edge of the moving board, below the red tape. **(2) Keep small hands clear of everything but the cord's handle.** Fingers can get pinched. (3) When all shaking has settled down, lower the "seismograph" pens onto fresh yellow Post-it's™ to make records of how large the shaking is on each platform. (4) Pull out the pin by pulling the cord *firmly to the right.* **This is the appropriate part for children to do.** (5) Peel off the seismograms and paste them into your handout on the next page.

II. If you use the shorter pin to make a smaller earthquake, the dog bone on the left usually barks and the one on the right (on the "rocks") usually does not bark. This method does not work as reliably, because the bones are very simple and not very consistent in their response to shaking. Follow the instructions in (I), but put in the **short** (left) pin and turn on both bones. **Hint:** Let the shaking and barking settle down *all the way.* Wait 10 seconds longer, and then pull **gently** *to the left* until the pin pops out. Lastly, **please turn off the bones!**

Paste Your Seismograms Here

⇒ The same shaking goes in at the bottom. ⇐

⇒ Different shaking comes out the top. ⇐

Valley Sediment (red): **Mountain Rocks (blue):**

References

Boore, D. M., W. B. Joyner, and T. E. Fumal, Estimation of response spectra and peak accelerations from western North American earthquakes: an interim report, *U.S. Geol. Surv. Open File Rep.*, **93-509**, 72 pp., 1993.

Ellsworth, W. L., 6. Earthquake History, 1769–1989, in *The San Andreas Fault System, California, U.S. Geol. Surv. Prof. Pap.*, **1515**, ed. R. E. Wallace, 153–187, 1990.

Wald, D. J., Strong Motion and Broadband Teleseismic Analysis of the 1991 Sierra Madre, California, Earthquake, *J. Geophys. Res.*, **97**, 11,033–11,046, 1992.

Wald, D. J, and T. H. Heaton, Spatial and temporal distribution of slip for the 1992 Landers, California, earthquake, *Bull. Seismol. Soc. Am.*, **84**, 668–691, 1994*a*.

Wald, D. J., and T. H. Heaton, A dislocation model of the 1994 Northridge earthquake determined from strong ground motion, *U.S. Geol. Surv. Open File Rep.*, **94-278**, 54 pp., 1994*b*.

Wald, D. J., T. H. Heaton, and D. V. Helmberger, Rupture model of the 1989 Loma Prieta earthquake from the inversion of strong motion and broadband teleseismic data, *Bull. Seism. Soc. Am.*, **81**, 1540–1572, 1991.

Wessel, p., and W. H. F. Smith, Free software helps map and display data, *Eos, Trans. AGU*, **72**, 441, 445–446, 1991.

Woodford, A. O., J. E. Schoellhamer, J. G. Vedder, and R. F. Yerkes, Geology of the Los Angeles Basin (Chapter 2, Part 5), in *Geology of Southern California, California Division of Mines Bulletin 170*, ed. R. H. Jahns, 65–81, 1954.