Passive Seismic Monitoring of Natural and Induced Earthquakes: Case Studies, Future Directions and Socio-Economic Relevance

Marco Bohnhoff, Georg Dresen, William L. Ellsworth, and Hisao Ito

Abstract An important discovery in crustal mechanics has been that the Earth's crust is commonly stressed close to failure, even in tectonically quiet areas. As a result, small natural or man-made perturbations to the local stress field may trigger earthquakes. To understand these processes, Passive Seismic Monitoring (PSM) with seismometer arrays is a widely used technique that has been successfully applied to study seismicity at different magnitude levels ranging from acoustic emissions generated in the laboratory under controlled conditions, to seismicity induced by hydraulic stimulations in geological reservoirs, and up to great earthquakes occurring along plate boundaries. In all these environments the appropriate deployment of seismic sensors, i.e., directly on the rock sample, at the earth's surface or in boreholes close to the seismic sources allows for the detection and location of brittle failure processes at sufficiently low magnitude-detection threshold and with adequate spatial resolution for further analysis. One principal aim is to develop an improved understanding of the physical processes occurring at the seismic source and their relationship to the host geologic environment. In this paper we review selected case studies and future directions of PSM efforts across a wide range of scales and environments. These include induced failure within small rock samples, hydrocarbon reservoirs, and natural seismicity at convergent and transform plate boundaries. Each example represents a milestone with regard to bridging the gap between laboratory-scale experiments under controlled boundary conditions and large-scale field studies. The common motivation for all studies is to refine the understanding of how earthquakes nucleate, how they proceed and how they interact in space and time. This is of special relevance at the larger end of the magnitude scale, i.e., for large devastating earthquakes due to their severe socio-economic impact.

Keywords Earthquakes · Passive Seismic monitoring · Borehole Seismology · Crustal mechanics · Physics of Faulting

Introduction

Global monitoring of seismicity detects the occurrence of earthquakes down to about M = 4. The resulting pattern of their distribution traces the plate boundaries and highlights the most active intraplate seismic zones. In many parts of the globe the detection threshold is lower because of the presence of regional and local seismic networks. Within regions such as the west coast of North America, Japan and Western Europe regional thresholds on the order of M = 1–2 have been achieved. Such well-designed local seismic networks not only record earthquake activity at low magnitude detection thresholds but also resolve the focal depths and focal mechanisms of earthquakes. The tools of modern Passive Seismic Monitoring, referred to as PSM in the following, allows the refinement of earlier

M. Bohnhoff (✉)
Department of Geophysics, Stanford University, Stanford, CA 94305-2215, USA
e-mail: marcob@stanford.edu

Marco Bohnhoff now at Helmholt - Centre Potsdam GFZ (bohnhoff@gfz-potsdam.de)

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seismotectonic models pioneered in the 1970s through the application of modern methods for determining crustal structure, locating earthquakes and determining focal mechanisms. These include 3-D mapping of active faults and fault systems, routine moment tensor determination of source processes, analysis of earthquake interaction, high-resolution characterization of active faults within hydrocarbon and geothermal reservoirs, and investigation of the systematics of the earthquake cycle for large magnitude earthquakes along plate-bounding faults.

The concepts behind PSM date back to the 1930s and 1940s and were accompanied by the quantification of the earthquake phenomenon. The introduction of the earthquake magnitude scale for regional events in California by C.F. Richter (1935) and the discovery of the earthquake frequency-magnitude relation by B. Gutenberg and C.F. Richter (1941) are certainly the most prominent and relevant examples. At about the same time, the idea of developing earthquake studies from regional and local seismic networks was introduced to study aftershock sequences in Japan (Imamura et al., 1932).

Based on technical developments in the 1960s that permitted the low-power operation and recording of many seismic stations on a common time base, new seismic networks and processing methods were developed that permitted the routine analysis of earthquakes of M < 2, commonly referred to as "microseismicity". A comprehensive review of principles and applications of microearthquake networks of this period was given by Lee and Stewart (1981). Using a similar approach extensive research using local arrays of seismic stations was undertaken to implement a nuclear test ban treaty. Over the last two decades, progress in the field of PSM has occurred primarily as a result of advances in seismic instrumentation and computation facilities to store and serve large data sets. The transition to digital high-frequency full waveform data acquisition systems with increased dynamic range also stimulated the development of more sophisticated analysis schemes allowing refinement of existing models at local, regional and plate-boundary scale.

A key objective of modern PSM is to collect data that can be used to resolve earthquake source processes in space and time during the rupture process. Seismic waves observed at a receiver carry information about the source process, but are also modified by propagation through the earth. When earth structure is sufficiently well known, it is possible to correct the observed waveforms for many propagation effects. As the waves propagate in the heterogeneous and inelastic earth, information about the source process is lost due to scattering and attenuation. Once lost, this information cannot be recovered, and so the solution is to place the receiver "close enough" to the source. Because the losses are greatest at the highest frequencies, to record a signal with a ~1 m wavelength the sensor must be within less than 1 km of the source, assuming anelastic and scattering losses corresponding to a damping factor of Q ~500. Clearly, if one intents to understand earthquake processes on a specific scale, one needs to record close to the source.

In this paper we summarize the principal objectives involved in PSM and review selected examples and future directions from key-locations representing various environments such as hydrocarbon and geothermal reservoirs, seismically quiet intra-plate regions, and large-scale transform faults and subduction zones. We also consider rock-deformation experiments in the laboratory thus giving examples that cover rupture length scales from millimeter to hundreds of km. Table 1 gives an overview on the different magnitude ranges and the relevant scales of rupture dimension, displacement, dominant frequency and seismic moment for the different environments discussed in this paper. We highlight that a comprehensive understanding of the physical processes responsible for brittle failure requires investigations that span this large spatial and frequency range. The effort to monitor these processes using adequately designed receiver geometries is important, especially with regard to socio-economic implications as shown in the cases of subduction megathrusts and large earthquakes along plate-bounding transform faults.

Quantifying the Earthquake Process

Earthquakes are the vibratory motion of the earth created by the sudden release of energy within the solid rock mass of the planet. Most earthquakes are caused by slip on faults, and as a consequence the term "earthquake" is commonly used to refer to the earthquake source process rather than the seismic waves it causes. Because the waves travel to great distances through the earth even for small underground disturbances they provide a powerful observational basis for studying the location, strength and fundamental nature of the earth-

Table 1 Overview on different earthquake magnitude ranges and relevant scales for rupture length, displacement, dominant frequency and seismic moment. Length and displacement scales are approximate and appropriate for crustal earthquakes with stress drops of 3 MPa. Note that ranges given may overlap between earthquake class depending on source-receiver distances and type of wave recorded

Magnitude range	Class	Length scale	Displacement scale	Frequency scale	Seismic moment*
8–10	Great	100–1,000 km	4–40 m	0.001–0.1 Hz	1 KAk–1 MAk
6–8	Large	10–100 km	0.4–4 m	0.01–1 Hz	1 Ak–1 KAk
4–6	Moderate	1–10 km	4–40 cm	0.1–10 Hz	1 mAk–1 Ak
2–4	Small	0.1–1 km	4–40 mm	1–100 Hz	1 μAk–1 mAk
0–2	Micro**	10–100 m	0.4–4 mm	10–1,000 Hz	1 nAk–1 μAk
–2–0	Nano	1–10 m	40–400 μm	0.1–10 kHz	1 pAk–1 nAk
–4 to –2	Pico	0.1–1 m	4–40 μm	1–100 kHz	1 fAk–1 pAk
–6 to –4	Femto	1–10 cm	0.4–4 μm	10–1,000 kHz	1 aAk–1 fAk
–8 to –6	Atto	1–10 mm	0.04–0.4 μm	1–100 MHz	1 tAk–1 aAk

* 1 Aki (Ak) is defined as 10^{18} Nm. The unit is named after Keiiti Aki, who pioneered the use of seismic moment in theory and practice. The International Association of Seismology and Physics of the Earth's Interior recommended in 2007 the adoption of the Aki as the standard unit of earthquake size.

** The term "microearthquake" traditionally refers to earthquakes $M < 3$. The earthquake class names used here are a compromise between the SI naming conventions, which would require that a microearthquake had a magnitude between $M = 2$ and $M = 4$, and traditional practice.

quake source. Seismology is the science of the analysis of these waves, and over the past century it has become a deep and sophisticated branch of mathematical physics (e.g., Aki and Richards, 2002). From the waves it is possible, in theory, to extract a detailed description of the earthquake source process in space and time. The study of the earthquake source, however, is of necessity an empirical science, as we have little control over when and where earthquakes occur. Aside from analog experiments performed in the laboratory or earthquakes induced by industrial modification of underground conditions, the seismologist must be prepared at all times to capture the earthquake when it happens.

Modern seismological instruments are designed to record the wide range of frequencies and amplitudes contained in the seismic waves of particular interest. Successful PSM also requires a geographic distribution of instruments that encircle the source. Only by recording waves from a range of azimuths and distances is it possible to accurately determine the location of the earthquake source (hypocenter) and determine its basic source properties (moment tensor, focal mechanism, etc.). For most natural earthquakes recorded by surface stations, this requires a network of instruments with at least one station within a focal depth of each earthquake. When these conditions are met, the initial point of rupture in an earthquake can be determined to a precision of a few hundred meters. Substantially higher precision locations can be obtained using the seismic wave field (Rubin et al., 1999; Waldhauser and Ellsworth, 2000).

Obtaining an accurate geographic description of where earthquakes occur is among the most basic steps toward developing a tectonic understanding of a region. Other physical measures of the complex mechanical event producing the earthquake take many forms, including the dimensions of the faulted region, the direction and amount of slip in both space and time, as well as traditional measures based on the amplitudes of the radiated elastic waves. To relate the characteristics of one event to another, the observed quantities must generally be summarized through the use of either an empirical relation, such as magnitude, or a quantity derived from a physical model, such as seismic moment. Based on recordings from the Southern California Seismic Network that initially consisted of ~10 stations at 100 km spacing, Richter (1935) developed the local magnitude scale as a first approach to quantify the earthquake size in a physical sense on an instrumental basis. He defined

$$M_l = \log_{10} A - \log_{10} A_0(\Delta)$$

where M_l is the event magnitude, A is the maximum amplitude recorded by the Wood-Anderson seismograph, and $\log_{10}A_0$ is the reference term used to account for amplitude attenuation with epicentral