A possible mechanism of M 9 earthquake generation cycles in the area of repeating M 7~8 earthquakes surrounded by aseismic sliding

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We propose a generation mechanism of a giant earthquake of magnitude (M) ~ 9 in subduction zones where only $M = 7 \sim 8$ earthquakes have been identified and the surroundings of the source areas are sliding aseismically. In an M 9 event, both the $M = 7 \sim 8$ source areas and the surrounding area rupture seismically and the coseismic slip amount is one order larger than that of $M = 7 \sim 8$ earthquakes. To reproduce such behavior, we assume that an M 9 earthquake occurrence is the fundamental rupture mode in the subduction zone, and the M 9 source area is modeled as a large fracture energy area whose nucleation size is comparable to the size of the source area. The $M = 7 \sim 8$ asperities are modeled as smaller fracture energy areas whose nucleation size is smaller than the asperity size. Based on these assumptions, we demonstrate a simple numerical simulation of earthquake generation cycles. The results are qualitatively consistent with the characteristics of the 2011 off the Pacific coast of Tohoku Earthquake and a number of phenomena observed prior to this event.

Key words: Earthquake generation cycle, hierarchical asperity model, giant earthquake, aseismic sliding, subduction zone, nucleation size, fracture energy, numerical simulation.

1. Introduction

The 2011 off the Pacific coast of Tohoku Earthquake (the 2011 Tohoku Earthquake) with a magnitude (M) of 9.0 occurred in a subduction zone along the Japan Trench (Earthquake Research Committee, 2011). One of the key questions provoked by this event is: “How could an M 9 earthquake occur in a subduction zone in which only $M = 7 \sim 8$ earthquakes have occurred repeatedly in the past 100 years (e.g., Yamanaka and Kikuchi, 2004)?” It should be noted that the occurrence of M 9 earthquakes cannot be explained by the combined rupture of the $M = 7 \sim 8$ earthquake sources. The slip amount in the M 9 event is one order larger than that in each $M = 7 \sim 8$ event. For example, off Miyagi, the central part of the 2011 Tohoku Earthquake, the slip amount was 1.8 m for the 1978 off Miyagi earthquake (Yamanaka and Kikuchi, 2004) and more than 10 m for the 2011 Tohoku Earthquake (e.g., Iinuma *et al.*, 2011). Although Kanamori *et al.* (2006) found an accumulation of slip deficit (3/4 of the plate convergence) off Miyagi over the past 70 years, we need to know how such a slip deficit can accumulate and how it can be released seismically.

In this letter, we propose a hypothesis to answer the key question posed above, and we demonstrate a conceptual numerical simulation based on our hypothesis. We then discuss the simulation results in comparison with the observations of the 2011 Tohoku Earthquake. Finally, we describe the model predictions, which can be evaluated in light of

future observation.

2. Model and Simulation Method

2.1 Hierarchical asperity model for M 9 earthquakes

We assume that the M 9 earthquake occurrence, which ruptured the entire seismogenic zone, is the fundamental rupture mode in the Japan trench. Furthermore, M 9 earthquakes are assumed to occur intermittently, with recurrence time interval of several hundreds to a thousand years, as indicated by geological data such as tsunami sediments (e.g., Minoura *et al.*, 2001). The key question then becomes: “How can $M = 7 \sim 8$ events occur within the source area of an M 9 event during the long-term seismic cycle of M 9 events?” Based on the concept of a hierarchical asperity model that was applied to the M 3 sequence within an M 5 asperity off Kamaishi along the Japan trench (Hori and Miyazaki, 2010), we speculate that such events could occur as follows.

If the frictional property in the M 9 source area is apparently stable enough, in other words, if the nucleation size is large enough, aseismic sliding occurs during the M 9 earthquake cycle. Such aseismic sliding in an asperity has been discussed in terms of smaller events (e.g., Kato, 2003; Chen and Lapusta, 2009). As a result, the aseismic sliding can load the smaller unstable locked patches within the M 9 source area. If the nucleation size and fracture energy of the patches are much smaller than they are in the rest of the source area, unstable slip occurs on the patches but does not propagate to their outside. This is a mechanism that can account for $M < 9$ events within the source area of an M 9 event. This mechanism is similar to the foreshock model (Matsu'ura *et al.*, 1992). The events on the unstable

patches can occur repeatedly until the accumulated strain energy is sufficient to cause an *M* 9 event in the entire seismogenic zone, as an unstable slip on a patch triggers the rupture of the area including large fracture energy. Note that similar multi-scale heterogeneity in fracture energy is assumed for the 2011 Tohoku Earthquake in Aochi and Ide (2011) although they focus on the dynamic rupture process.

In the following discussion, the unstable patches and the *M* 9 source area, including the unstable patches, are called regular asperities and hyper asperity, respectively. The source area outside the regular asperities is called a conditional asperity, since in this area both aseismic sliding and seismic slip occur, depending on the stress and fault strength conditions.

To demonstrate the above-mentioned phenomena, we set up a numerical model for earthquake cycles based on a quasi-dynamic force equilibrium and the rate- and state-dependent friction law, as reported by Rice (1993). As a friction law, we used a composite law proposed by Kato and Tullis (2001). In the framework of the rate- and state-dependent friction law, the nucleation size and fracture energy are proportional to the state evolution length d_c (Rubin and Ampuero, 2005). Hence, we model the asperity distribution via the heterogeneous distribution of d_c , as in Hillers *et al.* (2006).

We use the friction law as a mathematical model for fault friction. This model describes the relationships among slip velocity, shear stress in the slip direction and fault strength (Nakatani, 2001). The model also addresses the strength evolution of slip weakening and logarithmic healing. From this viewpoint, the state evolution length d_c controls the scale of weakening and healing, and its value should be chosen so as to reproduce the behavior of *M* 9 earthquake cycles. The dominant physical processes involved in weakening, such as thermal pressurization, and in healing, such as pressure solution, are another important dimension of this problem, but these questions lie outside the scope of this letter.

2.2 Model setup

We consider a planar dipping fault in an elastic half space in which the upper edge reaches the free surface. We show the spatial distributions of frictional parameters $A - B$ ($A = a\sigma$, $B = b\sigma$) and d_c in Fig. 1, where a , b and d_c are the frictional parameters and σ is the effective normal stress. Based upon rock friction experiments, $A - B$ primarily depends on temperature and hence depends on depth, and d_c increases where the temperature is high under wet conditions (Blanpied *et al.*, 1998). The area of $A - B < 0$ (velocity weakening in steady state) is the seismogenic zone here. d_c also has a larger value in the shallower part of the fault (Hillers *et al.*, 2006). For the sake of simplicity, A is assumed to be constant on the entire fault. Parameters B and d_c are, hence, assumed to depend on depth. d_c is large enough in conditional asperity to become conditionally stable (e.g., Boatwright and Cocco, 1996).

Thus, the entire seismogenic zone is modeled as one large hyper asperity. Of course, this is an over-simplification, and it results in simple recurrence of the entire seismogenic zone rupture for an *M* 9 event, a virtually homogeneous slip distribution in the strike direction, a positive stress drop

Fig. 1. Spatial distributions of frictional parameters (a) $A - B$ and (b) d_c on the model fault. Crosses indicate the cumulative slip shown in Fig. 2.

Fig. 2. Temporal variation of slip at the points shown in Fig. 1. x and d are strike and depth in Fig. 1. V_{pl} is the plate convergence rate (0.05 m/yr). Arrows indicate the timing of images shown in Fig. 3.

in the entire *M* 9 source area, no $M < 9$ aftershock occurrence in and just around the source area, and so on. To introduce such realistic complexity associated with an *M* 9 event, which is achieved by including heterogeneity in $A - B$ and d_c , will be the next step in our study.

For the sake of simplicity, we set only two regular asperities whose radii are about 50 km in the seismogenic zone, even as there are more similar size asperities and also much smaller asperities in the Tohoku subduction zone. The model is a simple and conceptual one and is not intended for any specific subduction zone. The distribution and size of the regular asperities significantly affect the entire rupture patterns and the recurrence time intervals. However, qualitative characteristics, such as $M < 9$ event occurrence within the *M* 9 source area shown below, can be reproduced in many cases that have a different size and distribution of regular asperities. In such cases, d_c in the regular asperity should be one order or more smaller than it is in the con-

Fig. 3. From (a) to (m) show spatio-temporal variations in the stress and slip velocity on the fault. Blue, white, yellow-green and red colors for the slip velocity correspond to locking, slip with plate convergence rate, aseismic slip (faster than the plate convergence rate) and seismic slip, respectively. Numerals attached to the arrows indicate the time intervals between two successive snapshots. (n) shows coseismic slip distribution and the moment magnitude for EQ1 and EQ5.

ditional asperity. This order difference in d_c is consistent with the idea of scaling in slip weakening distance (e.g., Ide and Aochi, 2005). A detailed description of the model parameters and simulation method is provided in Hori *et al.* (2009).

3. Results and Discussion

3.1 Slip history at selected points

In order to visualize the behavior of the entire earthquake cycle, the slip history at a number of sampling points is shown in Fig. 2. The points are at the same depth as those indicated in Fig. 1. In addition, the distribution of the coseismic slip, whose velocity is higher than 0.01 m/s, is shown for EQ1 and 5 in Fig. 3(n). On the regular asperities, earthquakes occur repeatedly ($x = 200, 400$ in Fig. 2). The slip amount is more than 5 times larger in the entire rupture area (giant earthquakes, EQ0, 5) than in the cases of rupture for each regular asperity (regular earthquakes, EQ1–4). This is consistent with the difference in slip amount for the 1978 off Miyagi earthquake and the 2011 Tohoku Earthquake.

The maximum slip is 53 m for EQ5 and the larger slip can be seen in the shallower part of the fault (Fig. 3(n)). This is consistent with one of the characteristics of the 2011 Tohoku Earthquake, namely the several-tens-of-meters slip along the Japan trench axis (e.g., Maeda *et al.*, 2011). Because the fault plane cut the free surface in our model, there was no loading source shallower than the fault plane. Hence, the slip deficit accumulated more easily than it did in the deeper portion of the seismogenic zone, which was loaded by the aseismic sliding to a deeper extent. In addition,

since the state evolution length d_c is large in the shallower part, slip cannot easily be triggered by $M < 9$ events. Furthermore, when seismic slip occurs in an M 9 event, the free surface enhances the sliding. These factors can explain the observed large slip near the surface in our model.

On the other hand, in the conditional asperity ($x = 100, 300, 500$ in Fig. 2), nearly constant aseismic slip and afterslip appear, as well as large coseismic slip and a locked state, depending upon the stage in the M 9 earthquake cycle. Such slip pattern variation can also be reproduced in the velocity-strengthening area with constant d_c (Kaneko *et al.*, 2010). Our model and their model are end-members of the heterogeneity in $A - B$ and d_c . We believe that the actual subduction plate boundary has the properties between them.

After EQ1 or 2, aseismic sliding appears whose slip velocity is 34~62% of the plate convergence rate. The southern half of the source area of the 2011 Tohoku Earthquake showed aseismic sliding with a velocity that was less than 50% of the convergence rate estimated from small repeating earthquakes between 1993 and 2007 (Uchida *et al.*, 2009). In this area, coseismic slip also occurred in the 2011 Tohoku Earthquake, though the slip amount was relatively small (e.g., Yoshida *et al.*, 2011).

3.2 Space-time evolutions of slip velocity and shear stress

Images showing spatio-temporal variations in slip velocity and shear stress are presented in Fig. 3. The entire area of hyper asperity is locked after an M 9 event. An image of the slip velocity for 47 years following the M 9 event is shown in Fig. 3(a). During the interseismic period, the

locked area shrinks and the aseismic slip area spreads. The shear stress becomes higher along the edge of the locked area.

When the aseismic slip area reaches the bottom of a regular asperity, the rupture begins (Fig. 3(b), EQ1). The rupture propagates within the regular asperity but decelerates beyond its boundaries (Fig. 3(c)). After the event, afterslip occurs around the asperity, especially in its deeper part (Fig. 3(d)). After EQ2, both asperities and their surrounding areas again become locked (Fig. 3(e)). Note that the total locked area is smaller than it was after an *M* 9 event (Fig. 3(a)).

After the aseismic slip reaches the regular asperity again, other regular earthquakes (EQ3, 4) and afterslip occur (Fig. 3(f, g, h)). It should be noted that afterslip spreads more widely in this case than it does following the former regular earthquakes (compare Fig. 3(d, h)). Such large afterslip areas for $M = 6 \sim 7$ events was observed from 2005 to 2010 along the landward side of the *M* 9 source area (Suito *et al.*, 2011). Since the wide afterslip in our model is caused by the higher stress level before an *M* 9 event, the observed afterslip is possibly a year-order precursor of the 2011 Tohoku Earthquake. Furthermore, the wide afterslip area in our model decelerated and again became locked, as shown in Fig. 3(i). In other words, the afterslip does not accelerate to produce the *M* 9 event as a preslip. This result is also consistent with the data.

The locked area shrinks further (compare Fig. 3(e, i)). When aseismic slip reaches the regular asperity for the third time, the stress level around the regular asperity is fairly high (Fig. 3(j)). Thus, the rupture does not decelerate outside the regular asperity and becomes an *M* 9 event (Fig. 3(k), EQ5). As shown in Fig. 3(j), the initiation of the *M* 9 event is triggered by the rupture of a regular asperity, similar to the case of *M* 7 events (Fig. 3(b, f)). Thus, a large preslip is not necessarily observed. Actually, no preslip larger than moment magnitude 7.3 is observed near the hypocenter of the 2011 Tohoku Earthquake (Hirose, 2011).

On the other hand, it is important to examine whether or not the initial rupture of the 2011 Tohoku Earthquake corresponds to an asperity that was previously identified in this area. In backward projection analyses (e.g., Wang and Mori, 2011), an initial rupture was identified in the first 40 seconds northwest of the hypocenter. This suggests that one of the asperities off Miyagi ruptured first.

After the *M* 9 event, the shallower part and the area in and around the regular asperities become locked first, and afterslip occurs in the surrounding area (Fig. 3(l)). Such spatio-temporal variations in the recovery of a locked state in the source area of the 2011 Tohoku Earthquake can be examined by analyzing the small repeating earthquake distribution, and more directly, by analyzing the seafloor deformation, as observed using ocean-bottom pressure gauges. One of the key issues here is whether or not a shorter recovery time was observed in past asperities and near the trench axis. More than a year after the *M* 9 event, the entire seismogenic zone is locked and afterslip continues in the deeper parts (Fig. 3(m)).

Since our simulation deals with quasi-dynamics, we cannot estimate the rupture propagation velocity quantitatively

(Lapusta and Liu, 2009). However, large d_c in the conditional asperity results in a lower weakening rate, and hence in slower rupture propagation than in ordinal events. This is consistent with the slow rupture velocity estimated for the 2011 Tohoku Earthquake (e.g., Wang and Mori, 2011). Although the weakening rate during the coseismic rupture has not yet been estimated, our model predicts a higher and lower weakening rate in the regular and conditional asperities, respectively. This is another validation of our hierarchical asperity model and the friction law we have used here.

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