1 The M_w 7.9 2014 intraplate intermediate-depth
2 Rat Islands earthquake and its relation to
3 regional tectonics
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Abstract:

The 2014 M_w 7.9 Rat Islands intermediate-depth (~ 107 km) earthquake is studied using seismic observations at teleseismic distances. The nodal plane with a steep-dip angle (87°) and a strike of 308° inferred from the Global CMT solution is selected as the causative fault plane because of the consistency with the distribution of the aftershocks relocated using the Joint Hypocenter Determination method. Its slip history is constrained using broadband P and SH waveforms and long-period (3-6 mHz) seismic waves in vertical and transverse components. The rupture process associated with this plane shows a rupture propagating at ~ 2 km/s. The rupture essentially breaks one large slip patch (peak slip amplitude of 3.5 m) located 30 km west of the hypocenter. The size of this slip patch extends from 80 to 140 km depth, which would be the largest depth extent of earthquake slip ever documented. We show that this earthquake breaks the entire cold core of the subducting slab defined by the depth of the 600 - 650°C isotherm. In addition, the lateral extent of the rupture appears to be bounded by the heterogeneities of subducting slab. Our results further indicate that, along this segment of Aleutian Arc, the intermediate-depth earthquakes might contribute to the westward transporting motion of Rat block. Lastly, the slip distribution of this earthquake as well as that of two other well studied $M_w > 7.5$ intermediate-depth earthquakes suggest that the coldest portion of the subducting slab, outlined by the double Wadati-Benioff zone, is seismic during large intermediate-depth earthquakes. Such a result is important for estimating the seismic hazard of future large intermediate-depth earthquakes.

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37 Key words:

38 Intermediate-depth earthquake; Earthquake source observations; Central Aleutian; Subduction
39 zone processes; Seismotectonics.

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Main text:

1. Introduction

Gutenberg and Richter (1949) defined the intermediate-depth earthquakes as those occurring at a depth between 70 to 300 km. Within this depth interval, the number of earthquakes exponentially decreases from the shallow depth to a minimum level near 300 km (e.g., Vassiliou and Hager, 1988). For the majority of the subduction zones, the intermediate-depth seismic activity is often distributed over two distinct planes with low seismic activity in between (double Wadati-Benioff zone (DWBZ); e.g., Hasegawa *et al.*, 1978; Brudzinski *et al.*, 2007). The upper plane of the DWBZ corresponds to the top portion of the descending slab (e.g., Hasegawa *et al.*, 1978). The lower plane of the DWBZ usually lies from ~10 to ~35 km below the upper plane (e.g., Brudzinski *et al.*, 2007). The separation between the two planes increases with the age of the subducting plate, and decreases as depth increases (Hasegawa *et al.*, 1978; Brudzinski *et al.*, 2007).

Seismic failure within the intermediate depth range essentially occurs inside the slab (Isacks *et al.*, 1968), but the mechanisms leading to the occurrence of these earthquakes are not yet clear. Isacks and Molnar (1969) have postulated that the shallowly penetrating slabs should be in downdip tension as a result of negative buoyancy with respect to the asthenosphere. Geodynamic mechanisms such as unbending of the elastic core of the subducted lithosphere (Engdahl and Scholz, 1977) and sagging of the slab under its own weight (Sleep, 1979) were later proposed to explain the observations that the upper plane of the DWBZs is often in downdip

compression, while only the lower plane is in downdip tension.

Although geodynamic models can predict a number of the first-order properties of the intermediate-depth and deep earthquakes using realistic mantle viscosity models (e.g., Vassiliou and Hager, 1988), they are not adequate to explain the gross effects of plate age and the cause of brittle failure under high confining pressure (Kirby, 1995). Since then, two major hypotheses have been advanced to explain the cause of intermediate-depth earthquakes. One is dehydration embrittlement; assisting the brittle fracture with the release of fluids from metamorphic reactions (e.g., Raleigh and Paterson, 1965; Green and Houston, 1995; Kirby *et al.*, 1996; Hacker *et al.*, 2003). The second is a shear-heating mechanism; shearing instability due to highly localized viscous creep in a pre-existing shear-zone (e.g., Ogawa, 1987; Kelemen and Hirth, 2007).

It is clear that relative small intermediate-depth earthquakes concentrate in the two planes of the DWBZ. However, there are not many investigations regarding the relationship between the rupture area of large intermediate-depth earthquakes and the DWBZ. The size of these large earthquakes would suggest a significant extent of the rupture, with the possibility that it breaks in between the double-seismic zone. If it does, it is not clear how this region fails. In fact, the unbending of the subducting slab should create a stress neutral plane in between the DWBZ that should prevent a shear rupture. Similarly, the relatively cold temperature of the slab inside this region might not be favorable for the rupture to occur.

Previous studies already indicate that the Gutenberg-Richter law is held for intermediate-depth earthquakes at a global scale (e.g., Green and Houston, 1995). Thus, the largest earthquakes release the majority of the global cumulative seismic moment at this depth range. From January 1, 1976 to January 1, 2015, the Global Centroid Moment Tensor catalog (GCMT; Dziewonski *et al.*, 1981; Ekström *et al.*, 2012) reports 160 earthquakes with $M_w > 7.5$. Fifteen of

95 them occurred in the intermediate-depth range. Previous examples suggest that such large
96 intermediate-depth earthquakes are capable to produce large ground accelerations and are
97 sometimes life-threatening (e.g., Ye *et al.*, 2014). The July 24, 2008 M_w 6.8 northern Iwate,
98 Japan, earthquake, with a centroid depth of 125 km, has produced ground acceleration up to 1.2g
99 (National Research Institute for Earth Science and Disaster Prevention, NIED), which is high
100 even for a shallow earthquake of similar magnitude. The 1977 M_w 7.5 Vrancea, Romania
101 earthquake, with a centroid of 83.6 km, has led to a fatality loss of 1,578 fatalities and an
102 economic loss of US\$2 billion (Georgescu and Pomonis, 2008). Hence, a better understanding of
103 large intermediate-depth earthquakes is not solely of scientific interest.

104 The 2014 M_w 7.9 Rat Islands earthquake, which occurred below the central segment of the
105 Aleutian Arc on June 23rd, 2014 at 20:53:09 UTC, is the largest intermediate-depth earthquake
106 recorded by modern instrument (Ye *et al.*, 2014). It is then important to study such a well-
107 recorded intermediate-depth earthquake in order to better understand the rupture process that
108 may possibly show the mechanisms controlling those large intermediate-depth earthquakes.

109 The Alaska-Aleutian subduction zone (Figure 1), where the Pacific plate subducts
110 beneath the North American plate at a rate of ~ 8 cm/yr (DeMets *et al.*, 2010), extends over 3800
111 km from the Gulf of Alaska to the Kamchatka peninsula (Ruppert *et al.*, 2007). All along the
112 Aleutian island chain, Lallemand and Oldow (2000) have evidenced an active partitioning of the
113 sense of displacement. As a result the intra-arc region is divided into clockwise-rotated and
114 westward translating blocks, deduced from the observation of subsiding summit basins along
115 with strike-slip faulting (e.g., Geist *et al.*, 1988; Krutikov *et al.*, 2008). Despite being a major
116 subduction zone the region has not experienced $M_w \geq 8.0$ earthquakes since the 1938-1965
117 earthquake sequence, which ruptured almost the entire arc (Johnson *et al.*, 1994; Ruppert *et al.*,

2007). The June 2014, M_w 7.9, Rat Islands intermediate-depth earthquake is one of the two largest earthquakes in this region since then. It occurs near the transition between the eastern end of the M_w 8.7 1965 Rat Islands earthquake (e.g., Johnson and Satake, 1996), and the western end of the M_w 8.6 1957 Great Aleutian earthquake (Johnson *et al.*, 1994; Figure 1).

Ye *et al.* (2014) have studied the 2014 M_w 7.9 Rat Islands intermediate-depth earthquake using teleseismic body waves. They concluded that neither the seismic waveform data nor the aftershocks from Preliminary Determination of Earthquake National Earthquake Information Center (PDE-NEIC; USGS) can distinguish the causative fault plane from the two nodal planes. In this paper, the rupture process of the 2014 M_w 7.9 intermediate-depth earthquake is studied using seismic observations at teleseismic distances. Then, we relocate the first three months of aftershocks in order to distinguish the causative fault plane from the two nodal planes. Finally, we investigate the relationship between the 2014 Rat Islands earthquake, the regional tectonic context of the Central Aleutian Arc, as well as the impact of the results on our understanding of large intermediate-depth earthquakes.

2. Finite-Fault inversion

The 2014 Rat Islands earthquake occurred on June 23rd, 2014 at 20:53:09 UTC. Its PDE epicenter (51.83°N ; 178.75°E) is ~ 60 km southwest of the Semisopochnoi Island, Alaska (Figure 1). The GCMT solution for this earthquake features a nearly pure double couple focal mechanism (93%; Figure 1). Its first nodal plane is a steeply-dipping (84°) fault plane, with a strike of 309° and a slip direction of -117° . Meanwhile, the second nodal plane has a shallow-dipping (27°) fault plane, with a strike of 207° and a slip direction of -13° , dominated by strike-slip motion. In order to determine the causative fault plane, we perform a series of finite-fault

inversions using teleseismic body waves and long-period seismic waves (see Section S1 of the Supplementary Material for details).

Three groups of inversions (labeled A to C), each using a different dataset, have been done. For each group, we perform one inversion with the steeply-dipping nodal plane (NP1), and one inversion with the shallow-dipping nodal plane (NP2). For inversion group A, we use only teleseismic ($30^\circ < \Delta < 90^\circ$) P-waves; the inverted rupture process is used to forward model the teleseismic SH-waves and the long-period seismic waves. For inversion group B, we use teleseismic P-waves and SH-waves, and the inverted rupture process is used to forward model the long-period seismic waves. For inversion group C, we use a combined dataset including teleseismic P- and SH-waves and long-period (3-6 mHz) seismic waves in vertical and transverse components, similar to what has been routinely done for the studies of large shallow earthquakes (e.g., Shao *et al.*, 2011). In all the inversions, we let the rupture initiate at the PDE-NEIC hypocenter and propagate either along the steeply-dipping fault plane (NP1) or the shallow-dipping fault plane (NP2). The steeply-dipping fault plane is 178.5 km along strike and 112.5 km along dip, divided into 315 rectangular sub-faults (8.5 km by 7.5 km). The shallow-dipping fault plane is 112.5 km along strike and 178.5 km along dip, divided into 315 rectangular sub-faults (7.5 km by 8.5 km).

Table 1 summarizes the misfit values of all the inversions. It shows that for each inversion group, the misfit obtained using NP1 as the causative fault plane is similar to that obtained using NP2. All the inversion groups tend to slightly favor the shallow dipping fault plane. However, none of these differences is significant enough to make a clear distinction. In fact, when long period seismic are combined with P and SH waves (Group C), the difference in misfit between the two fault planes becomes negligible. In Figure 2 we compare the inverted slip

distributions and rupture initiation contours for all the eight solutions. We can see that when only P-waves are used, the obtained slip distribution is rather patchy essentially because of a lack of depth resolution (Group A). The addition of SH waves renders the large slip patches more focussed (Group B). When the long-period seismic waves are added, the peak slip amplitude is reduced in comparison to the inversions of Group B. We also observe that the rupture models from Group C exhibit more slip at shallow depths. Thus, the addition of long-period seismic waves introduce a leaking of the slip from the main asperities towards shallow depths. We note that the overall rupture process is similar between the models from Group A, Group B, and those from Group C.

We now discuss the rupture process obtained for each of the two nodal planes. Each of the three inversion groups produce similar rupture process on a given fault plane. Thus, we restrict our discussion to the rupture models obtained from inversion group C (i.e., the one that uses both broadband body waves and long-period seismic waves).

2.1 Nodal Plane 1: The steeply-dipping fault plane

When we use the steeply-dipping fault plane (top row of Figure 2), the inverted slip distribution features essentially one slip patch with a peak slip amplitude of ~ 3.7 m, centered about 40 km northwest of the hypocenter. In addition, we observe two secondary patches of < 1 m slip amplitude; one at the hypocenter and one ~ 30 km southeast of the hypocenter. The average rupture speed is the sum at all points on the fault of the rupture time divided by the hypocentral distance and weighted by the slip. By doing so, we get an average rupture velocity of ~ 2.4 km/s. We observe that the rupture essentially propagates bilaterally breaking the major slip patch between 10 to 30 s. It subsequently breaks the southeastern slip second patch between 20 to 45 s. The centroid of the final slip distribution is located 30 km west of the

hypocenter and at a depth of 107.5 km. The seismic moment (M_0) is 1.0×10^{21} Nm ($M_w 7.9$). Figure 2 shows the moment rate function calculated for this model. It shows that ~90% of the moment is released within the first 40 s of the rupture, with ~70% of the total moment being released between 20 to 40 s. The number of sub-faults with slip >10% of the peak slip is 173. The average slip and rise time over these sub-faults are 1.3 m and 3.8 s, respectively. The mean slip rate, defined simply as the ratio of slip and rise time, is about 0.5 m/s, locally reaching 3.2 m/s. We have estimated the static stress drop to be about 3.8 to 6.2 MPa (see Section S2 in the Supplementary Material for a note on the stress drop estimation).

2.2 Nodal Plane 2: The shallow-dipping fault plane

For the shallow-dipping fault plane (bottom row of Figure 2), the source process propagates radially from the hypocenter at ~2.3 km/s. Most of the seismic moment is released from the rupture of one patch with large slip amplitude (~2.8 m) located about 30 km below the hypocenter. The other slipping regions, which overall surround the hypocenter, have lower slip amplitudes (<1.5 m). The centroid of the final slip distribution is located about 11 km east of the hypocenter at a depth of 115 km. The seismic moment (M_0) is 9.8×10^{20} Nm ($M_w 7.9$). Similar to the steeply-dipping fault plane, ~90% of the moment is released within the first 45 s of rupture, and ~70% of the total moment is released between 20 to 40 s. For this fault plane, 206 sub-faults have slip >10% of the peak slip. The average slip and rise time over these sub-faults are 1.08 m and 3.9 s, respectively. The mean slip rate is about 0.4 m/s, locally reaching 2.8 m/s. For this fault plane, the estimated static stress drop is about 2.2 to 5.1 MPa.

For the two fault planes, we observe that the slip distribution is relatively compact and located near the hypocenter. The moment rate function also looks very similar in both cases (Figure 2). These two features would explain why the seismic data alone could not distinguish

between the two possible causative fault planes. Generally, our results are consistent with Ye *et al.* (2014). They studied this earthquake using teleseismic body waves only. Although they have chosen the shallow-dipping fault plane as the causative fault plane, the difference in the waveform fit is very small between the two planes (Ye *et al.*, 2014).

We have also forward modelled the co-seismic static motions at the surface (see Section S3 in the Supplementary Material for discussion of the results). Because the 2014 earthquake is located in a remote area, only the static offsets from three GPS stations can be used. We find that, considering the uncertainty in the PDE-NEIC epicenter location, both models can fit these observations well, but the model with the steeply-dipping fault plane appears more favorable. However, because the predictions of the co-seismic static motions are sensitive to the slip distributions, this argument is subject to great uncertainties.

We note that the stress drop estimated for each fault plane is within the range inferred for other intermediate-depth earthquakes (0.2 - 10 MPa, with a mean value of 2 MPa; Houston and Williams, 1991). However, it should be emphasized that stress drop estimates usually suffer of substantial uncertainties associated to the methods of estimations and smoothness of the slip model (for instance, Ye *et al.* (2014) have found a stress drop of about 11 MPa for the 2014 Rat Islands earthquake).

3. Joint hypocenter determination (JHD) analysis

The 2014 Rat Islands earthquake is followed by a very active aftershock sequence, which is abnormal for an intermediate-depth earthquake. Perch and Houston (2004) have systematically studied aftershock sequences of 290 $M_w > 6.3$ intermediate-depth and deep earthquakes in the GCMT catalog from 1977 to 2002. They found that within 50 km of a mainshock GCMT

centroid and for the first 20 days after the mainshock origin time, the number of $m_b > 4.5$ aftershocks in the PDE-NEIC catalog is less than 10. As an example, the 1978 $M_w 7.8$ Kuril Islands intermediate-depth earthquake, which has similar magnitude as the 2014 Rat Islands mainshock, has only three such aftershocks. In contrast, we note that after the 2014 Rat Islands earthquake, the PDE-NEIC reports 27 $m_b > 4.5$ aftershocks within the same spatial and temporal criteria, about three times of the most active aftershock sequence reported by Perch and Houston (2004).

This abnormally active aftershock sequence of the 2014 Rat Islands earthquake provides a unique opportunity to constrain its causative fault plane. However, Ye *et al.* (2014) have pointed out that the distribution of PDE-NEIC aftershock locations do not show a clear trend that could help to discriminate the causative plane and its auxiliary. We suspect that this is a result of the errors associated with these network routine locations, particularly for the earthquakes with small magnitude ($m_b < 3.0$). Here, we apply a master event method (Joint Hypocenter Determination; See Supplementary Material S4 for more details on the method) to improve the relative locations of these aftershocks with respect to the hypocenter of the 2014 Rat Islands earthquake. As we show next, the relocated hypocenters clearly align along the steeply-dipping fault plane.

We use for the JHD relocation analysis the arrival times of seismic phases (P, S, pP, sP, PcP, and ScP) reported by the International Seismological Centre (ISC; International Seismological Centre, *EHB Bulletin*, <http://www.isc.ac.uk>, Internatl. Seis. Cent., Thatcham, United Kingdom, 2012). From June 23, 2014 to September 23, 2014, the ISC reports a total of 118 $m_b \geq 3$ earthquakes whose epicenters are located within 60 km of the main shock epicenter. Ninety-seven of them have over 10 reported arrival times and are selected for

the JHD analysis. After the relocation, we choose only 36 high quality JHD hypocenters for further discussions. For these earthquakes, the semi-major axis of the 90% confidence ellipse is less than 30 km. As expected, there is a clear bias towards earthquakes with larger magnitude. The selected dataset includes 67% of the $m_b \geq 4.0$ aftershocks (19 earthquakes) originally reported by the ISC catalog, but only 17% of the $3.0 \leq m_b < 4.0$ aftershocks (17 earthquakes). The mean of the horizontal error of these JHD locations is 6.2 km, while the mean vertical error is 19.8 km. The offsets between the original ISC locations and the corresponding JHD locations are significant. The horizontal offsets range from 1.2 to 31 km with a mean of 7 km, while the vertical offsets range from 0.3 to 41 km, with a mean of 14 km (see Section S4 of the Supplementary Material for a more detailed discussion on the results of the relocation and their uncertainties).

In a map view (Figure 3), we observe that most relocated aftershocks distribute along a 40 km long linear segment orienting northeast and that coincides with the width of the surface projection of the steeply-dipping fault plane. Vertically, the relocated aftershocks span a depth range from 70 to 150 km. It is noteworthy that the depth extension of the relocated aftershocks distribution is nearly the double of its horizontal extension.

Profiles (AA') and (BB') in Figure 3 are cross-sections that run perpendicular to the strike of the steeply-dipping and the shallow-dipping fault planes, respectively. For each cross-section we show the fault plane and the relocated aftershocks. To highlight the relationship between aftershocks distribution and slab position, we denote with a continuous purple line the top of the slab taken from the slab1.0 model of Hayes *et al.* (2012). We mark the bottom of an assumed 40 km thick slab by a purple dashed line (see Section 4.2 for detailed discussion about

this definition). Note that because the fault planes obliquely cut through the subducting interface, the depth of the slab interface changes along the fault plane strike on profile (AA') and (BB'). It can be seen that the relocated aftershocks align nicely along the fault plane on profile (AA'), while are much more diffuse on profile (BB'). We also note that most relocated aftershocks, with the exception of four earthquakes, are located inside the subducting slab and extend over its entire width. Profiles (CC') and (DD') in Figure 3 are fault plane view of the steeply-dipping and the shallow-dipping fault planes, respectively. We note that on profiles (CC') and (DD'), where we only project on the fault plane the aftershocks that are <25 km away from it. We observe that the aftershocks on profile (CC') concentrate on the region of large slip; they cut through the high slip patch on profile (DD'). We also observe that the slip associated with the steeply-dipping fault plane (profile CC') is essentially contained within the slab, which is not the case for the shallow-dipping fault plane.

Hence, although both waveform modeling and forward modeling of the co-seismic displacement fail to distinguish the causative fault plane, the relocated aftershocks distribution tends to agree better with the steeply-dipping fault plane.

4. Discussions

Before putting in perspective our results with respect to the local tectonics, as well as with respect to other large intermediate-depth earthquakes, let us quickly review the potential sources of errors and discuss whether they might undermine our further discussions. When we use teleseismic data to constrain the source model, the inverted spatial distribution of the fault slip is relative to the hypocenter. A similar argument also applies for the JHD aftershock locations. The location error of hypocenter then should not affect our argument about the

causative fault plane but might affect the comparison of the inverted slip model with the local tectonic features and near fault observations. In Section S3 of the Supplementary Material, we show that after shifting the epicenter location by 21 km N132°E, our preferred model could predict the observed coseismic static displacement at station AC66 reasonably well. This suggests that all the earthquake epicenters in this region might suffer the same systematic offset. Also, because the plate interface model Slab1.0 (Hayes *et al.*, 2012) is inferred from the PDE-NEIC catalog and the relocated ISC catalog (Engdahl *et al.*, 1998), the Slab1.0 model in this region might suffer a similar systematic offset as well. In contrast, the thickness of the slab that undergoes brittle deformation is inferred from relative earthquake distributions and thermal considerations and thus is more robust.

4.1 Correlation with the regional tectonics

The central segment of the Aleutian Arc, where the 2014 Rat Islands earthquake has occurred, is tectonically complicated. To the south, the strike of the Aleutian trench axis changes from ~250-260° on the eastern side to ~270-280° on the western side, and the direction of plate convergence changes from oblique to nearly parallel (i.e., nearly pure strike-slip) to the trench axis. To the north, the Bowers ridge, a remnant Cenozoic island arc (Wanke *et al.*, 2012), comes in to contact with the Aleutian Arc. In the Central Aleutian arc, three tectonic blocks, named Buldir, Rat, and Delarof block, make up the intra-arc region (Figure 4a; Geist *et al.*, 1988).

Based mainly on geomorphological observations, Geist *et al.* (1988) have concluded that these three blocks have rotated clockwise by 14°, 20°, 23°, respectively, around their poles of rotation near the northwestern corners of each blocks (Figure 5a). They further indicate that all the pivot points of each block should migrate westward to avoid compression between the blocks (see Figure 5a). If this speculation is correct, at least the northwest portion of the northern

boundary of the Rat block should be associated with dextral motion relative to the Bowers Ridge. However, no $M_w > 5.0$ shallow (< 30 km) strike-slip earthquakes along this boundary has occurred since 1976 (Ruppert *et al.*, 2012).

The 2014 earthquake has a significant but not uniform right-lateral strike-slip component motion ($\sim 45\%$ of the total seismic moment – $M_w 7.7$). Figure 5(bc) shows the predicted horizontal and vertical displacement field at the surface, respectively, using the preferred slip model. First, Figure 5b shows that most of the strike-slip motion concentrates on the western half of the fault plane, near the northwestern edge of the Rat block. It is dominated by the motion toward the trench, opposite to the horizontal motion expected by the locked section on the shallow portion of the plate interface. Secondly, Figure 5c shows that the occurrence of the 2014 Rat Islands earthquake has uplifted the Rat block, which is consistent with the set of islands demarcating its northern boundary. If we only focus on the fault-parallel motion (Figure 5d), we observe that it has opposite directions on the two sides of the fault, inducing a dextral motion. Thus, this moves the pivot point of the Rat block westward as predicted by Geist *et al.* (1988). In short, our results indicate the 2014 intermediate-depth earthquake has contributed to the westward transporting motion of the Rat block.

The subducting Pacific plate is complicated as well. The longitudinal strike change of the Aleutian trench axis is carried by the subducting slab (Hayes *et al.*, 2012). There is also the subduction of the Rat fracture zone (RFZ; Figure 4a), which is associated with the Kula ridge that subducted down the Aleutian trench > 10 Myr ago (Geist *et al.*, 1988). Furthermore, the purple star on Figure 4a shows the location of the 1965 Longshot underground nuclear explosion (Abe, 1972). Using the data from this explosion, Abe (1972) revealed the existence of a low-velocity region beneath it, which is interpreted as the result of a tear of the subducting Pacific

plate. As we discuss next, the lateral extent of the 2014 Rat Islands earthquake is in fact consistent with this complicated tectonic environment.

Figure 4a also shows the surface projection of the preferred solution for the 2014 Rat Islands earthquake. The two headed arrow denotes the estimated of along strike extension of its rupture (i.e., corresponding to 95% of the inverted seismic moment), spanning a fault length of about 127 km. We note that the heterogeneities of the subducting Pacific slab might have controlled the lateral extent of the 2014 Rat Islands earthquake. To the west, the rupture ends where it meets the assumed down-dip extension of the subducting RFZ. We also observe on this figure that the density of intermediate-depth earthquakes dramatically changes at the location where the rupture stops. To the east, the slip stops when it encounters the tearing of the subducting Pacific slab. Thus, we see that the fault boundaries (east and west) coincide with significant tectonic features of the down-going plate suggesting a structural control on the lateral extent of this earthquake.

Lastly, we notice an interesting correlation between the lateral extent of the slip and the geological features at the surface, though their physical relationship is not clear yet. The strike of the steeply-dipping fault plane (308°) for the 2014 earthquake is parallel to the northern boundary of the Rat block ($\sim 305^\circ$; Figure 4a) rather than the strike of the trench ($\sim 290^\circ$). The surface projection of the rupture to the west correlates with the extension of Murray canyon, which is the west boundary of the Rat block. To the east, where the slip stops, it is also correlated to the eastern edge of the Bowers ridge (3 km iso-bathymetry contour highlighted by a continuous black line Figure 4a). However, further investigations would be necessary to determine if there could be causal relationships between intermediate-depth earthquakes and tectonic features of the overriding plate.

4.2 Correlation between slip distribution of large intermediate-depth earthquakes with thermal structure of the oceanic plate before and after subduction.

Most intermediate-depth earthquakes are intraplate earthquakes associated with subducted oceanic plates (Kirby, 1995). Recent studies of earthquakes along the transform faults and beneath outer-rise regions have revealed that the maximum depth for brittle failure in the oceanic lithosphere corresponds to the approximate depth of the 600-650°C isotherm (e.g., Abercrombie and Ekström, 2001; Craig *et al.*, 2014). Coincidentally, a similar temperature criterion is also applied for intermediate-depth earthquakes at depths < 150 km, though the mechanism for rupture might be different. Yamasaki and Seno (2003) have found a good agreement between the dehydration loci of serpentine dehydration and the loci of intermediate-depth earthquakes for six subduction zones. Later, Brudzinski *et al.* (2007) have found evidence of DWBZs in all of the 16 subduction zones they surveyed. They have also noticed that the separation between the upper and lower planes of the DWBZs could be better explained by the dehydration of antigorite (serpentine). At the pressure of intermediate-depth earthquakes, the dehydration of serpentine occurs at ~600-650°C (e.g., Yamasaki and Seno, 2003; Hacker *et al.*, 2003). For the convenience of further discussion, we refer to the region of the subducting slab with temperature less than 600-650°C (i.e., the region between the solid and dashed lines on Figure 3 and 4b), as the cold core of the downgoing slab.

Using a simple half-space cooling model and the plate age of the subducting Pacific plate near the central Aleutian trench (~58 Ma), the 600-650°C isotherm depth is expected to be about 37 to 40 km below the surface. Because of the fast convergence rate (8 cm/yr), this thickness will vary only slightly at the source depth of the 2014 Rat Islands earthquake (~100 km). Craig *et al.* (2014) pointed out that the maximum earthquake depth in the outer-rise region

of the Aleutian Arc is ~30 km and no outer-rise thrust fault earthquakes have been observed yet. However, this is probably a result of either short observation period or low bending stresses near the bottom of the seismogenic zone (Craig *et al.*, 2014).

Figure 4b shows the reported ISC hypocenters (blue circles) from 1960 to 2012 projected on the steep-dip angle fault plane. It shows that the region highlighted by the solid and dashed lines host nearly all the intermediate-depth earthquakes of this catalog. Considering the errors of the earthquake locations, which is based mainly on teleseismic observations, this correlation is remarkable. Hence, on the relevant figures, the 40 km thick region between the continuous and dashed purple lines is the region of the subducting slab that is capable to host brittle failure before the subduction as well as after subducting to depth of 100 km. However, when we discuss the deeper seismicity, more precise thermal calculations are required. We also show on Figure 4b the distribution of the relocated aftershocks (green triangles). It is interesting to note that the regions of high slip during the 2014 Rat islands earthquake more or less fill the “holes” in the seismicity from 1960 to 2012. We also observe a correspondence between the seismicity before the 2014 earthquake and the distribution of the relocated aftershocks.

As we have mentioned above, the majority of the inverted slip for our preferred model is located inside the cold core of the slab, and so do the relocated aftershocks (Figure 4b), consistent with the above thermal based model. The portion of the slip outside the slab breaks after 30 s. This is the time where the moment-rate function begins to decrease, indicating the termination of the rupture process (Figure 2). Thus, it is likely that this spurious slip contribute only to the fit of the late part of the waveforms and are not a reliable portion of the source process. In contrast, if we choose the low angle fault instead, a significant portion of the inverted fault slip is located outside the cold core of the slab (Figure 3). Hence, our study suggests that the

steeply-dipping fault plane is more consistent with what we know of the characteristic of the subducting slab in this region. It further suggests that the 2014 M_w 7.9 Rat Islands earthquake has broken the entire cold slab core.

4.3 Downdip extension of large intermediate-depth earthquakes

Brudzinski et al. (2007) conducted an in-depth analysis of intermediate-depth seismicity. They observed that in Central Aleutian the seismic activity is distributed over two distinct planes with low seismic activity in between. Their study show that the two zones are separated by about 15-20 km. Using thermal-petrologic models for subduction zones, Hacker *et al.* (2003) have predicted that the intermediate-depth earthquake hypocenters correspond spatially to the dehydration reactions that occur in the crust and/or the uppermost mantle. In contrast, seismic gaps between the DWBZs correspond spatially to the cold core of the subducting slab. However, it is not clear whether this portion of the subducted slab is aseismic. Kirby *et al.* (1996) suspected that large intermediate-depth earthquakes concentrate near the top surface of the slab, inside the subducted crust. Tibi *et al.* (2003) surveyed six large intermediate-depth earthquakes with magnitude from 7.1 to 7.7. They found that along strike extensions of large intermediate-depth earthquakes are often several times of their in-depth extensions. However, the depth extensions are up to 13 km, which is thicker than the average thickness of the oceanic crust but is much more thinner than the depth extension of the 2014 Rat Islands earthquake. As mentioned above, our rupture model reveals that this earthquake has broken the entire cold slab core.

In the fifteen $M_w > 7.5$ intermediate-depth earthquakes since 1976, the 1993 M_w 7.6 Kushiro-Oki, Japan, earthquake (Ide and Takeo, 1996), and the 2005 M_w 7.7 Tarapaca, Chile, earthquake (Delouis and Legrand, 2007), are the only two intermediate-depth earthquakes that 1) its causative fault plane is defined using the aftershock distribution located with a local seismic

network and 2) the slip distribution is constrained using local strong-motion observations. Both of them occurred in fast subduction zones. In Figure 6(abc), we show a cross-section along a line running perpendicular to the strike of the causative fault plane for the 2014 Aleutian earthquake, the 2005 Chile earthquake and the 1993 Japan earthquakes. We also show the projections of the determined causative fault plane for these three earthquakes (this study; Ide and Takeo, 1996; Delouis and Legrand, 2007). A red line is used to denote the region with significant slip. Again, we approximate the cold core of the slab with the continuous and dashed purple lines again based on the Slab1.0 model and the age of the subducting seafloor. As for the Aleutian region, there is an excellent agreement between the cold portion of the slab and the background seismicity. The figure also clearly illustrates that most of the cold core slab thickness, if not the whole, is broken during those earthquakes (this study; Ide and Takeo, 1996; Peyrat *et al.*, 2006; Delouis and Legrand, 2007). Thus, available seismological evidence suggests that seismic gaps between the DWBZs are not aseismic and can rupture during large intermediate-depth earthquakes.

We note that serpentine dehydration embrittlement (Yamasaki and Seno, 2003) or shear-heating instability mechanism due to creep (Kelemen and Hirth, 2007) typically occurs at relative high temperature (600-650°C; 600-800°C) in comparison to the cold core of the slab (<600-650°C). However, under certain adjustments, these mechanisms can occur at colder temperatures. For instance, if minerals other than antigorite are involved in the dehydration embrittlement process, the dehydration reactions can occur at lower temperatures (Hacker *et al.*, 2003). In regards to shear-heating, once the rupture is triggered, the shear heating generated during the co-seismic rupture is faster than the dissipation by conduction. The melting of rock along the narrow fault zone lubricates the fault by balancing the confining pressure dynamically, and thus leads to thermal runaway (e.g., Kanamori *et al.*, 1998; Prieto *et al.*, 2013).

We also note that the spatiotemporal pattern of the intermediate-depth earthquakes has great similarity with interplate earthquakes in the shallow portion of the subduction zone, where megathrust earthquakes occur. As shown by the diagram in Figure 7, the downdip seismogenic limit for interplate earthquakes in most subduction zones appears to agree with either a maximum temperature of 350°C or the thrust intersection with the forearc serpentinitized mantle (e.g., Hyndman *et al.*, 1997). There is often a tapered transition downdip of large earthquake rupture zone, accompanied by creeping motion and that extent down to the 450° isotherm. During the interseismic period, the region with significant slip during the future large earthquake (also often colder than 350°C in temperature) is silent and with frequent small to moderate earthquakes, as well as creeping motion, along the tapered transition zone. For the intermediate-depth region, frequent small to moderate earthquakes mark the DWBZs, while the elastic energy is accumulated in the colder portion of the slab, which is silent during the interseismic period. A large intermediate-depth earthquake thus occurs when the rupture of the colder portion of the slab is triggered dynamically (Figure 7). A careful study of some of others large intermediate-depth earthquakes would allow to see if this is a consistent pattern for large intermediate-depth earthquakes, and thus would enable to see whether the analogy stands.

5. Conclusions

In this study, we have first performed a set of finite-fault inversions using body waves and long-period (3-6 mHz) seismic waves in order to invert for the slip history of the 2014 Rat Islands earthquake. We find that by using this type of data alone, we cannot determine the causative fault plane, the misfit reached by the inversions for each fault plane being the same.

Thus, using phase picks reported by the ISC, we have relocated three months of

481 aftershocks from the abnormally active aftershock sequence of the 2014 earthquake using a JHD
482 method. We find that they tend to align preferentially along the fault plane that has a dip of 87°
483 and a strike of 308° .

484 The rupture process associated with this plane shows a bilateral rupture at ~ 2 km/s
485 breaking essentially one large slip patch (3.5 m of peak slip amplitude) west of the hypocenter.
486 The inverted slip distribution extends from 80 to 140 km depth, which would be the largest depth
487 extent of earthquake slip ever documented.

488 We find that the lateral extent of this earthquake is bounded by two major tectonic
489 features of the subducting slab: The Rat Fracture zone to the west and a lithospheric tear to the
490 east. From the calculated static displacement field, we suspect that intermediate-depth
491 earthquakes might contribute to the motion of the tectonic block at the surface. We also observe
492 an interesting correlation between the preferred fault plane and the geological features observed
493 at the surface.

494 We show that the inferred slip distribution of the 2014 earthquake is confined inside a
495 region delimited by the top of the slab and the depth of the $600\text{--}650^\circ\text{C}$ isotherm. Inside this
496 region is also contained most, if not all, of the intermediate-depth earthquakes prior to the 2014
497 mainshock, as well as the relocated aftershocks.

498 A similar observation is made for two other large intermediate-depth earthquakes (the
499 1993, Koshi-Oki, Japan, earthquake, and the 2005, Tarapaca, Chile, earthquake), suggesting
500 that seismic gaps between the DWBZs are not aseismic and can rupture during large
501 intermediate-depth earthquakes.

502 This is of importance regarding seismic hazards assessment. As the strain drop of
503 earthquakes is roughly constant, independent of earthquake depth and magnitude (Vallée, 2013),

504 whether the rupture of large intermediate-depth earthquakes is limited to the subducting crust
505 (Kirby *et al.*, 1996) or the entire elastic core of the subducting slab will strongly affect the
506 magnitude estimates of potential earthquakes.

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Figures:

Figure 1. Tectonic setting of the Alaska-Aleutian region. The black arrows show the motion of the Pacific Plate with respect to the fixed North American Plate (DeMets *et al.*, 2010). The focal mechanisms of all $M_w \geq 6.0$ earthquakes from the GCMT catalogue from January 1, 1976 to June 23, 2014 are shown color-coded with respect to depth. The $M_w 7.9$ June 23, 2014 is shown by the largest focal mechanism (yellow). The dark and light purple regions are the rupture extent of the 1957 $M_w 8.6$ Great Aleutian earthquake (Johnson *et al.*, 1994), and the 1965 $M_w 8.7$ Rat Islands earthquake (Johnson and Satake, 1996), respectively. In the bottom left corner, we show the station distribution used for the finite-fault inversions (red triangles), and the epicenter of the 2014 earthquake (blue star).

Figure 2. The top row shows the inversions that use the steeply-dipping fault plane (Nodal Plane 1), and the bottom row shows the inversions that use the shallow-dipping fault plane (Nodal Plane 2). The furthest left column shows the moment rate functions calculated from the rupture models obtained by the inversion group C. The remaining three columns show the rupture processes obtained for each group of inversion. In these figures, the black lines show rupture time contours every 10 s. The blue arrows indicate the slip direction. The red star shows the location of the hypocenter at 107.5 km depth for both planes. Table 1 shows a description of each of the inversions.

Figure 3. [Left panel] Map view of the slip distribution obtained for the two nodal planes along with the $m_b \geq 3$ high quality JHD aftershocks (i.e., with a 90% confidence ellipse axes < 30 km), color-coded with respect to depth. [Middle panel] Profile (AA') and (BB') are cross-sections that run perpendicular to the strike of Nodal Plane 1 (steeply-dipping fault) and Nodal Plane 2 (shallow-dipping fault), respectively. The black line is the projection of the fault plane. The continuous purple line shows the top of the slab from the USGS Slab1.0 model (Hayes *et al.*, 2012). The purple dashed line represents the bottom of the slab (600–650°C isotherm). We also project the high quality JHD aftershocks that are within the cross-section boxes shown on the left panel. [Right panel] Fault plane view of the obtained slip distribution for the two nodal planes. The continuous and dashed purple lines represent the intersection of the slab boundaries with the fault plane. Here, we only project the high quality JHD aftershocks that are < 25 km away from the fault plane. Note that on these profiles, the depth markers are just shown as indicators. For the Nodal Plane 1, the x-axis is 178.5 km long and the y-axis is 112.5 km long. For the Nodal Plane 2, the x-axis is 112.5 km long and the y-axis is 178.5 km long.

Figure 4. (A) Map of the tectonic blocks from Geist *et al.* (1988). The continuous line to the north shows the 3 km iso-bathymetry contour, highlighting the Bowers ridge. The line named RFZ, south of the subduction trench, shows the Rat Fracture Zone and the dashed line shows its assumed down-dip extension. The dots shows intermediate-depth earthquakes ($70 < \text{depth} < 300$) from the ISC catalog. All earthquakes are color-coded with respect to their hypocentral depth. The slip distribution of the preferred fault plane (i.e., the steeply-dipping one) is shown and a two headed arrow shows the lateral extent of 95% of the slip. The pink star shows the location of the 1965 Longshot explosion experiment. (B) Fault plane view of the slip distribution obtained for the preferred fault plane. Note that we only show the slip that is larger than 10% of the peak slip amplitude. The blue dots show the ISC hypocenters at intermediate-depths from 1960 to 2012, while the green triangles show the high quality JHD aftershocks. We only show

earthquakes that are < 25 km away from the fault plane. The continuous and dashed lines highlight the cold core of the subducting slab.

Figure 5. (A) Rotation of blocks along an arc with a constant orientation of forces (reproduction of Figure 10 in Geist *et al.*, 1988). The dashed lines denote the original positions of the arc and the solid lines show the current positions after block rotation. Note that the pivot of these blocks is transported to the west. BB stands for Buldir block, RB stands for Rat block, and DB stands for Delarof block. (B) Co-seismic horizontal surface displacement vectors (cm) calculated from the slip distribution of the preferred model (i.e., the steeply-dipping fault plane; Figure 2). The black dashed line show the surface projection of the fault plane. The continuous black line outlines the boundaries of the Rat block (Geist *et al.*, 1988). (C) Co-seismic vertical surface displacement field (cm). (D) Co-seismic horizontal fault-parallel displacement field (cm). Red colors are for east-southeast motion, while blue colors are for west-northwest motion.

June 23, 2014 Aleutian earthquake

June 13, 2005 Chile earthquake

January 15, 1993 Japan earthquake

Figure 6. Set of cross-sections running perpendicular to the strike of the fault plane for the 2014 Rat Islands, Aleutian, earthquake (A), the 2005 Tarapaca, Chile, earthquake (B), and the 1993, Kushiro-Oki, Japan, earthquake (C). On each figure, the trench is located on the left side. The continuous purple line shows the top of the slab from the USGS Slab1.0 model (Hayes *et al.*, 2012). The purple dashed line represents the bottom of the slab (600-650°C isotherm). The dots shows the reported ISC hypocenters from 1960 to 2012. The black lines show the causative fault plane (this study for the 2014 Aleutian earthquake; Delouis and Legrand (2007) for the 2005 Tarapaca, Chile earthquake; Ide and Takeo (1996) for the 1993 Kushiro-Oki, Japan earthquake), while the thick red line shows the portion of the fault where most of the slip occurs.

Figure 7. Schematic illustrating the analogy between shallow interplate earthquakes and intermediate-depth intraplate earthquakes.

Table 1. Summary of the 3 groups of inversions carried out in this study. For each, the first row shows the name of the group, the second row, the dataset used, the third row, the misfit obtained when using the steeply-dipping fault plane, and the fourth row, the misfit obtained when using the shallow-dipping fault plane. Waveform fit for Inversion Group C is shown on Figures S7 to S12 in the Supplementary Material.

	Inversion A	Inversion B	Inversion C
Dataset	Only P-waves	P- and SH-waves	P-waves, SH-waves, and long-period seismic waves
Nodal Plane 1 (steeply-dipping)	0.263	0.246	0.236 (Figures S7 to S9)
Nodal Plane 2 (shallow-dipping)	0.254	0.239	0.230 (Figures S10 to S12)

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