LARGE AND REPEATING SLOW SLIP EVENTS

IN THE IZU-BONIN ARC

FROM SPACE GEODETIC DATA

(伊豆小笠原弧における巨大スロー地震および繰り返し

スロー地震の宇宙測地学的研究)



by

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The Izu-Bonin arc lies along the convergent boundary where the Pacific Plate subducts beneath the Philippine Sea Plate. In the first half of my three-year doctoral course, I focused on the slow deformation on the Izu Islands, and later in the second half, I focused on the slow deformation on the Bonin Islands.

The first half of the study, described in Chapter V, is published as a paper, "Transient crustal movement in the northern Izu–Bonin arc starting in 2004: A large slow slip event or a slow back-arc rifting event?". Horizontal velocities of continuous Global Navigation Satellite System (GNSS) stations on the Izu Islands move eastward by up to ~1 cm/year relative to the stable part of the Philippine Sea Plate suggesting active back-arc rifting behind the northern part of the arc. We confirmed the eastward movement of the Izu Islands explained by Nishimura (2011), and later discussed the sudden accelerated movement in the Izu Islands detected to have occurred in the middle of 2004.

I mainly discussed this acceleration and make further analysis to find out the possible cause of this acceleration. Here I report that such transient eastward acceleration, starting in the middle of 2004, resulted in ~3 cm extra movements in three years. I compare three different mechanisms possibly responsible for this transient movement, i.e. (1) postseismic movement of the 2004 September earthquake sequence off the Kii Peninsula far to the west, (2) a temporary activation of the back-arc rifting to the west dynamically triggered by seismic waves from a nearby earthquake, and (3) a large slow slip event in the Izu-Bonin Trench to the east.

By comparing crustal movements in different regions, the first possibility can be shown unlikely. It is difficult to rule out the second possibility, but current evidences support the third possibility, i.e. a large slow slip event with moment magnitude of \sim 7.5 may have occurred there.

In Chapter VI, I describe the result of my study about the slow deformation of the Bonin Islands. These islands are located to the south of Japan, very close to the northern edge of the Mariana Arc. I focus on the repeating slow slip events (SSEs) revealed by the GNSS data from stations in the Hahajima and Chichijima Islands. Numbers of slow slip events (SSE) have been found in various subduction zones around the world. SSEs were first found in Japan. These studies were followed by reports of similar event in regions including the Cascadia subduction zone, North America. The geodetic observation by continuous GNSS stations operated by Geospatial Information Authority of Japan (GSI) contributed a great deal on finding such events in Japan.

Aside from the GNSS data from GSI (formally known as GEONET), we add the additional data from National Astronomical Observatory of Japan (NAO) as well as the supporting evidence from their Very Long Baseline Interferometry (VLBI) data. Using data from the GNSS data from the Hahajima and Chichijima Islands, we focus our study on the repeating SSEs in the latest decade, reporting that at least 5 SSEs have occurred within 10 years with recurrence intervals of ~2 years. These 5 SSEs have similar characters in the time constants and we modeled the dislocation of the fault patch in these SSEs by using a rectangular fault plane model. We constructed simple fault models assuming the rectangular faults in an elastic half-space to explain the observed displacement vectors. The total moment magnitudes of the SSEs are

estimated as 6.8 - 7.0, assuming the shear modulus of 40 GPa.

Both studies confirm the occurrences of SSEs in the Izu-Bonin Islands. This is expected to contribute to SSE studies as the newly confirmed series of the possible repeating SSEs in Japan in addition to well-known SSEs, e.g. in the Bungo Channel, the Boso Peninsula, and the Iriomote Island (SW Ryukyu). These studies have also revealed that SSEs often play important roles in the plate convergence in the Mariana type subduction zones.

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The purpose of this research is to use the GNSS (Global Navigation Satellite System) measurements to investigate crustal deformation, particularly related to the slow deformation in the Izu-Bonin Arc. The dissertation is composed of several chapters, describing the main idea of using GNSS for geodetic purposes, the study of accelerated slow deformation in the Izu Islands, and the research of possible repeating slow deformation in the Bonin Islands.

For this purpose, time series of daily receiver positions are created over the period of up to 18 years in the Izu Islands (1997 to 2015), and 10 years in the Bonin Islands (2006 to 2016). Following the introduction in Chapter I - IV, the later chapters describe the research and the results in details. Chapter V discusses the accelerated eastward movement in the Izu Islands in 2004 and I infer the cause of this acceleration, which leads to the finding of the evidence for the occurrence of a Mw7.5 SSE (the largest SSE observed in Japan). Chapter VI discusses the repeating SSEs occurring in the Bonin Islands with the recurrence interval of ~2 years recorded over the last 10 years. Chapter VII contains the general conclusions and closing remarks.

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This chapter contains the general introduction of geodesy, space geodesy, and the GNSS and GEONET that are mainly used in this study. This chapter also includes the introduction of seismicity and tectonic of Japan especially in the Izu and Bonin Island Arcs as the study area, including the types of subduction zones categorized based on several points of view.

1. Geodesy

Geodesy is the scientific discipline that deals with the measurement and representation of the shape, the movement and rotation, and the gravitational field of the earth, in a three dimensional time-varying space. Positioning is one of the main concerns in geodesy and it is performed in the global scale on the Earth with time-variable shape. Positioning provides the valuable information on geodynamic phenomena including plate motion, crustal deformation, solid earth tides, and polar motion.

2. Space geodesy and satellite geodesy

By mean of space geodesy, we mainly talk about the aspects of geodesy studied by using natural and/or artificial celestial bodies as observed objects or as observing platforms. Space geodesy is thus defined through the observation techniques, referred to as space geodetic techniques, or methods. Space geodesy evolves rapidly in the second half of the twentieth century. It became possible to deploy and use artificial satellites either to study size and figure of the earth from space or to observe them as targets from the surface of the earth.

Today, space geodetic techniques are the primary tools we study size, figure, and deformation of the earth, and its motion as a finite body in the inertial reference system. Space geodetic techniques thus are the fundamental tools for geodesy, astrometry, and geodynamics. Space geodetic observations contain information about the position and movement of the observed object and the observer.

The use of artificial earth satellites for geodetic purposes is also referred to as satellite geodesy. Satellite geodesy is the surveying discipline using the earth orbiting satellites to obtain the geodetic data. It includes several techniques such as Global Navigation Satellite System (GNSS) as represented by Global Positioning System (GPS), Very Long Baseline Interferometry (VLBI) and Satellite Laser Ranging (SLR). The main goals of satellite geodesy are: 1. determining the figure of the earth, positioning, and navigation., 2. determining the geodynamical phenomena, such as crustal dynamics and polar motion. The satellite geodetic data can be applied to diverse fields such as navigation, hydrography, oceanography and geophysics. There are several techniques (methods) based on the instrument platform used to obtain geodetic data.

a. Earth-to-space method,

In this method, the satellite is observed with ground-based instruments. Laser ranging and GNSS are the example instruments operated based on this

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method. GNSS are dedicated radio positioning services. The GNSS data are mainly used in this study, so further details explanation about these satellite systems will be provided separately in the next part.

b. Space-to-earth method

In this method, the satellite carries an instrument or sensor as part of its payload to observe the earth. InSAR is operated using this method. InSAR, stands for Interferometric Synthetic Aperture Radar, is a radar technique used in geodesy and remote sensing of crustal deformation.

c. Space-to-space method

Satellite-to-satellite tracking works based on this method. In this method, the satellite uses its instruments to track or to be tracked by another satellite. Time-variable gravity field of the Earth is measured in this way using twin satellites, such as GRACE (Gravity Recovery and Climate Experiment).

3. GNSS

GNSS is a constellation of many satellites which is used for navigation and precise geodetic position measurements. It refers to a collection of satellite positioning systems that are operating or planned globally. Really global GNSS systems include GPS (United States), GLONASS (Rusia), Galileo (EU), Compass/Beidou (China), and regional GNSS systems include IRNSS (India's next generation regional system), and QZSS (Japanese quasi-zenith satellite system).

GPS (Global Positioning System) is a specific GNSS developed in United

States of America, formally named NAVSTAR GPS, standing for Navigation Satellite Timing And Ranging Global Positioning System. It was first developed by the Department of Defense in the 1970's and 1980's and originally operated as a mean of global navigation primarily for military purpose. It was designed to provide accurate, real-time, unambiguous range measurements for point positioning, to enable real-time navigation for mobile users, and to serve an unlimited number of users anywhere on the earth's surface. GLONASS is the Russian GNSS. As of 2013, only the NAVSTAR GPS and GLONASS are fully operational.

4. Geodetic measurements

Plate motions were first measured in the 1980s using VLBI (Herring et al., 1986) and SLR (Christodoulidis et al., 1985). In the mid 1990's, SLR and, especially, VLBI continued to play an important part in the realization of global reference frames, but the vast majority of geodetic reference stations were GNSS stations. This technology is relatively inexpensive, lightweight and robust, and as a result there are now thousands of continuous GPS (CGPS) stations worldwide. Dense networks of GNSS stations were deployed in many plate boundary zones by the end of the 1990s.

Points on the Earth's surface change their location due to a variety of mechanisms:

- Large scale plate motion, plate tectonics
- Episodic motion of tectonic origin, especially close to faults
- Periodic displacements due to Earth tides

- Postglacial land uplift due to isostatic adjustment
- Seasonal displacements due to surface loads, e.g. snow
- Various anthropogenic movements

Geodetic observations at many active plate margins reveal relatively steady aseismic motion during the time between major earthquakes. In several arcs, accurate convergence rates have been accurately determined by geodetic measurement using GNSS. Geodetic GNSS measurement data are collected either in field campaigns, in which an area is surveyed for a limited period and later resurveyed to observe the displacement, or by continuous arrays designed to continuously track satellites over longer periods of time. Data from receivers are analyzed to produce time series of station coordinates and other data such as delays in propagating media. Horizontal velocities, mostly due to motion of the earth's tectonic plates are represented on the map by arrows extending from individual sites. These time series are useful for studying wide range of geodynamic phenomena including plate motion, mountain building, earthquake deformation cycle, postglacial rebound, and environmental loading. Station coordinate time series are expressed in a spatial reference frame, which is typically a global, earth-centered, earth-fixed (ECEF) reference frame.

A significant number of stations have eventually recorded coseismic jumps as well as other discontinuities including postseismic transient deformation. Some of the recorded discontinuities are non-tectonics in nature, for example the antenna change or replacement, or monument location change for technical purpose. Geodetic observations of postseismic transient deformation associated with large earthquakes lead to adding of the nonlinear components to the GNSS study. The existence of postseismic transient deformation has been known for many decades (Okada and Nagata, 1953; Kanamori, 1973; Thatcher and Rundle, 1984). However, only since ground motion can be recorded continuously using GNSS especially CGPS, a lot of studies could provide the observational evidence for the precise functional form of its time dependence (e.g Shen et al., 1994; Heki et al., 1997; Perfettini et al., 2010).

The main content of this chapter is the explanation of the subduction zones and their seismicity. This chapter also includes the explanation of regular earthquakes. It also focuses on the detailed explanation about slow earthquakes, the main subject in this study.

1. Interaction of plates

Tectonic plates are massive pieces of the earth's lithosphere that interact with each other along their boundaries. Plate boundaries can be characterized as the place where the plates separate, slide alongside each other, or collide into each other. Each tectonic plate moves over a certain distance through time. Each plates slides in a different direction at a different speed. The boundaries of plates covering the surface of the earth are classified into three types (simulated in Figure 2.1):

(a). Convergent boundaries

These boundaries occur where one plate subducts underneath another plate with density lower than the subducting plate, or collides with another plate in the case that both plates are composed of continental material. They are characterized by the compressive stress. As an oceanic plate subducts underneath another plate at a convergence boundary, such area is called a subduction zone. This will be described further in the next part.

(b). Divergent boundaries

These boundaries occur where new lithosphere (plate) is produced and plates move away from each other at spreading centers. They are characterized by the tensional stress. The space created can be filled with new crustal materials from molten magma at depth. Divergent boundaries can form within continents but will eventually open up and become ocean basins. The boundaries on land will initially produce rifts, leading to producing of rift valleys, and the boundaries under the sea, which is the place of most active divergent plate boundaries, occur between oceanic plates and are often called mid-oceanic ridge.

(c). Transform boundaries

These boundaries occur where one plate laterally slides past another, displacing spreading ridge. They can occur underwater or on land, where crust neither created nor destroyed. They are characterized by the shear stress.



Figure 2.1. Schematic illustration of three plate boundaries.

The movement of plates, which causes stress for both stresses, lead to the formation of faults. In the term of faulting, compressive stress produces reverse faults, tensional stress produces normal faults, and shear stress produces transform (or strike-slip) faults. Transform fault often means large-scale strike-slip faults connecting segments of mid-oceanic ridges in the ocean. Because of the friction, the plates cannot simply slide past each other. When stress in rock exceeds a threshold due to secular build-up on both plates, it will release the energy and cause an earthquake.

1.1. Subduction zones

Plate tectonic theory recognizes that the earth's surface is composed of a mosaic of interacting lithospheric plates, where the lithosphere consists of crust (continental or oceanic) and associated upper mantle, with a typical thickness of ~100 km. Oceanic plates are created at mid oceanic ridges (divergent or accretionary plate boundaries) by seafloor spreading and destroyed at convergent or destructive plate boundaries, the subsurface continuations of which are known as subduction zones.

A subduction zone is a region of the earth's crust where the tectonic plates meet and collide. Figure 2.2 shows a cartoon of a subduction zone. The movement of this plate is decided by their mass. The more buoyant plate, normally continental (but could possibly be the oceanic) plate will force the other plate, an oceanic plate, go down beneath it. Subduction zones are characterized by trenches, lines of volcanoes parallel to the trenches, the mountain building, and deep seismic zones dipping from the trenches landward. As the area where the intensive geodynamics processes occur, strong mechanical deformations and complex geochemical processes in subduction zones are responsible for the diversity of various geological structures observed on the surface. Part of the material of the subducted plate is recycled back to the surface and the remainder is mixed back into the earth's deeper mantle. This process balances the creation of lithosphere that occurs at the mid-ocean ridges system.



Figure 2.2. Subduction zones and types of plate boundaries (Cross section by Josè F. Virgil from This Dynamic Planet – a wall map produced jointly by the U.S. Geological Survey, the Smithsonian Institution, and the U.S. Naval Research Laboratory).

In subduction zone, plate subduction forms a trench and uplift area parallel to the trench. Stress and phase changes in the upper part of the cold descending plate produce large earthquakes in a narrow band called the Wadati-Benioff zone. The plate is heated as it descends, and the resulting release of water leads to melting of the overlying mantle. This melt rises to produce the linear volcanic chains that are one of the most striking features of subduction zones. The surface expression of each subduction system is known as an island (continental) arc system. Arc systems are often divided into:

a. Fore-arc region, between the trench and the volcanic front

b. Volcanic arc, the chain of active volcanoes running parallel to the trench

c. Back-arc region, furthest from the trench

1.2. Comparative subductology

Subduction zones can be classified in two types, on the basis of the nature of the crust in the overriding plate and on the age of the subducting plate. There are several different kinds of such classifications. The first classification yields two broad categories: a. those beneath an oceanic plate, which is known as an "intra-oceanic convergent margin", as in the Mariana or Tonga trenches, and b. those beneath a continental plate, which is known as an "Andean-type convergent margin", along the west coast of South America. The second classification yields two end-member types. It depends on the age of the seafloor being subducted, and the arc-normal stress, either extensional or compressional, in regions behind the volcanic front.

a. Chilean-type subduction zones

In Chilean-type subduction zones subduct young (< 50 million years old), that is, hot and thin lithosphere. Such relatively buoyant lithosphere resists subduction and results in a shallowly dipping seismic zone, shallow trench, great thrust earthquakes, and back-arc folding and thrust

faulting. This buoyant lithosphere is associated with Wadati-Benioff zones that dip gently. They have a lot of seismic activities, and are associated with back-arc compression. This type of subduction zones are characterized by strong coupling between two plates. They typically have thick accretionary prisms.

b. Mariana-type subduction zones

In Mariana-type subduction zones, old (> 100 million years old), dense (also cold and thick) lithosphere subducts. It has greater density than the underlying mantle so it readily sinks to the mantle, characterized by a steeply - dipping Wadati-Benioff zone, deep trench. They have small seismic activity in plate interfaces (absence of great thrust faults), and are often associated with back-arc extension. Old oceanic lithosphere tend to sink vertically in addition to moving down-dip. A consequence of this motion is that the slab also rolls back; that is, the hinge at which it bends moves away from the volcanic arc.

This extension is usually accommodated by the development of a small mid-ocean ridge spreading center just behind, or within, the volcanic arc. These small ocean basins are known as back-arc basins, and their development is thought to be episodic. They begin by splitting apart the weakest part of the arc system, the active volcanic chain. Sea-floor spreading in the basins moves one part of the arc away from the trench, where it eventually becomes extinct and forms a remnant arc. The other part of the volcanic chain remains active and moves with the fore-arc and the trench. Because the seafloor of the Pacific is much older in the western Pacific than in the Eastern Pacific, most Mariana-type subduction zones are often found in the western Pacific and most Chilean-type margins lie along the margins of the Eastern Pacific.



Figure 2.3. Two end-members of comparative subductology, i.e. Chilean and Mariana types. A key factor is the age of subducting lithosphere (Figure modified after Uyeda and Kanamori, 1979)

Comparative studies of different subduction zones are instructive in this regard and leads to the recognition of two basic and contrasted modes controlled by the strength of coupling between down-going and over-riding plates. In this part, I discuss the comparative subductology which compares two type of subduction zone, the Chilean type and the Mariana type. This Mariana type has been long time believed to have no record of large earthquakes. In this study we are trying to see if there is any SSE recurring in a Mariana type subduction zone. The Bonin Islands as our study area are located to the north of the Mariana Arc, one end of member of comparative subductology. The angle of the deep seismic zone under the Izu-Bonin arc is ~45 degrees or more, which is much steeper than the northeast Japan arc (~30 degrees).

2. Fast earthquakes vs slow earthquakes

Earthquakes very often occur in boundaries between plates. Interplate earthquakes are caused by the movement over an area of a part of the plate interface or the seismogenic zone. This zone 'locks' between the earthquakes, such that stress builds up, and it is then released as an earthquake. Above and below seismogenic zone, stress cannot build up and the movement between plates occur relatively smooth through time. In plate boundaries, the crust is stressed by plate movement. When the stress exceeds their strength, the rocks on the fault surface rupture and energy is released. The rupture generally occurs along the fault, which are considered as sources of seismic waves. Therefore, the rapid slip of rocks along a fault results in an earthquake. As a consequence of a gradual stress buildup in a region, stress eventually exceeds some threshold value or critical local strength, greater than that the rock can withstand. Then a rupture starts. The spatial scale of the earthquake rupture ranges over the order of 10^{-1} to 10^5 m in micro- to large earthquakes.

An earthquake is a unified physical process originating from the release of energy accumulated by long-term plate motion, followed by the propagation of seismic waves in underground elastic materials, and surface shaking that may cause significant damages. Seismic energy is released during seismic rupture, which is a mixture of shear fracture and frictional slip along near-planar surfaces (fault planes) in rocks at depth. This definition of an earthquake was established in the early 1960s, coinciding with the emergence of the theory of plate tectonics.



Figure 2.4. The seismograms produced by fast (top) and slow (bottom) earthquakes (Source: Pacific Northwest Seismic Network, www.pnsn.org).

While regular earthquakes are catastrophic events with rupture velocities governed by elastic wave speed, the processes that underlie slow fault slip phenomena, including recent discoveries of tectonic tremors, slow-slip events and low-frequency earthquakes, are less well understood. Regular earthquakes take place rapidly, while slow earthquakes occur on time scales that may range up to months and years. They can have moment magnitudes as large as 7 or more, and may be precursors to larger regular earthquakes. Slow earthquakes, on the other hand, propagate slowly and do not produce high-frequency seismic energy.

Figure 2.4 shows the seismograms of fast (regular) and slow earthquakes. The seismogram of the slow earthquake look different from those produced by regular earthquakes. In the figure, the top seismogram is from the 2001 Nisqually earthquake, western North America. There are well-defined peaks that indicate the arrivals of the seismic waves that shake the ground. The bottom recording shows the ground motion from a slow earthquake, which is often referred to as tectonic (non-volcanic) tremor. Unlike the recording from a regular earthquake, tremor looks like a disorganized seismic wave arrivals without distinct peaks. In addition, the seismic waves from a large, regular earthquake like the Nisqually earthquake are much larger than the seismic waves that make up tremors. That is why people can feel a regular earthquake but not a slow earthquake.

In the latest two decades, an expanding variety of unusual earthquakes have been discovered. Space geodetic observations of surface movements by GNSS enabled the scientist to find the slower type of earthquakes, which occur in the period of minutes to years. Some phenomenon including low-frequency earthquake, very-low-frequency earthquake, slow slip event (SSE), episodic tremor and slip (ETS) have been found in various plate interfaces. The characteristics of slow earthquakes are quite different from those of ordinary earthquakes. The seismic moments are estimated for various slow earthquakes and compared with the duration of events. The seismic moment rate of slow earthquakes is almost constant, between 10^{12} and 10^{13} Nm/s. The difference between slow and fast (ordinary) earthquakes increases with the seismic moment. The difference is also observed in the stress drop associated with these events. Although the stress drop for fast earthquakes is in the range of 1–10 MPa, the stress drop for SSEs larger than M_w 6 is estimated to lie in the range of 0.01–0.1 MPa. Similarly, the scaled energies are about 10^{-5} and 10^{-10} for fast and slow earthquakes, respectively. Scaled energy is a term used to define the ratio between seismic energy and moment. It is also proportional to the radiated energy scaled by the fault area and slip (Kanamori and Rivera, 2006).

Some of these events occurred at the same time and in the same place, suggesting a close relationship and perhaps a common origin. A unifying characteristic of these events is that they have much longer durations than ordinary earthquakes of comparable seismic moments. For that reason, scientist refer to them as slow earthquakes.

2.1. Scaling law of slow earthquakes

Recent developments of study of earthquakes have expanded the knowledge of the physics of earthquakes. Newly discovered slow earthquakes are qualitatively different phenomena from ordinary fast earthquakes and provide independent information on slow deformation. Many numerical simulations have been carried out to model both fast and slow earthquakes, but problems remain, especially with scaling laws. Each of these slow earthquakes has been demonstrated to arise from shear slip, just like regular earthquakes, but with longer characteristic durations and radiating much less seismic energy.

It is important to study various SSEs to understand their behaviors and explain their characteristic of time, relationships with any seismic activities, and common location in where they could occur, in order to understand seismic hazard in subduction zones. Ide et al., 2007 formulated the scaling law classifying various types of slow earthquakes, to give better knowledge of the plate subduction process and their characteristic. It enlightened the fact that the moment released during SSEs appears to be proportional to their duration, which differs from the earthquakes behavior where seismic moment grows as the cube of the duration.



Figure 2.5. Scaling law of slow earthquakes (Ide et al., 2007). Seismic moment is proportional to the characteristic duration of the event for slow earthquakes.

Ide et al. 2007 show that these slow events follow a simple, unified

scaling relationship that clearly differentiates their behavior from that of regular earthquakes. Figure 2.5 shows the relationship between the moment magnitude (seismic moment), characteristic duration for various kinds of slow earthquakes. Their seismic moment is proportional to the characteristic duration, and their moment rate function is more or less constant. This scaling demonstrates that they can be thought of as different manifestations of the same phenomena and that they comprise a new earthquake category. Ide et al. (2007) proposed that this new scaling law unifies a diverse kinds of slow seismic events and may lead to a better understanding of the plate subduction process and large earthquake generation.

2.2. Slow Slip Events

In the latest two decades, the development of space geodetic technique enabled discoveries of various type of unusual earthquakes, including the slower type of earthquakes which occurs in longer duration from normal fast earthquake, lasting from seconds to months. Slow slip events (SSE), in particular, occur when faults slip as in regular earthquake, but the do too slowly so it does not radiate seismic waves. Because of the lack of seismic waves, SSEs are not damaging like regular earthquakes. This also makes them more difficult to detect, since they can only be detected with geodetic instruments such as GNSS stations or tilt meters coupled to the Earth.

The difficulty of detection lead for SSEs to remain undiscovered until the late 1990s, when studies of GNSS data from the Nankai subduction zone in southwest Japan and the Cascadia subduction zone on the Pacific coast of North America revealed periodic changes in GNSS velocities. In these areas, the coupling of tectonic plates and elastic loading cause GNSS monuments to move away from the coast relative to the stable interior of the tectonic plate at a constant rate, with a sudden movements back towards the coast during earthquakes.

SSE was first observed as the periodic change in the GNSS data, revealed the occurrence at the depths of 30-50 km, close to the downdip limit of strongly coupled subduction interfaces (Southwest Japan, Hirose et al., 1999; Cascadia, Dragert et al., 2001). SSE is inferred to occur on conditionally stable portions of the plate interface, in the transition from stick-slip (velocity weakening) behavior to aseismic creep (velocity strengthening) (Dragert et al., 2001). These regions were interpreted as the expression of the brittle-ductile transition zone located at the down-dip limit of the seismogenic zone. Above this zone and up to shallow depths, the interface accumulates slip deficit, which is mostly released in interplate thrust earthquakes. Below it, the plates are freely slipping. More recently, SSEs at the shallower depths were also found to have occurred at least in three subduction zones, e.g. the Boso Peninsula, Japan, by Ozawa et al. (2003) and Sagiya (2004), Hikurangi, New Zealand, by Douglas et al. (2005), McCaffrey et al. (2008), Wallace and Beavan (2010), Nicoya, Costa Rica, by Outerbridge et al. (2010).

These SSE last from days to months and occur along the subduction interface with a mechanism releasing some of the stress accumulated by plate convergence. Several SSEs have been shown to trigger seismicity with magnitudes in the M5 class, i.e SSE in Boso peninsula, Japan (Ozawa et al., 2007) and SSE in Hikurangi subduction zone, New Zealand (Wallace et al., 2012). Current studies also provide the evidence that the M9.0 Tohoku earthquake (and its M7.3 foreshock) were preceded by slow slip (Kato et al., 2012; Ito et al., 2013). SSEs in Nankai Trough are known to be accompanied by abundant tectonic tremor (Hirose and Obara, 2010).

SSEs were discovered in other subduction zones and even in some nonsubduction environments. The increasing number of SSEs observed in several subduction zones has offered the possibility to examine their scaling relations (Ide et al., 2007). These studies have enlightened the fact that the moment released during SSEs appears to be proportional to their duration. This differs from the regular earthquake behavior, i.e. seismic moment grows as the cube of the duration.

2.3. Physical and mathematical approach for slow deformation

Understanding the physics of slow deformation, especially SSEs, and how these events differ from regular, fast, high-frequency earthquakes have been the most challenging part in geoscience. Theoretically, the easiest way to explain the physics of slow deformation is by explaining the overriding and downward plunging process of the plates, where the plates rub each other at the plate interface and this motion is governed by frictional force between their surface.

However, the effect of both physical and chemical environment seems to complicate the mechanism of the slip. The present of minerals and fluid may also affect the motion and friction of these surfaces. Several evidences suggest that the presence of fluids play a role in slow slip mechanism (Vidale et al., 2012). These various conditions leave important conundrums in the physics of slow slip. Several observations and laboratories models have provided some insight on its mechanisms but the fundamental physics of slow fault rupture remain unsolved.

The rate-and-state friction law is designed to explain the behavior during regular earthquake. This law explains how the friction varies with the physical and chemical character of plate interface. Describing the slow slip with the rate-and-state friction law requires the attention on the mechanism on how slips fail to accelerate to regular earthquakes. The stress release should be much lower than regular earthquakes. There are some possible mechanisms on how the slow slip events generated by using the rate-and-state friction law.

One commonly known model is the spring-block slider model. This model suggests a steady state velocity weakening at low slip speeds, but strengthening of the faults at higher speeds. Slow events may start when the available velocity-weakening fault was too long for steady sliding but too short for dynamic instability (Rubin, 2008). Marone (1998) explained the frictional strength ruling the slow slip as:

$$\tau = \bar{\sigma} \left[f^* + a \ln \frac{v}{v^*} + b \ln \frac{v^* \theta}{D_c} \right]$$
 1)

where $\bar{\sigma}$ is the effective normal stress, f^* and V^* are the reference value of the friction coefficient and sliding velocity, V is the slip speed, θ is the state variable, D_c is the characteristic slip distance for state evolution, and a and bare the empirical coefficient of the order of 10⁻². For a > b, the surface is
steady state velocity strengthening and sliding is stable. For a < b, the surface is steady state velocity weakening and unstable sliding is possible.

Marone et al., 1991 explained the mathematical approach on modeling the time evolution of the displacement by in slow deformation. Such temporal change of displacement is modeled with exponential and logarithmic functions, explaining the energy release process and how it decays over time. This mathematical model using exponential and logarithmic functions are described in the following equations:

Exponential model

$$SSE = 1 - \exp\left(\frac{-(t-t_0)}{\tau}\right)$$
 2)

Logarithmic model

$$SSE = \log\left(\frac{(t-t_0)}{\tau} + 1.0\right),\tag{3}$$

where

- t_0 is the onset time of the slow deformation
- t is the time
- τ is the time constant

Chapter III Japan - Tectonic Setting and Seismicity

This chapter contains the introduction of the Japan area. We describe the tectonic setting of the whole Japan and seismic activity there. We also describe how GNSS has been applied in this area to study crustal deformation. The final part will explain the Izu-Bonin-Mariana (IBM) arc system, where most of our study takes place. It will also include the brief explanation about crustal activities there.

1. Tectonic setting of Japan

Japan is a tectonically active region and it is widely acknowledged to be one of the best-studied arc-trench systems in the western Pacific area. Intensive monitoring of seismicity and crustal deformation, combined with studies of active faults, has allowed a detailed picture of tectonic processes and deformation over different timescales to be drawn over recent decades. The Japanese Islands lie at the junction of four major tectonic plates – the Pacific and the Philippine Sea Plates (oceanic plates) and the North American (or Okhotsk) and the Eurasian (or Amurian) Plates (continental plates) (Figure 3.1). The Pacific Plate moves towards the WNW at a rate of about 5 cm/year and subducts beneath the Izu-Bonin (or Izu-Ogasawara) Arc. The Philippine Sea Plate moves towards the NW at a rate of approximately 5 cm/year (Wei and Seno, 1998) and is subducting beneath SW Japan and the Ryukyu Arc. In SW Japan, the volcanic front lies parallel to the Ryukyu Trench and the Nankai Trough.



Figure 3.1. Current tectonic setting of the Japanese Islands (NUMO-TR-04-04, www.numo.or.jp)

The Japanese islands consist of five different arcs: the Kuril, the Northeast Japan, the Izu-Bonin, the Southwest Japan, and the Ryukyu Arcs. The Northeast Japan Arc meets the Southwest Japan Arc in central Honshu, and the Izu-Bonin Arc collides with these two arcs. Each island arc is accompanied with a trench in parallel: the Kuril Arc – the Kuril Trench, the Northeast Japan Arc, the Japan Trench, the Izu-Bonin Arc – the Izu-Bonin Trench, the Southwest Japan Arc – the Nankai Trough, and the Ryukyu Arc – the Ryukyu Trench. These trenches are divided into two series: The line of Kuril, the Japan and the Izu-Bonin Trenches; and the line of Nankai Trough and the Ryukyu Trench. The arc-trench system in Japan, therefore, is classified into two systems: The eastern Japan arc system (the Kuril, the Northeast Japan and the Izu-Bonin Arcs), and the western Japan arc system (the Southwest Japan and the Ryukyu Arcs). Tectonism and volcanism in the eastern Japan arc system and in the western Japan arc system are mainly regulated by the Pacific Plate and the Philippine Sea Plate movements, respectively. The Izu-Bonin Arc (also called the Izu-Ogasawara Arc), ~1100 km long and ~300-400 km wide, collided with the central Honshu at the northern end and connected with the Mariana Arc at the southern end.

2. GNSS stations in Japan

The GNSS Earth Observation Network System (GEONET) is a permanent nationwide GNSS array operated by Geospatial Information Authority of Japan (GSI). GSI operates GNSS-based stations that cover the Japanese Archipelago with over 1300 stations covering the Japanese archipelago at an average separation of ~20 km for crustal monitoring and GNSS surveys in Japan. From GEONET data, daily station estimation of positions revealed coseismic and postseismic displacements for many earthquakes that have occurred since 1994. It has also revealed plate motions and interseismic deformation along the plate boundaries (e.g., Sagiya, 2004). On the routine basis. the GEONET GPS data are processed with the Bernesse 5.0 software to estimate the daily coordinates of the stations, and we used the F-3 solution (Nakagawa et al., 2009).

3. SSE in Japan

In the last decade, SSEs have been identified at many subduction margins worldwide that are well instrumented with GNSS. Studies of SSEs and associated seismic phenomena provide important insights into the mechanics and physical conditions at subduction zone plate interfaces. In particular, SSEs in the Boso Peninsula, Kanto District, Japan (Ozawa et al., 2007a) and in the Hikurangi subduction zone in New Zealand (Wallace et al., 2012), have been shown to trigger seismic activities including M5 class earthquakes. In Japan, the M_w9.0 Tohoku-oki earthquake was preceded by the rapid afterslip of its M7.3 foreshock (Kato et al., 2012; Ito et al., 2013). It is therefore critical to study SSEs, and their relationships with triggered seismicity, in order to understand seismic hazard in subduction zones.

SSEs are inferred to occur on conditionally stable portions of the plate interface, in the transition from stick-slip (velocity weakening) behavior to aseismic creep (velocity strengthening) (Dragert et al., 2001; Larson et al., 2004; Ohta et al., 2004, 2006; Wallace and Beavan, 2010). In the Nankai and Cascadia subduction zones, SSEs are accompanied by abundant tectonic tremors that are concurrent and co-located with migrating geodetically resolved slow slips (Hirose and Obara, 2010; Bartlow et al., 2011). However in other areas, such as the Bungo channel in Japan or the Guerrero seismic gap in Mexico, tremors are offset downdip from the slipping region (Hirose et al., 2010; Kostoglodov et al., 2010). Since the relationships between SSEs and tectonic tremors can vary by location, it is important to study SSEs in as many regions as possible to sample the full range of behaviors.

4. Izu-Bonin arc

The Izu-Bonin-Mariana (IBM) Arc system is located in the western Pacific, extends more than 2800 km in north-south. This intra-oceanic convergent zone is the result of a multistage subduction of the Pacific plate beneath the Philippine Sea Plate. There are more than 20 volcanic islands along the Izu-Bonin-Mariana Arc, as well as many submarine volcanoes. The Pacific Plate subducts into the Izu-Bonin Trench at a rate of ~50 mm/year, and the age of the subducting plate is ~132 Ma. The Izu-Bonin Arc (also called Izu-Ogasawara Arc), 1100 km long and 300-400 km wide, collides with Central Honshu at the northern end and connected with the Mariana Arc at the southern end. The straight volcanic front clearly runs in the center of the island arc, dividing it into the outer arc and the inner arc. The outer arc has the non-volcanic landforms with gentle slopes, and the inner arc has volcanoes and complicated landforms including ridges, seamounts, and basins.

Ogasawara Ridge, located in the southern part of the outer arc is a nonvolcanic ridge 400 km long and 50 to 70 km wide. The Shichito-Iwojima Ridge, situated in the center of the island arc consists of active volcanoes, such as the Izu-Oshima, Miyakejima, and Iwojima volcanoes, along the volcanic front. Some volcanoes emerged to be islands and many others are below sea level. The IzuBonin Trench is an oceanic trench in the western Pacific Ocean, consisting of the Izu Trench (at the north) and the Bonin Trench (at the south, west of the Ogasawara Plateau). It stretches from Japan to the northernmost section of the Mariana Trench. The Izu-Bonin Trench is an extension of the Japan Trench, where the Pacific Plate subducts beneath the Philippine Sea Plate, creating the Izu Islands and the Bonin Islands on the Izu-Bonin-Mariana arc system. This Izu-Bonin arc will be the focus of this study, and I explore the possibility of the occurrence of SSEs there.

In this chapter, we will discuss the methods and procedure to analyze the GNSS data to obtain the time series and describe the mathematical model used in modeling them.

1. Time series

Station position time series are most commonly specified in a global or local Cartesian coordinate system. The most commonly used global reference system is the well-known earth-centered-earth-fixed Cartesian axis system $\{X, Y, Z\}$ whose Z axis roughly coincides with the earth's spin axis. The most common local Cartesian coordinate system are described as $\{E, N, U\}$ axes that are oriented east, north, and up, respectively. Typically the model parameters are computed in global Cartesian coordinates and converted to local Cartesian coordinates. The standard linear model for the trajectory of a GNSS station (within a given reference frame) consists of the time in *x*-axis and the displacement in *y*-axis. This displacement will be shown as a discontinuity indicating the sudden jump from a coseismic step or an increasing displacement indicating the slow slip in some period of time.

1.1. Time constant and onset time

The time constant for the decays for each station and each event were determined by searching over the range of values and choosing the decay time associated with the minimum misfit between the observation and the model. The afterslip and slow earthquakes are mostly explained and modeled with the logarithmic and exponential models, as described in Chapter II. The preferred direction of estimating time constant is that for which the decay is maximum. If there is a-priori knowledge about the same event in the same area, we can simply follow the information about the direction of the displacement, otherwise, the iteration and checking the residual can be done to confirm this direction.

1.2. Plotting GNSS data

To test and better understand our modeling work on the geodetic time series in this research, I will draw figures using some examples from the GNSS data in Japan. The position time series are cleaned (i.e. outliers are removed) and modeled independently in the north, east and up directions. We added some step to remove discontinuities due to the antenna changes and maintenance works of the instruments. The observed displacement is plotted in *x*-*y* time series with the displacement in the *y*-axis and time in *x*-axis. Modeling the displacement is started by plotting the events seen in the raw data plot followed by estimating time constant of each detected transient event.



Figure 4.1. Flowchart describing the procedure to plot the raw GNSS data to obtain the final time series. This time series will be used for the next process of data analysis.

Plotting the data starts with the GNSS raw data processing. This process is aimed at obtaining the clear image of any suspected transient events in the target stations. The flowchart of the procedure is shown in Figure 4.1. Figure 4.2 describe the step-by-step procedures to plot the displacement of a GNSS station as shown in the flowchart. The displacement vectors obtained by analyzing the time series are plotted as arrows extending from individual GNSS stations in the map. The procedure is followed by checking possible causes of transient movements. Causes of these changes include transient movements due to natural events such as afterslip, postseismic viscous relaxation, SSE, or simply replacement or maintenance of the GNSS antenna and other instruments. Figuring out the details of each event, we can continue working on the purpose of our study, which is, in this case, finding the possibility of any SSE.



Figure 4.2. Processing the raw GNSS data, a). Raw data from a GNSS receiver. Some undulations appeared as suspected events and need to be

estimated. The red curve shows the fitting process to model the displacement. Here no appropriate models are estimated, and so the red curve is very inconsistent with the observed data, b). Fitting the data by estimating some parameters for the suspected events. The red curve shows improved fits compared to a). The vertical dashed blue lines indicate the starting times inferred by seeing the plot. In several events, the assumed starting time look inconsistent with the assumed starting time of the event. This onset time need to be carefully tuned by searching for the value bringing the least misfit. c). The time series showing that the starting time of the events in black dots coincide well to the model shown in the red curve, indicating optimization of the onset times of the events. Dashed blue lines indicating the modeled onset times are more consistent with the observed data. d). The time series showing the displacement by optimizing the time constant. It shows clear consistency between the model and the observed data. There, the waveforms of the three transient displacements look very similar, with only small amount of noise. This is considered the final result of the time series, providing good estimations of the onset time and time constant of suspected transient deformation events.

2. Okada's DC3D solution

DC3D is the subroutine package by Okada (1992), to calculate displacement and its space derivative at an arbitrary point on the surface or inside of the semiinfinite medium due to dislocation of a finite rectangular fault. There are several parameters included in the calculation in using the Okada model, i.e. the location and depth of the fault, orientation (dip and strike angle) of the fault, dimension (width and length) of the fault, and the slip length and direction. Each parameter has to be optimized to obtain the best value which leads us to the best rupture modeling of seismic events.

One of the simplest ways to obtain the best values in such calculation is using the grid search method. It helps us to find the best parameters by modeling several values of each parameter and finding the results with the least root-mean-squares (rms) or the results with least errors. The value with less error is considered as the better parameter to model the fault. This fault estimation will be used to calculated the deformation at GNSS receivers using model parameters, and select the best model by checking the consistency between the calculated and the observed displacements.



Figure 4.3. Fault geometry (Okada, 1992).

2.1. Seismic moment (Mo) and moment magnitude (M_w)

Best estimated fault parameters and consistency of observed and calculated displacement is analyzed further by checking the seismic moment to understand the stress release of each event. Seismic moment is a measure of the mechanical energy released in an earthquake based on the area of fault rupture, the average amount of slip, and the rigidity of the rocks. Seismic moment can be obtained by the relation of the fault dimension and dislocation/slip (assuming a rigidity of 40 GPa for Izu-Bonin arc) as explain in the equation:

$$M_o = \mu D A \tag{4}$$

where

 μ is the rigidity

D is the length of displacement

A is the area of the fault that moved

Seismic moment provides estimate of overall size of the seismic source. The unit is Nm (Newton meter), similar dimension to Joule, the unit of energy. So it is also a measure of the mechanical energy released in an earthquake. Moment magnitude (M_w) is the scale derived based on the concept of seismic moment. The seismic moment can be used to derive the moment magnitude (Kanamori, 1978) using the equation:

$$M_w = \frac{\log_{10} M_o - 9.1}{1.5}$$
 5)

This chapter contains the first half of the study. First I confirmed the eastward movement of the Izu islands, and later discussed the sudden accelerated movement of them detected to have occurred in the middle of 2004. I mainly discussed this acceleration and make further analysis to find out the possible cause of this acceleration.

1. Abstract

The Izu-Bonin arc lies along the convergent boundary where the Pacific Plate subducts beneath the Philippine Sea Plate. Horizontal velocities of continuous Global Navigation Satellite System (GNSS) stations on the Izu Islands move eastward by up to ~1 cm/year relative to the stable part of the Philippine Sea Plate (SPH) suggesting active back-arc rifting behind the northern part of the arc. Here I report that such eastward movements transiently accelerated in the middle of 2004 resulting in ~3 cm extra movements in three years. I compare three different mechanisms possibly responsible for this transient movement, i.e. (1) postseismic movement of the 2004 September earthquake sequence off the Kii Peninsula far to the west, (2) a temporary activation of the back-arc rifting to the west dynamically triggered by seismic waves from a nearby earthquake, and (3) a large slow slip event in the Izu-Bonin Trench to the east. By comparing crustal movements in different regions, the first possibility can be shown unlikely. It is difficult to rule

out the second possibility, but current evidences support the third possibility, i.e. a large slow slip event with moment magnitude of ~7.5 may have occurred there.

2. Introduction - Izu Islands

The Pacific (PA) and Philippine Sea (PH) plates are subducting beneath Northeast and Southwest Japan arcs at the Japan Trench and the Nankai Trough, respectively. Southward extension of the Japan Trench is the Izu-Bonin and the Mariana Trenches, where Pacific plate subducts beneath the Philippine Sea plate (Figure 5.1 and 5.2). In the northernmost part of the Izu-Bonin arc, the movement of Pacific plate relative to the Philippine Sea plate is ~50 mm/yr toward N84W (Argus et al., 2011). In convergent plate boundaries, plate interfaces are often locked and move episodically as interplate earthquakes (including afterslips), and slow slip events (SSE). There are no historical M8 class interplate thrust earthquakes known to have occurred in the northern Izu-Bonin arc. It is not well known how the two plates converge there owing to the lack of appropriate geodetic observations.

Back-arc of the northern Izu-Bonin Arc (the Izu Islands) is considered to be in the initial rifting stage (e.g. Tamaki, 1985). In fact, a chain of topography suggesting active E-W rifting with width of ~30 km have been identified to the west of the Izu volcanic arc (Taylor et al., 1991). In the southern Izu-Bonin Arc (beneath the Bonin Islands), such back-arc spreading does not occur. Further to the south, however, mature active back-arc spreading occurs in the Mariana Arc (Figure 5.3.a).



Figure 5.1. Plate boundaries in and around Japan. Red circle and black square (with 1-sigma confidence ellipse) show the PH Euler poles of the NNR-MORVEL (Argus et al., 2010) and from the present study, respectively.

The active back-arc spreading in the Mariana Trough has been directly measured by GNSS as eastward movements of the Mariana Islands relative to the stable part of Philippine Sea plate (SPH) (Kato et al., 2003). Likewise, Nishimura (2011) showed that the GNSS stations in the Izu Islands are moving eastward relative to SPH by 2-9 mm/year, and attributed it to the active back-arc rifting behind the Izu arc. In divergent plate boundaries on land, rifting episodes lasting for years occur and are often followed by post-rifting relaxation (e.g. Heki et al., 1993; Wright et al., 2012). However, behaviors of back-arc spreading/rifting have been poorly known due to the lack of geodetic observations near submarine rift axes.



Figure 5.2. Map of the northern Philippine Sea plate. Observed velocity vectors of three GNSS stations in the stable Philippine Sea plate (SPH), Minami-Daitojima, Okino-Torishima, and Hahajima, are used to define the reference frame fixed to SPH. Red arrows shows the observed velocities and green arrows show velocities calculated using the Euler pole and the rotation rate estimated using these three velocity vectors.

In this chapter, I report that transient eastward crustal movement of the Izu Islands relative to SPH started in middle 2004 and lasted for a few years. I propose several geophysical mechanisms, such as postseismic movement of a large earthquake, temporary activation of back-arc rifting, and an independent silent earthquake, as candidates responsible for the event, and discuss which one best explains the observations.



Figure 5.3. (a). Map of Izu-Bonin-Mariana arc system. Back-arc in Izu Islands is considered to be in the initial rifting stage (e.g. Tamaki, 1985). There is no back-arc spreading behind the Bonin Islands, and further to the south, mature active back-arc spreading occurs in the Mariana Arc. (b) Map of the northern part of the Izu-Bonin Arc (Izu Islands). Here the observed pre-2004 velocities of the four GNSS stations are compared with those calculated using the Euler vector of SPH. Residual (black arrows) show eastward direction suggesting the active back-arc rifting.

3. GNSS data in the Philippine Sea plate

3.1. Secular velocity

First, I confirm the secular eastward movements of the Izu Islands relative to SPH as reported by Nishimura (2011), in three steps, i.e. 1) defining the SPH Euler vector using GNSS stations in the stable part of PH, 2) calculating velocities at GNSS stations in the Izu Islands using the estimated Euler vector, and 3) deriving the movements of the Izu Islands with respect to SPH as the differences between the observed and calculated velocities. For this purpose, I use velocities before the start of the transient movement in the middle of 2004 (referred to as "pre-2004" velocities in this dissertation).

Figure 5.2 shows that the velocities of three stations on SPH, Minami-Daitojima, Okino-Torishima and Hahajima, in the F3 solution (Nakagawa et al., 2009). These velocities can be expressed as the clockwise rotation around the Euler pole at (48.5N 152.6E) of ~0.899 deg/Ma, which is close to the NNR-MORVEL values (46.02N 148.64E, 0.910 deg/Ma) (Argus et al., 2011). Hahajima, Bonin Islands, is located only ~100 km from the trench, but its velocity suggests that the island is fixed to SPH to a large extent in a time scale exceeding 10 years (back-arc rifting does not occur behind the Bonin Islands).

Figure 5.3.b shows that the observed pre-2004 velocities of four stations in the Izu Islands (Aogashima, Hachijojima, Mikurajima and Shikinejima) deviate significantly from calculated vectors. The three southern islands (Aogashima, Hachijojima, Mikurajima) show eastward residual velocities (black arrows) of ~1 cm/year. These islands are on the eastern flank of the rift axis, and their residual velocities would reflect E-W tensile strain coming from the back-arc rifting to the west of these islands (red double line in Figure3a). This is consistent with the earlier work by Nishimura (2011). In Shikinejima, the northernmost of the four islands, the residual velocity has eastward component coming from the back-arc rifting, but it is somewhat smaller (~0.5 cm/yr) than the other three islands. It also has significant southward component (~1.5 cm/yr), which is due to the north-south compression caused by the collision of the northernmost PH with the Honshu Island (Nishimura, 2011).

3.2. Eastward Acceleration of the Izu Islands in 2004



Figure 5.4. Time series of four stations in the Izu Islands (see Figure 5.1 and 5.3 for positions) relative to SPH until 2010 December. Eastward acceleration started in the middle of 2004 (a, b). I also show that the north (c) and up (d) components of Aogashima do not show significant changes in 2004. Hachijojima shows a step in 2002 August associated with a shallow earthquake swarm that started on Aug. 13 and culminated 2-3 days later

(JMA, 2003). In (a), I give the equation used to model the excessive movement *u*. The black curves in (a) and (b) show best-fit models in which I assumed that the transient component started on July 17, 2004. The standard deviations of the individual daily solutions in the six time series are, downward from the top, 3.1 mm, 4.3 mm, 6.0 mm, 6.1 mm, 7.1 mm, and 14.4 mm.

Figure 5.4 shows time series of four GNSS stations in the Izu Islands (see Figure 5.3 for locations). They are expressed in the kinematic reference frame fixed to SPH defined using the pre-2004 velocities of three stations in SPH (Figure 5.2). Because SPH is fixed, these stations show persistent eastward velocity (positive trend) throughout the period. In order to avoid the influence of the large-scale intrusion episode that started in 2000 summer of a dike connecting the Kozushima and Miyakejima (Ozawa et al., 2004), the data only after 2001 were used for the two northern stations (Shikinejima and Mikurajima) (Figure 5.4.a). This intrusion episode, whose duration is indicated as a blue line, did not influence the secular eastward movements of the two southern stations (Hachijojima and Aogashima) (Figure 5.4.b).

Data after the 2011 Tohoku-oki earthquake are not used because large postseismic signals prevail throughout the country. Although a local earthquake swarm in 2002 August at Hachijojima (JMA, 2003) caused a steplike displacement of the Hachijojima GNSS station, its post-rifting transient movement was insignificant. Figure 5.4.a and b clearly show that the eastward movements have accelerated significantly in middle 2004. The excessive movement (departure from the extrapolation of the pre-2004 trend) u is a function of Δt , the time after the onset of the event, and can be expressed as,

$$u = A \log \left(1 + \Delta t / \tau\right) \tag{6}$$

I used 0.05 year for the time constant τ for all stations and components, and estimated *A* for individual stations and components. This time constant τ has been inferred by minimizing the post-fit residuals of the east component time series of Mikurajima. The transient movement decays in a few years, resulting in ~3 cm excessive eastward movements in three years. The selection of τ is not sensitive to the cumulative movements; changing τ to 0.10 year alters the cumulative movement less than 2 mm. The movements lack north and up components (Figure 5.4.c, d) and are nearly perpendicular to the trench and the rift axes. The limited distribution of the GNSS stations (located one-dimensionally along the volcanic arc) makes the geophysical interpretation of the acceleration non-unique. I will discuss this issue in the next section.

4. Geophysical models of the transient crustal movements

4.1. Contemporary seismic events and three hypotheses

Before discussing the mechanism of the transient movement of the Izu Islands, I examine if its start coincides with a certain earthquake. Figure 5.6 compares the post-fit residuals at three stations for various onset times of the transient movement. The observed time series are best explained by an onset time in the middle of 2004, but the time resolution is not better than a month.



Figure 5.5. Epicenters of the three earthquakes in 2004 which might have triggered the acceleration of eastward movement of the Izu Islands. 1). Mw 6.7 interplate earthquake on May 29 (Close to Boso Triple Junction, ~200 km east of Izu), 2. Mw 5.6 interplate earthquake on July 17 (on subducting PA slab surface, at Sagami Trough), and 3). Mw 7.2 - 7.4 - 6.6 of foreshock - mainshock - aftershock of September 5-6 earthquakes sequence off the Kii peninsula (within subducting PH slab, ~250 km west of Izu).

In this time window, three relatively large earthquakes occurred near the Izu Islands (epicenters shown in Figure 5.5). The first is the $M_w 6.7$ interplate thrust event on May 29 close to the Boso triple junction, ~200 km east of the Izu Islands (May 29 earthquake). The second is the $M_w 5.6$ event on July 17, also an interplate earthquake on the subducting PA slab surface, at the Sagami Trough (July 17 earthquake). The last one is the September 5-6 earthquake

sequence composed of the foreshock (M_w 7.2), the main shock (M_w 7.4), and the largest aftershock (M_w 6.6), off the Kii Peninsula. They occurred within the subducting PH slab, ~250 km west of the Izu Islands (September 5-6 earthquakes). Seismic intensities were 2-3 for July 17 and September 5-6 earthquakes, but it was 1 for May 29 earthquake in the Izu Islands.



Figure 5.6. Comparison of the root-mean-squares (RMS) of post-fit residuals of the east component time series (Mikurajima, Hachijojima, and Aogashima) for different onset times of the acceleration, and the three vertical lines show occurrences of three nearby earthquakes in 2004 (epicenters shown in Figure 5.5).

The eastward transient movement could be explained by three scenarios related to this earthquake. The first one is that the eastward movement represents the postseismic movement of September 5 earthquake (Hypothesis A). The movement could be also interpreted as the acceleration of the backarc rifting, or a rifting event, dynamically triggered by July 17 earthquake (Hypothesis B). The eastward movement could also be the afterslip of May 29 earthquake (or an independent SSE) at the Izu-Bonin Trench (Hypothesis C). In Section 4.3, I show that Hypothesis A is unlikely. In Sections 4.4 and 4.5, I will discuss which of the other two hypotheses are more likely from geophysical points of view.

4.2. Start of the transient crustal movements

Figure 5.6 does not have sufficient resolution to identify the exact starting date of the transient movement. Therefore, I further pursue this issue at Mikurajima, Hachijojima and Aogashima stations by using Akaike's Information Criterion (AIC) (Figure 5.7). There I first set up a moving time window of the width of ± 150 days. I then fit the time series within the window in two different ways, i.e. the first case with a simple straight line and the second case with an SSE starting at the center of the window. The decrease of AIC (- Δ AIC) in the second case gives the measure of the significance of the SSE onset at the center of the window. By moving the window in time (time step was set to two days), I expect to get a sharp peak of $-\Delta AIC$ at the most likely start time of the transient movements. This is similar to the method Nishimura et al. (2013) and Nishimura (2014) used to detect small SSEs in Japan, and to the method Heki and Enomoto (2015) detected trend changes in ionospheric total electron content. The only difference from these past studies is that they looked for discontinuities (Nishimura et al., 2013) or bending (Heki and Enomoto, 2015) while we look for the start of the change expressed with equation (5) in this study.



Figure 5.7. 2003-2006 east component time series of GNSS stations at Mikurajima (grey), Hachijojima (red) and Aogashima (blue). The significance of the start of a temporary movement inferred as - Δ AIC, obtained using the moving time window of ±150 days, is shown at the bottom with the same colors as the time series. The occurrence times of the three mid-2004 earthquakes possibly related to the temporary movement of the Izu Islands are marked with gray dashed lines. - Δ AIC show peaks closest to the July 17 earthquake.

- 5. Mechanism of the transient movement of the Izu Islands
 - 5.1. Hypothesis A: Postseismic deformation of the 2004 September earthquake sequence

Generally speaking, two major mechanisms of postseismic crustal movements are the afterslip and the viscous relaxation of the upper mantle. Here I discuss if the 2004 Kii Peninsula earthquake sequence can account for the observed eastward transient movement of the Izu Islands. At first, I note that the onset of the transient movement seems to deviate significantly from September 5-6 (Figure 5.7). This makes Hypothesis A less likely than the other hypotheses. In addition to this time difference, Figure 5.8 and 5.10 provide evidence strong enough to rule out Hypothesis A.



Figure 5.8. Map of the epicenter of the foreshock and the main shock of the 2004 September 5-6 earthquake sequence off Kii Peninsula. Coseismic crustal movements of the foreshock and the main shock are shown with thin gray arrows. Three-year postseismic displacements by the viscoelastic relaxation of the upper mantle with viscosity 10^{19} Pa s (Suito and Ozawa, 2009) are shown with black arrows.

The possibility of afterslip (at a fault patch close to the main shock) can be simply ruled out by comparing the time series of GNSS stations in Izu Islands in Figure 5.4.a, b (these stations show only slow transient movements without significant discontinuous steps) with a point much closer to the epicenter (the location is shown in Figure 5.9 and time series plot shown in Figure 5.10). For example, the Shima GNSS station, Central Japan, showed the largest coseismic step (~5.4 cm, southward) by this earthquake sequence (Figure 5.10.b). However, its three year postseismic movement is only ~3.2 cm (this includes contributions from afterslip and viscous relaxation). This indicates that an afterslip exceeding the main shock in moment release did not occur. If the eastward acceleration of the Izu stations were due to the afterslip of the 2004 earthquake sequence, they should have shown coseismic jumps larger than the three-year cumulative eastward movements. Figure 5.4.a and b shows that this is not the case.

Figure 5.8 gives coseismic crustal movements of the foreshock and the main shock of the 2004 September 5 earthquake off Southeast Kii Peninsula at GEONET stations, calculated using the fault parameters in Hashimoto et al. (2005) and an elastic half space (Okada, 1992). They are below 1 mm at the Izu Islands, which is consistent with the lack of coseismic steps in the time series (Figure 5.4.a-c). Such coseismic displacement was small but clear (0.5/0.8 cm) at the Muroto/Susami2 stations (Figure 5.10.a), ~250/150 km to the west of the epicenters. Observations also show that there were little post-2004 transient movements at Muroto/Susami2 stations, in contrast to the clear transient movements in the Izu Islands (Figure 5.4.a and b).

I consider that the viscous relaxation cannot explain the eastward transient movement of the Izu Islands (Figure 5.4.a, b), either. At first, a numerical calculation does not support this. Suito and Ozawa (2009) showed the three-year postseismic displacement field of this earthquake due to viscous relaxation of the upper mantle with the viscosity of 1×10^{-19} Pa.s. They are plotted as black arrows in Figure 5.8 In the Izu Islands studied here, these vectors are ~1 mm or less (~0.97 mm northward in Mikurajima, and ~1.1 m northwestward in Shikinejima). In contrast, the observed movements (Figure 5.4.a and b) are eastward and up to a few cm there.



Figure 5.9. Map of the epicenter of the September 5 earthquake. Red circles show 3 stations, Muroto, Susami2 and Shima, whose time series are shown in Figure 5.10.

By reducing the assumed upper mantle viscosity, we could increase the movements in the Izu Islands. However, this requires that GNSS stations on the opposite side of the epicenter to behave similarly. The Izu stations are located ~250 km to the "east" of the epicenter. The Muroto station, Shikoku, is located ~250 km to the "west" of the epicenters. Figure 5.10.a shows that

Muroto does not show a measurable westward transient movements after 2004 September. If the eastward movement of the Izu Islands was caused by the viscous relaxation of the 2004 Kii Peninsula earthquake, a similar westward movement should emerge at Muroto. This is obviously not the case.



Figure 5.10. The eastward time series of Muroto and Susami2 stations and the north component of Shima station. The solid and dashed vertical lines in the time series indicate September 5 and July 17, respectively.

Figure 5.10.a also shows that the Susami2 station, located only ~150 km WNW of the epicenters (location shown in Figure 5.9), jumped westward by ~8 mm in the earthquakes but exhibit little postseismic transients. Unless I assume significantly non-uniform upper mantle viscosity, viscoelastic relaxation following the 2004 earthquake sequence would not be able to explain the eastward acceleration of the Izu Islands. I also note that co-existence of the stations with and without transient movements excludes the

reference frame problem, i.e. leakage of the movement(s) of fixed reference point(s) used in the F3 solution, is not responsible for the observed transient movements.

5.2. Hypothesis B: A slow rifting event triggered by a nearby earthquake

This transient movement accelerates the secular eastward velocity caused by the back-arc rifting. Hence, it could be interpreted as a temporary activation of the rifting, or a rifting event. Lack of clear discontinuities in the time series (Figure 5.4.a and b) suggests that the whole sequence proceeded as a slow event. Hence I tentatively call it a slow rifting event (SRE) analogous to SSE. Figure 5.6 suggested that the onset of this movement seems to coincide with July 17 M_w5.6 earthquake at the Sagami Trough. An earthquake could trigger rifting by two mechanisms, i.e. by static and by dynamic stress perturbations. Both May 29 and July 17 earthquakes cause static stress changes that encourage the rifting event. However, the E-W tensile stress increase at the rift axis by the larger event (May 29 earthquake) is only ~1 kPa, and dynamic triggering by the shaking would be more likely. It is known that SSEs in convergent plate boundaries are often dynamically triggered by seismic waves from remote earthquakes (e.g. Itaba and Ando, 2011). The seismic waves from July 17 earthquake (seismic intensity was 2 at Hachijojima and Aogashima, and 3 in Mikurajima (JMA, 2004)) might have dynamically triggered the SRE in the Izu back-arc.

A rifting event in a young continental rift involves both normal faulting and magmatism in the shallower and deeper parts, respectively (e.g. Calais et al., 2008). The Izu back-arc rifting is also a young rift, and both of them may have occurred in this episode. However, strain partitioning between them cannot be constrained with the limited geodetic data in this case (the GNSS stations are only on the east side of the source at similar distances from the rift axis).



Figure 5.11. Distribution of M≥3 earthquakes 2000-2010 in the Izu Islands from Japan Meteorological Agency catalog (Kei Katsumata, pers. comm.) in a horizontal map (a) and cross section (b). There are earthquake swarms north of Mikurajima (A, 2000), south of Aogashima (B, 2001), and beneath Hachijojima (C, 2002) and west of Hachijojima (D, 2002-2003). The cumulative number of earthquakes shown in (c) does not show notable change after the mid-2004 onset of the transient crustal movement in the Izu Islands (vertical solid line). The approximate position of the back-arc rift axis

is shown with double lines.

For example, the eastward movements of 3-4 cm of the Izu Islands can be explained as the response of an elastic half space to the intrusion of a dike extending by 200 km or more along the axis of back-arc rifting. There, if the dike extends from surface to the depth of 10 km, the rifting of 4-20 cm in three years would explain the observed eastward displacements. It is noteworthy that several earthquake swarms preceded the SRE in this region. After the major dike intrusion between the Miyakejima and Kozushima (Ozawa et al., 2004) in 2000 summer, there were swarm activities to the south of Aogashima in middle 2001, west of Hachijojima from 2002 to 2003 (Figure 5.11). However, the seismicity did not increase along the rift axis after the start of the middle 2004 transient movement (Figure 5.11). Therefore, this possible SRE would not have been associated with simultaneous dike intrusions at depth.

Figure 5.12 shows that slow slips along a long normal fault running above the rift axis can also explain the eastward movements. Four fault segments dipping eastward by 30 degrees from surface to 10 km depth, with the slip of 7, 15, 15, 12 cm, from north to south, can reproduce the observed displacements. I emphasize that the model of the deformation source is not unique, i.e. I do not have information to further constrain the source of deformation. For example, if the faults extend down to 20 km, half of the slip would make similar displacement fields.

The causal relationship between these earthquake swarms in 2000-2003

and the possible SRE is unknown. Figure 5.4 shows that the earthquake swarms did not cause significant surface movements in these islands. The only exception is the 2002 August swarm activity in Hachijojima, which caused eastward jump of the GNSS station of a few centimeters. This event was, however, not followed by immediate post-rifting transient movements. On the other hand, lack of such jumps before the onset of the 2004 acceleration suggests that large-scale dike intrusions did not take place immediately before this hypothetical SRE.



Figure 5.12. Excessive movements of Izu islands due to transient crustal movement in July 2004. Green arrows show three-year cumulative excessive

movements of the four stations, Shikinejima, Mikurajima, Hachijojima and Aogashima. Red arrows are the calculated displacements of the four Izu stations by the slips of 4, 10, 12, and 8 cm of the four segments, from north to south, of the normal faults drawn with dashed lines. Contour lines show the depths of the PA slab surface with 20 km interval, and the rift axis is assumed to align between the 140 and 160 km depth contours.

Considering these points, the post-2004 eastward acceleration of the Izu Islands would not be the post-rifting stress diffusion as seen in NE Iceland after the Krafla rifting episode (e.g. Heki et al., 1993). However, these minor intrusion events in 2001-2003 may have localized the strain with magmatic heat and enhanced east-west tensile stress in the overlying lithosphere. This may have made it sensitive to dynamic stress perturbations, and slow normal faulting may have been triggered in the shallow part of the back-arc rift axis in response to the seismic wave of the 2004 July 17 earthquake.

5.3. Hypothesis C: A large SSE to the east

As the third hypothesis, I examine the possibility that the eastward transient movements were caused by an interplate large SSE at the Izu-Bonin Trench. The May 29 $M_w 6.7$ earthquake near the Boso triple junction (Figure 5.8 and 5.10) did not cause coseismic displacements exceeding 2 mm in the Izu Islands, which is consistent with the lack of discontinuities in the time series (Figure 5.4). If its afterslip area have expanded southward covering N-S extent > 200 km, it would let all the four islands move eastward as seen in Figure 5.4.a and b. Such an expansion of the afterslip area may also have
occurred in the down-dip direction, which might have induced the July 17 Sagami Trough earthquake. Generally speaking, SSEs could start without triggering earthquakes (Rubinstein et al., 2009), and the lack of such candidate seismic event would not be a problem in Hypothesis C.



Figure 5.13. Cumulative excessive movements of the GNSS stations due to the transient movements starting in mid-2004. These movements are shown as blue arrows in the map. Green circles in show the three stations whose time series are given in Figure 5.14. Contour lines show the depths of the PA slab surface with 20 km interval, and the SSE of 20 cm dislocation is assumed to have occurred at the PA slab surface with depths 20-60 km (two rectangles drawn with red dashed lines). Thin gray arrows are calculated displacements at grid points by this hypothetical SSE.

Figure 5.13 shows that fault patches on the PA slab surface, with the total length of 260 km (or more), could cause fairly uniform eastward motion of Mikurajima, Hajijojima, and Aogashima. I assumed 100 km width of the fault patch with depth range of 20-60 km. This depth corresponds to those of the SSEs repeating beneath the Bonin Islands further to the south (Arisa and Heki, 2015). I assumed 20 cm slip of the faults in the direction of plate convergence (N84W), resulting in the total $M_w \sim 7.5$. Needless to say, this model is non-unique. For example, even if I halve the fault width, the same surface displacements in the Izu Islands would be realized by doubling the slip. The total M_w keeps robust against such changes. According to the scaling law of slow slips (Ide et al., 2007), an SSE of $M_w7.5$ would occur over a timescale of 1-2 years, which is consistent with the observations (Figure 5.4). By this slip, Aogashima is expected to subside by ~ 8 mm. This is, however, not large enough to be detected because the post-fit residual of the time series is larger than this value (Figure 5.4.d).

Figure 5.14 shows time series of three additional GNSS points (Ooshima in Izu-Oshima, and Tateyama and Maruyama in the southern part of the Boso Peninsula). Because these stations are near (the first) or beyond (the second and the third) the plate boundary (Sagami Trough), I removed the pre-2004 trends from the entire time series in Figure 5.14 to isolate the transient components. They show clear southeastward transient movements starting in the middle of 2004. Their movements are similar to those expected by this possible SSE. The observed vectors exceed the calculated ones in the Boso Peninsula stations, and this suggests that the actual slip was non-uniform and

was larger in the northern part of the fault. Considering that the SRE in Hypothesis B cannot move the two Boso Peninsula stations more than 1 mm, I consider Hypothesis C more likely than Hypothesis B.



Figure 5.14. De-trended time series of southeast (N135E) component of Tateyama and Maruyama, and the south-southeast (N120E) component of Ooshima stations. The solid vertical line in indicates 2004 July 17.

The SSE shown in Figure 5.13 and 5.14 is the largest ever observed in Japan. There are several series of repeating long-term SSEs in various places in Japan, e.g. southwest Ryukyu (average M_w 6.6) (Heki and Kataoka, 2008), Hyuganada (M_w 6.7-6.8) (Yarai and Ozawa, 2013), and the Boso Peninsula (M_w 6.6) (Ozawa et al., 2007). The SSE that started in 2000 off the Tokai district, Central Japan, had been considered the largest SSE in Japan, but its M_w has recently been revised downward from 7.0-7.1 to 6.6 (Ochi and Kato, 2013). The SSE repeating every 6-7 years beneath the Bungo Channel would then be the largest SSE observed in Japan. For example, Ozawa et al. (2013) estimated the fault dimension of 100 km x 100 km and the slip 10-30 cm (M_w 7.0-7.1) for those in 1997, 2003, and 2010.

The SSE in Figure 5.13 and 5.14 is ~4 times as large in seismic moment as these Bungo Channel SSEs. Outside Japan, one of the best-studied repeating SSE would be those in the Cascadia subduction zone, Canada. They have M_w 6.3-6.8 (Szeliga et al., 2008). The SSE shown in Figure 5.6 is comparable only to the M_w 7.5 silent earthquake that occurred in the Guerrero seismic gap, Mexico, in 2001 (Kostoglodov et al., 2003).

The slip accumulation rate of repeating SSE should not exceed the plate convergence rates. The PA-PH plate convergence rate in this region is ~49 mm/yr. If I assume 20 cm slip by an SSE, their average recurrence should not be less than 5 years. The 2004 event is the only transient eastward movement episode in the observed time span (1997-2015), and the percentage of slips accommodated by repeating SSE would not be large in the northernmost Izu-Bonin Arc.

6. Conclusions

Among the three hypotheses given in the previous chapter, I consider Hypothesis C the most likely and Hypothesis A the least likely. Nevertheless, the idea of SRE in Hypothesis B is still attractive considering that Hypothesis C requires the occurrence of a very large (M_w 7.5) SSE and that the onset time (Figure 5.7) favors Hypothesis B. The problem will be finally solved when the ocean floor movement of the next event is monitored with a submarine benchmark located somewhere between the Izu Islands and the Izu-Bonin Trench. The mixture of Hypotheses B and C will also be possible, i.e. the SRE and SSE, to the west and east of the Izu Islands, respectively, may have occurred simultaneously encouraging each other.



Figure 5.15. Schematic illustration of the movement of a GNSS point B relative to point A (distance between A and B) on an island arc in response to episodic occurrences of SSE (a) and SRE (b). The velocities are drawn relative to the reference point A. The black and white arrows are velocities in interseismic (interrifting) times (v_1) and during the events (v_2). The top panels of (a, b) show the trenchward movement of the point B relative to the reference point A (length of AB). Reversal (a) or acceleration (b) occurs during the events. In the case (c), SSE is assumed to recur in a subduction zone where both active back-arc rifting and interplate coupling at the trench co-exist. Both B and C move trenchward during SSE, while the polarity of v_1 (interseismic velocity) depends on the distance from the trench.

Figure 5.15.a depicts hypothetical cycles of SSE in a subduction zone without

back-arc spreading. Secular landward movements of the fore-arc would develop east-west compressional stress within the lithosphere. This accumulation eventually leads to the fault slip at the trench (SSE in this case). Figure 5.15.b shows hypothetical recurrence of SRE in a young back-arc rifting axis. Secular trenchward movements of the fore-arc would develop east-west tensile stress within the lithosphere. This accumulation eventually leads to the failure of the rocks at the axis, and would let an SRE start and release accumulated tensile stress. Hypothesis C of the present case corresponds to Figure 5.15.c, i.e. SSE occurs at the trench in a subduction zone with active back-arc rifting. In this case, interevent velocity at a station close to the rift (Point B in Figure 5.15.c, and the Izu Islands correspond to this) would be trenchward (eastward) and SSE causes further movements in the same direction. However, if there is an island closer to the trench (Point C in Figure 5.15.c), its inter- and co-event movements would be different from Point B. In this chapter, I focus on the repeating slow slip events (SSEs) in the Bonin Islands revealed by the GNSS data from stations in Hahajima and Chichijima Islands. There are five SSEs detected to have occurred in this area within 10 years record, with the recurrence interval of ~2 years, I constructed simple fault models to assume the rectangular faults in an elastic half-space (Okada, 1992) to explain the displacement. The total M_w of the SSE is estimated as 6.8 – 7.0, assuming the rigidity of 40 GPa.

1. Abstract

The Izu-Bonin Arc lies along the convergent boundary where the Pacific Plate subducts beneath the Philippine Sea Plate. I analyzed the movements of four GNSS stations and one Very Long Baseline Interferometry (VLBI) station located in two of the Bonin Islands, i.e. Hahajima and Chichijima, and searched for signatures of SSEs beneath these islands. These islands are only ~100 km away from the trench, and are suitable for detecting SSEs. During 2007-2015, we found five SSEs, and they showed fairly uniform recurrence intervals somewhat shorter than 2 years and similar displacement vectors. Such uniformness would reflect that these fault patches are mechanically isolated from surrounding plate boundaries. I inferred simple fault models responsible for the observed displacements. To overcome the difficulty arising from the limited distribution of

geodetic points, I relied on numbers of external constraints, e.g. fault patches lie on the prescribed plate boundary, cumulative slip accumulation rate does not exceed the plate convergence rate, etc. The estimated M_w of the SSEs ranged from 6.8 to 7.0. The SSEs showed waveforms characterized by gradual start and stop, in contrast to other repeating SSEs showing distinct starts and stops such as those in the Bungo Channel and the Boso Peninsula. The slow onsets of SSEs may suggest that the slow slip started within a small patch and subsequently expanded over a larger area. These results suggest that plate convergences in the Marianatype subduction zones are largely accommodated by repeating SSEs.

2. Introduction

The Izu-Bonin Trench is an oceanic trench in the Western Pacific Ocean, consisting of the Izu Trench (in the north) and the Bonin Trench (in the south, west of the Ogasawara Plateau). It stretches from Japan to the northernmost section of the Mariana Trench. Back-arc of the Izu Islands is considered to be in the initial rifting stage (e.g. Tamaki, 1985) and a chain of topography suggesting active E-W rifting with width of ~30 km has been identified to the west of the Izu volcanic arc (Taylor et al., 1991), but such back-arc spreading does not occur beneath the Bonin Islands, and mature active back-arc spreading occurs further to the south in the Mariana Arc (Figure 6.1b).

The angle of the deep seismic zone under the Izu-Bonin arc is 45 degrees or more, which is much steeper than the northeast Japan arc (about 30 degrees), and the Bonin Islands are located to the north of the Mariana Arc, one end of member of comparative subductology. In this study I try to see if there is any possible slow slip events occurring in Mariana-type subduction zone.

- 3. Space geodetic data in Izu-Bonin arc
 - 3.1. GNSS data in Bonin Island arc

I used the F3 coordinate solution of the GEONET GNSS stations installed on two of the Bonin Islands, Hahajima and Chichijima (Figure 6.2a), to study crustal movements. The coordinate data was found to be very noisy during 1996 – 2006, and the significant improvement around 2006 enabled us to analyze 10 years data (2006-2015). During this period, the old Chichijima station ended operation in 2011, which contribute only 4 years data. Another GNSS station in Chichijima (New Chichijima station) station started observation in 2006, and have been providing data up to now.

Small number of GNSS stations with available data might result in less accurate estimation of fault parameters and affect the final results of SSEs. Hence, we add data from an independent station installed by NAO (National Astronomical Observatory of Japan) in Chichijima. A large difference between coordinate data sets from NAO and GEONE GNSS stations comes from the difference in data processing. The GEONET F3 solution is based on relative positioning, while the coordinates of the NAO station have been estimated by Precise Point Positioning (PPP) technique. For the coordinates of Chichijima and Hahajima (Figure 6.2b), the two stations often show similar short-term (and possibly spurious) "movements", and this would reflect the relative positioning approach in the data processing. On the other hand, NAO station coordinates do not show such systematic coordinate migrations.

There is also a difference in the treatments of coordinate jumps associated with antenna changes. For the GEONET stations, such jump information is provided from the GSI webpage, and we can correct for them. In the time series of both Chichijima and Hahajima stations (Figure 6.2b), such jumps have been already removed. For the NAO data, we estimated jumps at the time of the antenna change together with other parameters related to SSE.

3.2. SSE signatures from GNSS data

Generally speaking, long distance from trench to an island would limit the chance to detect the SSE unless this event has huge size and ruptures an extensive fault. This limited chance was found in detecting SSEs in the Izu Islands (~200 km from the trench), i.e. I found one large SSE (M_w >7.5) to have occurred in this area, causing transient eastward movements starting in the middle of 2004 (see the previous chapter).

The Bonin Islands, located to the south of this area, is only ~100 km from the trench, and offer higher probability to detect smaller SSEs. This is also an ideal place to analyze and study the plate convergence beneath the trench (Figure 6.1b). To see the daily movements of the Bonin Islands, GNSS data were plotted together in both horizontal and vertical components (Figure 6.2b, d, e) and we used the exponential function (Eq. 2) to model these SSEs. In these figures, we removed trends during interseismic (i.e. inter-SSE) periods. We can see several clear episodic eastward movements, possibly SSE

signatures, in all the four GNSS stations. The most significant movement is seen in EW component, and much less in NS component, indicating that such transient displacements in these two islands are almost eastward with small amount of N-S components. Mainly because of the lower precision of the GNSS in the vertical component, no clear motion is seen in the time series. Nevertheless, due to the limited number of GNSS stations, we use the vertical component as well as horizontal components in order to extract as much information as possible from surface displacement data.

To determine the time constants and the onset times of the SSEs, one could refer to a-priori knowledge of similar past SSE examples in the same area. Since there have been no previous studies on SSEs in the Bonin Islands, I had to determine them by grid search. The time constants of each event (an exponential decay function is assumed) was inferred by grid-search over a certain range of values in 0.01 year step, and selected the decay time constants resulting in the minimum misfit between model and observations in the time series of the EW components. We inferred the onset times by the same method, and found that the time constant of 0.1 year gives the best fit for all of the SSEs. The only exception is the 2008 SSE, for which time constant of 0.05 years resulted in a better fit. The cumulative displacements in the north, east and up directions were estimated in this way, which shows how much a certain station moved by an event. Figure 6.2b, c shows the time series of all components showing signatures of repeating SSEs and a few regular earthquakes.

3.3. VLBI data (VERA)

Taking the insufficient amount of data into consideration, I added the additional data from VLBI (Very Long Baseline Interferometry) to confirm the existence of SSE signatures seen in the GNSS time series. VLBI, the space geodetic techniques originating in late 1960s, is an interferometric technique in which radio signals emitted from distant quasars are recorded separately by multiple radio telescopes deployed around the globe. A VLBI facility is extremely bulky and expensive, and only a few stations in a country are used mainly for global studies (measurements of plate motion and earth rotation to contribute to the definition of the terrestrial reference frame). It is operated less frequently than GNSS, typically one observing session in a week or two, and one observing session results in the determination of the three components of a baseline vector, a vector connecting the radio telescopes.

The National Astronomical Observatory of Japan has installed the VERA (VLBI Exploration of Radio Astrometry) VLBI network consisting of four stations: Oshu (Iwate), Ogasawara in Chichijima (Bonin), Ishigaki in the Ishigaki Island, the Ryukyu Islands (Okinawa), and Iriki in Kyushu (Kagoshima) (Figure 6.1a). VERA was constructed to obtain 3-dimensional map of the Milky Way galaxy by measuring the distances and motions of compact radio sources within our galaxy with unprecedentedly high accuracy, using relative VLBI technique and annual parallax measurements. The construction of VERA was completed in 2002, and it is under regular operation since 2003 fall.



Figure 6.1. (a). Map of Japan. Blue circles show four astrometric VLBI (VLBI Exploration of Radio Astrometry, VERA) stations in Japan. The baseline length time series from two baselines are used in this study (marked with red dots inside the blue circles). The red line shows the Iriki-Ogasawara baseline, and the blue line shows the Ogasawara-Ishigaki baseline (the time series are shown in Figure 6.2c, (b) The IBM (Izu-Bonin-Mariana) Arc System. Back-arc of the Izu Islands is considered to be in the initial rifting stage (e.g. Tamaki, 1985), marked with green line. There is no back-arc spreading beneath the Bonin Islands, and further to the south, mature active back-arc spreading occurs in the Mariana Arc (marked with blue line). Green dashed rectangle shows the study area, shown in (c). (c). Map of the study area in the Bonin Arc, with the epicenter of two earthquakes in February 27, 2008 and December 22, 2010. (d) The Bonin Arc which consisting of two

relatively large islands (Hahajima and Chichijima) and small miscellaneous islands. GNSS stations are located in these two islands (written in black).



Figure 6.2. (a). Map of Chichijima and Hahajima of the Bonin Islands with GNSS and VERA stations marked with squares and a circle, respectively. (b). Detrended eastward component from 4 GNSS stations. (c) Time series of baseline lengths of Iriki-Ogasawara (IRK-OGA) and Ogasawara-Ishigaki (OGA-ISG) baselines shown in red and blue, respectively. Change in the north (d) and vertical (e) components of the Hahajima GNSS station. The vertical dashed lines in (b) and (c) are indicating the signatures of SSEs repeating in a relatively deep segment (blue), a shallow SSE triggered by the 2008 February regular earthquake (green) and the 2010 December outer-rise earthquake (black). The horizontal axes in (b) – (e) show the time in year.

Geodetic VLBI observing session conducted by VERA Ogasawara in Chichijima, together with other 2 VERA stations in Iriki (Kagoshima) and Ishigaki Island, would provide additional information on the movement of Chichijima and reinforce the observations of repeating SSE as seen in the GNSS time series. The geodetic VLBI observing sessions are less frequent than GNSS, and this makes it impossible to treat these data in the same way as other GNSS data. Figure 6.2c shows the time series of Chichijima-Ishigaki and Iriki-Chichijima baseline lengths.

3.4. SSE signatures from GNSS and VLBI data

The SSE signatures observed in the time series of GNSS stations (Figure 6.2b) and VLBI data of VERA (Figure 6.2c), can be categorized based on its related event such as earthquakes. In the figure, we marked the occurrence of earthquakes and SSEs with dashed lines of different colors. According to the JMA catalogue, there were two earthquakes with detectable crustal deformation in this area during 2007-2015: 1). M_w 6.6 earthquake on February 27, 2008, followed by a SSE, is marked as the green dashed vertical line (this will be discussed separately), and 2). M_w 7.4 outer-rise earthquake on December 22, 2010 (Figure 6.1c), marked in black dashed vertical line. The 2010 earthquake showed a significant coseismic jump recorded by all four stations, but did not show any afterslip signatures. This earthquake made it difficult to isolate the signal of the 2011 SSE, which started only 2 month after the 2010 December earthquake.



Figure 6.3. (a-e). The cumulative horizontal velocity field of the 5 SSEs from 2007 to 2014. Red arrows are the observed displacements from the GNSS data from GSI and NAO. Error ellipses indicate 95% uncertainties. (f) Average displacement vectors of GNSS stations in the Bonin Islands of the 5 SSEs.

Other slow displacement signatures whose start times are marked with orange dashed lines indicate the repeating SSE. They do not have any triggering earthquakes, and recur fairly regularly. There are five such SSEs: 1). SSE starting around August, 2007, 2). SSE starting around July 2009, 3). SSE starting around February, 2011, 4). SSE starting around October, 2012, and 5). SSE starting around November, 2014. These SSEs have fairly uniform recurrence intervals of ~2 years.

The cumulative displacement vectors estimated from the time series are plotted as the arrows extending from individual GNSS stations in the map (Figure 6.3a-e). From the figure, one can notice that the displacements of five SSEs are very uniform in both size and direction. Hence, we calculated their average, which resembles to individual SSEs very well (Figure 6.3f). I will use these average vectors to infer geometry and slip vectors of the fault.

To see the difference of the 2008 SSE and other repeating SSEs, we compare the displacement vectors observed at the two islands. The EW and NS components of the Hahajima station (time series in Figure 6.2b and map in Figure 6.3a-f) show that the displacements are mostly eastward with small amount of southward deviation. Hahajima station moved ~17.4 mm, ~15.9 mm, ~12.2 mm, ~13.2 mm, and ~13.3 mm due to the SSEs in 2007, 2009, 2011, 2012 and 2014, respectively. The Hahajima station in total has moved approximately ~72 mm due to these 5 SSEs.

Time series and the horizontal displacement vectors of Chichijima (the new GEONET point) station (Figure 6.2b and Figure 6.3a-f, respectively) indicate that SSE moved the Chichijima island mostly eastward with small amount of northward deviation. The Chichijima station has moved ~13.6 mm, ~16.7 mm, ~8.1 mm, ~10.0 mm, and ~11.4 mm, due to the SSE from 2007 to 2014, respectively. The total movement due to the 5 SSEs (9 years) is ~56 mm. Chichijima (the old GEONET point) only provide two clear displacement data for the first two SSEs, and showed eastward displacements with small northward deviation. The SSE in 2007 and 2009 have moved this station by ~12.0 mm and ~15.6 mm, respectively.

Such displacement vectors, northward/southward deflection of the Chichijima/Hahajima stations, are peculiar to the repeating SSEs. In contrast, Chichijima/Hahajima showed southward/northward deflection in the 2008 SSE (discussed later in a separate section). These difference reflect the horizontal position of the fault patches that moved in SSEs.

The NAO Chichijima station also showed displacements similar to other GNSS points in Chichijima. It moved ~10.8 mm, ~14.1 mm, ~8.2 mm, ~9.8 mm, and ~6.4 mm due to the 5 SSEs from 2007 to 2014, respectively. The total movement was ~49 mm. From the time series and horizontal displacement maps, one may notice that the 2011 SSE shows somewhat 'strange' vectors, which is probably due to the influence of the M_w7.4 outer rise earthquake on December 22, 2010. Because the 2011 SSE occurred only two months after this earthquake, inter-parameter correlation made it difficult to separate the displacement vectors in the 2010 and 2011 events.

4. Result

4.1. Fault Estimation

We use the formulation of Okada (1992) which gives explicit solutions for the surface displacement due to a dislocation of a rectangular fault in an isotropic elastic half-space. DC3D is the subroutine package based on Okada (1992). There are several parameters to be estimated. They include the location and depth of the fault, dip and strike angle, size and dimension of the fault, and the slip length and direction. The fault of the repeating SSEs in the Bonin Islands is modeled with two rectangular patches along the prescribed slab surface (as seen in Figure 6.1c and illustrated Figure 6.4a).

All parameters estimation was done by combining grid search and leastsquares method. I estimated the slip vectors by the least square method. In doing this, I fixed other parameters, i.e fault position (latitude, longitude, and depth), orientation (strike, dip), and the size (length, width) of the faults. The values of these fixed parameters were optimized to minimize the post-fit residuals of the three components of the displacement vectors of the GNSS stations. In doing the grid search of fault parameters, I constrained the parameters so that they satisfy external conditions as described in the next section. Estimated fault parameters were used to calculate seismic moments and M_w of individual SSEs.

4.2. External constraints

Before estimating the fault parameters caused by the repeating SSE, we should take account of the situation that GNSS stations in this area are located only in two tiny islands, Chichijima and Hahajima. Among others, the fault size and the fault slip lengths have strong negative correlation. When a larger fault is assumed, the estimated slip will become smaller, and vice versa.

To overcome the limitation in assuming the fault parameters with the limited distribution of GNSS stations, we take several external constraints into account, i.e. 1) fixing the fault to the slab surface, and 2) considering the PH-PA plate convergence rate. The first constraint assumes that SSEs occurred at the plate interface. Because of the first constraint, if we give the latitude and longitude of the fault center, its depth and orientation (strike and dip) are readily given. The slab geometry information is extracted from the model for the Pacific Plate slab (earthquake.usgs.gov/data/slab). This model is widely known as Slab 1.0, which is a three-dimensional compilation of global subduction geometries, separated into regional models for individual subduction zones. These values are plotted in Figure 6.1c with the contour interval of 20 km.

Regarding the second constraint, I took account of the PH-PA plate convergence rate. As mentioned earlier, fault size and slip length are negatively correlated. Too small a fault would force us to assume too long slip vectors. Then the cumulative slip rate would exceed the plate convergence rate (~5 cm/year beneath the Bonin Islands). Hence, I can constrain the minimum fault size. In order to clearly show this, I plot the relationship between the assumed fault size and the estimated slip vectors in Figure 6.4b. There I compare 5 cases assuming the different fault width (with same length of fault of 120 km) and show the amount of slip necessary to explain the observed surface displacements. In this simulation, we use the average displacement vectors, and the fault geometry shown in Figure 6.4a.

In the simulation, I tried 5 cases: 1) Case 1, assuming the narrowest fault with the width 100 km (50 km in each patch), 2) Case 2, assuming 106 km for the fault width (53 km in each patch), 3) Case 3, assuming the width of 115 km for the fault width (57.5 km in each patch), 4) Case 4, assuming 127 km for the width (63.5 km in each patch), and 5) Case 5, assuming the largest fault with the width of 144 km (72 km in each patch). Then, the estimated slips and M_w are summarized in Table 6.1.



Figure 6.4. (a) Vertical cross section of the assumed geometry of the fault responsible for the SSEs repeating beneath the Bonin Islands. The fault is composed of two patches, the deeper Fault 1 and the shallower Fault 2. D_1 and D_2 refer to the depth of the fault measure as the distance between the fault reference points (P_1 and P_2) to the surface. C_1 and C_2 are the center of the fault, and δ_1 and δ_2 refer to the dip angles of the faults. (b) Amount of estimated fault slips assuming 5 different fault sizes. Blue stars show M_w and the red stars show the lengths of the slip vectors. Although the estimated slip length is sensitive to the fault size, M_w remains similar for all the assumed fault sizes.

Table 6.1. Simulating the relation between the fault width and the slip and M_w (rigidity: 40 GPa).

Case	width	width vs slip		slip (mm)		moment	M
	(km)	width	slip	fault 1	fault 2	(1.e19Nm)	IVIW
1	100	smaller	larger	50.28	69.72	2.88	6.91
2	106	Î	ſ	49.03	60.97	2.81	6.90
3	115			47.31	52.68	2.76	6.89
4	127	\Downarrow	Ų	45.05	44.95	2.74	6.89
5	144	larger	smaller	42.14	37.88	2.78	6.90

Cases 1 and 2 with narrow faults (width: 100 and 106 km) give the slip of 6 cm and 5.5 cm in one year. Slip accumulation rate can be shown as the amount of slip in one year by dividing the slip with the average recurrence interval (2 years). Because these rates exceed the plate convergence rate, these cases can be ruled out from the consideration. Cases 3, 4 and 5 with fault widths of 115 km, 127 and 144 km, give the slip of 5 cm, 4.5 cm and 4 cm in one year, respectively. These rates are similar to or less than the plate convergence rate. In all cases, M_w remains similar and seem to have no direct correlation with the variation of the fault size (Figure 6.4b). It is necessary to note that we also include the vertical displacement in obtaining the best model. Because we have limited number of GNSS points, I had to use as many data as possible to fully utilize the observations.

The slip rate does not let us decide which of the Cases 3-5 is the most realistic. Among them, Case 3 resulted in the smallest difference between the modeled and observed surface displacements, and I consider it the best fault geometry to explain the repeating SSE beneath the Bonin Islands (Figure 6.5a).

4.3. Fault estimation for each SSE

Average SSE, assuming the Case 3 fault size, can be explained by the dislocation of ~47 mm at the deeper fault and ~53 mm at the shallower one, both in S82E direction. If we assure 40 GPa for the rigidity, the total moment becomes ~2.76 x 10^{19} Nm, equivalent to M_w 6.89 (Figure 6.5a). The fault patch is centered at 26°40" N, 142°00" E and 142°20" E for deeper and shallower faults, respectively. The fault center depths of the deeper (shallower) fault is 48.8 km (30.0 km). The deeper and shallower patches dip westward by 30.0° and 24.5°, respectively, with the strike angle (clockwise from north) of 175°. An important result is that the estimated slip of the shallower segment was somewhat larger than that of the deeper segment, and such difference could come from the smaller coupling of the plate interface in the deeper part.

The model of average SSE is applied to each SSE indicating that all SSE rupture the same segment. Following the best model in Case 3, we assume the fault length of 120 km and presumed the width of 115 km (which will be adjusted for individual SSEs) and other parameters remain similar to ones in Case 3. The adjustment and results are described as follows: The 2007 and 2009 SSEs are modeled by adjusting the fault width to accommodate the slip estimation. Fault width is increased a little bit from the original 115 km, because the 115 km width results in 75 mm and 68 mm slip per year, and this exceeds the plate convergence rate here. To make the rate below the plate convergence rate, their fault width were changed to be 161 km and 128 km, respectively. These assumptions result in the estimated slip of ~60 mm of the deeper fault and ~40 mm of the shallower one for SSE 2007, and ~50 mm for

both the deeper and the shallower faults for SSE 2009. The 2007 and 2009 SSEs have released the seismic moments of ~ 3.86×10^{19} Nm and ~ 3.07×10^{19} Nm, which correspond to M_w 7.0 and 6.9, respectively.



Figure 6.5. Fault estimation. (a) Average of 5 SSEs in 2007, 2009, 2011, 2012, and 2014. (b) Another SSE triggered by an earthquake on February 27. 2008. Both map show the assumed fault patches that slipped in the SSEs. The rectangles in all panels show the surface projection of the assumed fault patches. Panels in both figures: (1) Horizontal displacement at GNSS stations, red and green arrows are the observed and calculated displacement, respectively, (2) Vertical displacements are marked as red and blue arrows, indicating the observed and calculated displacement, respectively, (3) Modeled horizontal displacement at grid points are marked as thin gray arrows, and (4). Modeled vertical displacement at grid points. The estimated fault displacements are shown as thick black arrows in all panels.



Figure 6.6. (a) N135E time series of the Misho station, Shikoku, showing the SSE signals in the Bungo Channel. (b) N150E time series of the Chiba-Ohara station, Kanto District, showing the SSEs in the Boso Peninsula. (c) East component time series of Hahajima showing the SSE signatures in the Bonin Islands. Green dashed vertical lines show the repeating SSEs and blue dashed vertical lines show the time of the 2011 Tohoku-oki earthquake. This earthquake may resulted in the earlier recurrence of SSE in the Bungo Channel and the Boso peninsula. Red dashed vertical lines show two earthquakes during the studied period. Panels a, b and c all show the time constant of each SSE, indicating variable time contants in the Bungo Channel and relatively uniform time constants in the Bonin SSE.

Due to the disturbance by the signals of the $M_w7.3$ outer rise earthquake on December 22, 2010, which occurred 2 month before the start of this SSE, estimation of the station displacements and the fault parameters for the 2011 SSE was difficult. The best assumption of fault width is 98 km in width, suggesting that this SSE is causing dislocation of ~5 mm at the deeper fault and ~81 mm at the shallower one, with the total seismic moment of 2.02 x 10^{19} Nm, equivalent to M_w 6.8. These estimations have larger uncertainties than other SSEs.

SSF #	fault size (km)		slip ((mm)	moment	M
551 "	length	width	segment 1	segment 2	(1.e19Nm)	IV IW
Average	120	115	47.31	52.68	2.76	6.89
2007	120	161	59.75	40.24	3.86	6.99
2009	120	128	50.37	49.59	3.07	6.92
2011	120	98	4.53	81.39	2.02	6.80
2012	120	112	45.80	47.22	2.50	6.87
2014	120	112	59.94	37.61	2.41	6.85

 Table 6.2. The fault estimation of five SSEs.

Both of the 2012 and 2014 SSEs are assumed to have the width of 112 km, and the slip of ~46 mm at the deeper fault and ~47 mm at the shallower one, with total seismic moment of 2.50×10^{19} Nm (M_w 6.9) for the 2012SSE, and the slip of ~52 mm at the deeper fault and ~38 mm at the shallower one, with total seismic moment of 2.41×10^{19} Nm (M_w 6.9) for the 2014 SSE. Table 6.2 summarizes the results of the estimated fault parameters for individual SSEs.

4.4. SSE following February 27, 2008, earthquake

As we recognized earlier, there is one slow movement showed up in 2008 in the time series (Figure 6.2b), and it is not included in the repeated SSE. This slow movement is found to start following the M_w 6.2 earthquake on February 27, 2008, at the plate boundary between the Bonin Islands and the trench (Figure 6.1c). This SSE is unique in several points, i.e., it started following an earthquake, and the GNSS stations in the Bonin Islands detected no significant coseismic steps but only slow displacement. This SSE is also different from other repeating SSEs in the displacement vectors, i.e. Chichijima moved toward ESE while Hahajima moved toward ENE.

To identify the fault patch that moved in this SSE, we repeated the same procedure as we did for the average displacement vectors of the repeating SSEs. We optimized some parameter including the position (longitude and latitude), depth, fault size (length and width), and rake angle, and estimated the dislocation vector. Assuming fault of 54 km in length and 26 km in width, we could reproduce the observed displacement vectors (Figure 6.5a). The fault center is at depth of 18 km, and dips by 20° westward. The slip in this SSE is ~230 mm with total seismic moment of 1.30×10^{19} Nm, equivalent to M_w6.7. This magnitude is much larger than that of the February 27 earthquake (M_w6.2). Hence, we cannot simply consider this as an afterslip of the February earthquake, i.e. this would be an independent SSE triggered by the February earthquake.

The position and the size of the inferred fault suggests that the fault of this SSE is located at the up-dip part of the repeating SSE fault segment (i.e. these two patches do not overlap with each other). This is the reason why we considered this SSE is not a part of the repeating SSEs in the Bonin Islands. The estimated slip exceeds 20 cm, and is much larger than the average of the repeating SSEs (maximum dislocation of 10 cm in 2 years). We assume that the plate boundary beneath the Bonin Islands release the strain every 1.5-2.0 years in the deeper part (depth 20-40 km). At the same time, the shallower part (depth 10-20 km) release the strain less frequently. The recurrence interval would possibly exceed 10 years because our 10 year time window included only one example.

5. Discussion: Comparing the Bonin SSEs with other SSEs in Japan

I analyze the repeating SSEs occurring in the relatively deeper part of plate interface beneath the Bonin Islands. I also found one SSE in the shallower part. Next, I compare various aspects of these SSEs with other repeating SSEs in Japan, e.g. the SSEs in the Boso Peninsula, the Bungo Channel, Northern Hokkaido (Dohoku), and Iriomote (SW Ryukyu).

5.1. Recurrence interval and time constant

Figures 6.6a and 6.6b show the time series of the Misho station, Shikoku, which records the SSEs in the Bungo Channel, and the Chiba-Ohara station which records the SSEs in the Boso Peninsula, both in the period from 1996 to 2016 (~20 years). The dashed green lines and a blue line show the repeating SSE and the 2011 Tohoku-oki earthquake, respectively. Repeating SSEs in the Bungo Channel recur every ~6 years, and those in the Boso Peninsula recur every 5-6 years. One can see that both SSEs occurred earlier than average intervals after the 2011 Tohoku-oki earthquake (e.g. only ~4 years from the previous event in the Boso Peninsula). Figure 6c shows the time series of Hahajima, vertical dashed green lines show SSEs and the blue line shows the 2011 Tohoku-oki earthquake. It can be clearly seen that the

2011 Tohoku earthquake did not influence the recurrence of SSE in the Bonin Islands.



Figure 6.7. Comparison of waveforms from the time series of various series of repeating SSEs. In (a), I compare the difference in how an SSE stops. SSEs in the upper panel, i.e. Boso Peninsula (2007 SSE observed at the Chibaohara station), Bungo Channel (2003 SSE observed at the Misho station), Northern Hokkaido (Dohoku) (2012 SSE observed at the Horonobe station), tend to stop in a distinct manner. On the other hand, the Bonin Islands SSE time series (Average SSE) shown in the lower panel does not have a distinct stop time, i.e. SSE stops gradually. In (b), I compare how SSEs start. The starting time of the Boso Peninsula SSE, the Bungo Channel SSE, and the Iriomote (SW Ryukyu) SSE, all shown in the upper panel, have discrete start times. On the other hand, those in the lower panel, the average waveform at Hahajima, the Bonin Islands and the 2012 Dohoku SSE, show gradual starts (the vertical

blue lines indicate approximate start times) and initial acceleration phases.

Comparing the 5 SSEs in the Bonin Islands and 4 SSEs in the Bungo Channel (excluding the last one, whose time constant is difficult to infer due to the postseismic deformation of the Tohoku-oki earthquake), the time constants of the SSEs in the latter series are fairly variable, ranging from 0.14 to 0.31 years (Figure 6.6b). Their recurrence intervals have been quite uniform for the first 3 SSEs, but the latest SSE in 2014 started significantly earlier, which is also possibly caused by the 2011 Tohoku-oki earthquake. SSEs in the Boso Peninsula have uniform time constants (they are much smaller than those of the Bonin Islands SSEs) and almost similar recurrence intervals. One SSE in 2014 seems to have a longer time constant than the earlier SSEs, but this is probably an artifact due to the rapid postseismic movement of the 2011 Tohoku-oki earthquake.

This difference in the uniformness of the time constants and recurrence intervals of the SSEs repeating in the Bungo Channel and the Boso Peninsula, and in the Bonin Islands might be related to their interaction with the adjacent area. The Bungo Channel and the Boso Peninsula SSEs have mechanical interaction with adjacent faults in the seismogenic depths. On the other hand, Bonin area is isolated from large-scale seismogenic plate interfaces, and has relatively small interaction with them.

5.2. Waveform of the time series

The waveforms of SSEs in the Boso Peninsula, the Bungo Channel, and Dohoku show clear distinct stops of SSE (upper panel in Figure 6.7a). On the other hand, the waveforms of the Bonin Islands SSEs (lower panel in Figure 6.7a) have gradually increasing starts and gradually decreasing terminations. For the SSEs with discrete stops, exponential functions might better approximate the observations, while those with gradual stops might be better approximated by logarithmic functions (Figure 6.8a).



Figure 6.8. Illustration of the stop and start of slow deformation, a). Illustration of slow deformation with a clear stop (blue curve) (as seen in upper panel of Figure 6.7a), which is better approximated with exponential functions, and the slow deformation a indistinct stop (red curve) (as seen in lower panel in Figure 6.7a), which is better approximated with logarithmic functions. (b) Illustration of starting phases of slow deformation. Upper panel (blue curve) shows the typical afterslip and have clear start of the slow movements. Also some SSEs, as seen in the Boso Peninsula, the Bungo Channel, and Iriomote (SW Ryukyu), have distinct starts as shown with the red curve. Other SSEs, such as those found in the Bonin Islands and Dohoku, start is indistinct and is followed by initial acceleration phase.

Afterslips often start immediately after earthquakes, and some SSEs start also abruptly. In the Bonin Islands, one SSE was preceded by an earthquake (the 2008 SSE), and other repeating SSEs did not have any triggering earthquakes (other repeating SSEs). Due to large noises, it is not clear if the former SSE started abruptly. When a slow deformation starts instantaneously, it can be interpreted as the simultaneous start of the whole fault patch. In case of the repeating SSE in the Bonin Islands and Dohoku 2012 SSE (lower panel in Figure 7b), the slow slip event seem to start slowly and accelerate in time. In these cases, the fault may have started slipping in a small patch and expanded in area with time. This would have caused the indistinct starts and faster displacements afterwards (Figure 6.8b).

6. Conclusions

This study confirms the occurrence of SSEs in the Bonin Islands. Most SSEs started without any triggering earthquake, and recurred with the average interval of ~2 years. Using the displacement data of the GNSS stations in 10 years (2006 - 2016), I estimated the fault parameters of individual SSEs, and found that M_w ranged from 6.8 to 7.0. The fault dimension was inferred to be 120 km in length and 100 - 160 km in width. There was one SSE that followed an interplate earthquake on February 27, 2008. In this SSE, a much shallower fault patch than the repeating SSE seems to have slipped. These observations suggest that the plate interface beneath the Bonin Islands are coupled and the accumulated strain is released by repeating SSEs.

In Chapter 3, I showed that the Bonin Islands are mostly fixed to the stable part of the Philippine Sea Plate as long as we discuss long-term average velocities. This means that inter-SSE stress is completely released by SSEs, and no longterm stress accumulation occur in this plate boundary. This is consistent with the lack of large historical interplate thrust earthquakes in the Bonin Islands.

The SSEs in the Bonin Islands are unique in their uniform recurrence intervals and time constants, and this is considered to reflect the mechanical isolation of this area from seismogenic fault segments. This can be confirmed from their stable recurrence without any disturbances by the 2011 Tohoku-oki earthquake, in contrast to other repeating SSEs in Japan, i.e., those in the Bungo Channel and the Boso Peninsula.

The Bonin Islands SSEs show waveforms characterized by indistinct start and stop of individual events. The initial acceleration phase of the Bonin Islands SSEs could be explained by the temporal growth of the fault from a small patch to a large segment. Further studies and comparison with other SSEs are needed for the better understanding of such differences of SSE time series waveforms. Space geodetic observations of SSEs are often difficult. This could be because only small islands are available for GNSS stations or because these islands are far from plate boundaries. This is the case with the Izu-Bonin Arc, an island chain composed of sparsely distributed small islands. Although the plate subduction process in this region draw interests of many researchers, geodetic observations have been limited there because the islands are often very small and are without any populations. Some islands in the Izu-Bonin Arc are volcanically active and volcanic inflation/deflation signals often cause significant disturbances/noises in the GNSS station position data.

This dissertation has focused on space geodetic measurements of slow deformation in the Izu-Bonin Arc. Using the limited data from not only from GSI but also from NAO, I investigated the signals of the occurrence of SSEs. With 20 years of data from GNSS stations installed in the Izu-Bonin Arc, I could analyze the movement of these islands and confirm the occurrence of number of SSEs.

In Chapter V, I have studied the data from the GEONET F3 solution of GNSS stations on several islands of the Izu Arc. I focus on the islands of Shikinejima, Mikurajima, Hachijojima, and Aogashima. In the Japanese Islands, an M_w7 SSE can be easily observed as seen in the Bungo Channel and the Boso Peninsula. However, the Izu Islands are located ~200 km from the trench, and it is quite a challenge to observe signals of M_w7 SSEs there. 20 years of data (1996-2016) in the Izu Islands lead to the finding of SSE with M_w of 7.5 or more, which seems to have started in the middle of 2004. This finding encourages us to move southward to the Bonin Arc

which is located only ~100 km from the trench. There, I analyzed the space geodetic data 2006-2016 from the Bonin Islands.

In Chapter VI I investigated the displacement of Chichijima and Hahajima, using the GNSS data from GSI and NAO, and found significant signatures of repeating SSEs. These SSEs seems to have repeated every ~2 years and slip occurred over almost the same fault patch. We observed at least 5 SSEs, M_w 6.8-7.0, have occurred by the beginning of 2016. This is much less than M_w ~7.5 SSE that occurred in 2004 in the Izu Islands. These results answer our questions about the possibility of SSEs in the Izu-Bonin Arc, which is a typical low-stress subduction zone. This also satisfies our purpose to find the SSEs in remote islands with only limited number of geodetic instruments. In the future, this study is hopefully complemented by other geodetic instruments such as sea-floor positioning, which will help the researcher to fully understand the characteristic of the plate convergence in the Mariana-type subduction zones.

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