Evaluation of dip angles of active faults beneath the Osaka Plain inferred from a 2D numerical analysis of visco-elasto-plastic models

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Abstract

The geometries (i.e., dip angles) of active faults from the surface to the seismogenic zone are among the most important factors used to evaluate earthquake ground motion, which is crucial to seismic hazard assessments in urban areas. In Osaka, a metropolitan city in Japan, there are several active faults (e.g., the Uemachi and Ikoma faults), which are inferred from the topography, the attitude of active faults in surface trenches, the seismic reflection profile at shallow depths (less than 2 km), and the three-dimensional distribution of the Quaternary sedimentary layers. The Uemachi and Ikoma faults are N–S-striking fault systems with total lengths of 42 km and 38 km, respectively, with the former being located ~12 km west of the latter; however, the geometries of each of the active faults within the seismogenic zone is not clear. In this study, to examine the geometries of the Uemachi and Ikoma faults from the surface to the seismogenic zone, we analyze the development of the geological structures of sedimentary layers based on numerical simulations of a two-dimensional visco-elasto-plastic body under a horizontal compressive stress field, including preexisting linear high-strained weak zones (i.e., faults) and surface sedimentation processes, and evaluate the relationship between the observed geological structures of the Quaternary sediments (i.e., the Osaka Group) in the Osaka Plain and the model results. Based on a comparison between the simulation results and the geological observations/interpretation, we propose geometries of the Uemachi and Ikoma faults from the surface to the seismogenic zone. When the friction coefficient of the faults is ~0.5, the dip angles of the Uemachi and Ikoma faults near the surface are ~30°–40° and the Uemachi fault has a downward convex curve at the bottom of the seismogenic zone but does not converge to the Ikoma fault. Based on the analysis in this study, the dip angle of the Uemachi fault zone is estimated to be approximately 30°–40°, and the downward extension of the Uemachi fault zone nearly coincides with the epicenter of the 2018 northern Osaka earthquake.

Keywords: dip angle, fault geometry, visco-elasto-plastic simulation, Uemachi fault zone, Ikoma fault zone, Osaka Group, Osaka Plain, 2018 northern Osaka earthquake, active fault, seismogenic source fault

Introduction

The 2018 northern Osaka earthquake that occurred on June 18, 2018, was located at the junction of the Arima-Takatsuki fault zone and the Ikoma fault zone. The focal mechanism solution for this earthquake showed a N–S-striking reverse fault type (Kato and Ueda 2019). If the dip angle of the Uemachi fault zone, which is a reverse fault with a N–S strike and E dip, is approximately 40°, the downward extension of the Uemachi fault zone would nearly coincide with the epicenter of the earthquake (Fig. 1). However, the relationship between this earthquake and the Uemachi fault zone is still unknown because the attitude (i.e., the strike and dip angle) of the Uemachi fault zone at the depth of the seismogenic zone is not well understood. Similar to the Uemachi fault zone, the attitude of active faults, including the Ikoma fault zone, in the Osaka Plain within the seismogenic zone (at a depth of ~10–15 km) is not clear. The topography, the attitudes of active faults in surface trenches, seismic reflection profiles at shallow depths (less than 2 km), and the three-dimensional distribution of the Quaternary strata based on bored geological columns have been used by the Headquarters for Earthquake Research Promotion of MEXT and the Central Disaster Prevention Council to evaluate the activities of each active fault. The deep underground attitudes of active faults have been estimated assuming that the attitude near the surface hardly changes with depth (e.g., Headquarters for Earthquake Research Promotion of MEXT 2001, 2004; Director General for Disaster Management 2006; Fig. 1). To mitigate earthquake damage, it is essential to predict ground motion caused by earthquakes that occur on active faults, and the attitude of the source fault in the seismogenic zone is the most important factor used to predict the ground motion.

To estimate the dip angles of the active faults at depth in the Osaka Plain, numerical analyses were performed to determine how the deformation of the strata and the ground surface changes as a result of different fault dip angles; the simulation results can be compared to the distribution of the strata in the Osaka Group and the inclination angle of the axial plane of the subsurface flexure of the Uemachi fault (Ishiyama 2003; Iwasaki 2016; Fig. 1). However, these different dip angles were obtained by forcibly deforming the pre-deposited strata via displacement at the fault; factors, such as the fault behavior under the assumed stress field, the rheological properties of the rocks and sediments, and sedimentation and erosion, have not been considered.

In this study, to examine the geometries of the Uemachi and Ikoma faults at the depth of the seismogenic zone, we analyze the development of geological structures in the sedimentary layers based on numerical simulations of a two-dimensional visco-elasto-plastic body under a horizontal compressive stress field, including preexisting linear high-strained weak zones (i.e., faults) and surface sedimentation processes, and evaluate the relationship between the observed geological structures of the Quaternary sediments (i.e., the Osaka Group) in the Osaka Plain and the model results.

Constraints on the numerical simulation: Quaternary geology in the Osaka Plain

The Osaka sedimentary basin is an oval topographical depression surrounded by mountains, and the Osaka Bay and the Osaka Plain are located in its western and eastern regions, respectively. The boundary between the Osaka Plain and the surrounding mountains is demarcated to the north by the ENE–WSW-trending Arima-Takatsuki fault zone and to the east by the N–S-trending Ikoma fault zone. The N–S-trending Uemachi fault zone is located in the central part of the Osaka Plain (Fig. 2a). Quaternary crustal deformation in and around the Osaka sedimentary basin is referred to as the Rokko movements (Ikebe and Huzita 1966; Huzita 1968, 1990) and result from the subduction of the Pacific Plate at the Japan Trench and that of the Philippine Sea Plate at the Nankai Trough (Huzita 1968; Itoh et al. 2000). Based on a comparison between the present stress conditions, the direction of the horizontal maximum stress (σH_{max}) is currently nearly E–W (Tsukahara and Kobayashi 1991; Terakawa and Matsu'ura 2010), and the conditions determined by inverting the fault-slip data from active faults that have exhibited cumulative displacement for the past ~10^5 years, it has been suggested that the stress field in central Japan has been uniform and stable for the past ~10^5 years (Tsutsumi et al. 2012). Wesnousky et al. (1982) estimated the geological horizontal shortening strain rate of central Japan, including in the Kinki region, to be 16–26 × 10^{-9}/yr, based on earthquake records with a magnitude of 6.9 or greater and the displacement rates of active faults for the last ~400 years. We first summarize the Quaternary geology in the Osaka Plain, especially that of the Osaka Group, Uemachi fault zone, and Ikoma fault zone, because the geometry in these regions constrains the model parameters and results.

Osaka Group

The Osaka sedimentary basin is filled by the strata of the Plio–Middle Pleistocene Osaka Group, Middle–Upper Pleistocene terrace deposits and their corresponding sediments, and Upper Pleistocene–Holocene alluvium. The strata are 1,000–2,000-m thick at the deposition center and 200–400-m thick at the margins (Itihara 1993). Yoshikawa and Mitamura (1999) reported that the Quaternary system in the Osaka Plain consists of unconsolidated clay, silt, sand, and gravel layers with a thickness of more than 1,500 m and dozens of volcanic ash layers; this system has been divided into the Miyakojima Formation in the lower part, the Tanaka Formation in the upper part, and the Namba Formation in the uppermost part (Fig. 2b). The Miyakojima Formation consists of freshwater layers primarily composed of gravel, sand, and silt layers, whereas the Tanaka and Namba formations consist of freshwater sand and gravel layers and 21 marine clay layers (Ma-1, Ma0, Ma0.5, Ma1, Ma1.3, Ma1.5, Ma1.7, Ma2,..., Ma11 (1), Ma11 (2), Ma12, and Ma13).

Since the 1995 Hyogoken-Nambu earthquake, multiple drilling and seismic reflection surveys have been conducted in the Osaka Plain, revealing details concerning its subsurface structure. Ikebe et al. (1970) analyzed nine deep drilling cores (OD-1–OD-9) and found that the Osaka Plain can be divided into two areas (i.e., west and east Osaka) by the N–S-trending Uemachi fault zone running through the central part of the plain. The basement depth in west Osaka is more than 1,000 m (because the OD-1 core did not reach basement rock at a depth of 907 m), whereas the basement depth in the Uemachi Upland and its northern extension is as shallow as 656 m (OD-2; Fig. 2b). The basement depth in the central part of the Osaka Plain was examined based on the basement structures inferred from gravity anomalies (Nakagawa et al. 1996a; Kansai Geo-informatics Council 1998) and reflection seismic surveys (Ikebe et al. 1970). Using this information, Uchiyama et al. (2001) and the Osaka Prefecture (2004) found that the depth of the basement surface in west Osaka is nearly 1,500 m and does not significantly change to the west; meanwhile, in east Osaka, the depth is approximately 800 m near the Uemachi Upland and more than 1,500 m near the Onchi River and the surface of the basement slopes to the east (Fig. 3). In addition, the strata thickness does not change in west Osaka, whereas the strata thickness in east Osaka tends to increase eastward (Uchiyama et al. 2001). In detail, in east Osaka, the thickness of the

Miyakojima Formation is approximately 550 m at the Uemachi Upland and more than 700 m near the Onchi River and the thickness of the lower Tanaka Formation (Ma-1–Ma6) is approximately 250 m at the Uemachi Upland and 850 m near the Onchi River.

According to Uchiyama et al. (2001), the sedimentation rate from ~1.2 million to ~0.05 million years ago (Ma) in west Osaka, obtained from the OD-1 core, gradually decreased from 0.7 m/kyr to 0.2 m/kyr and the rate of the decrease in the sedimentation rate was higher after ~0.4 Ma. In east Osaka, in the western region near the Uemachi Upland (OD-2, OD-9, and YU), the sedimentation rate from ~1.2 Ma to ~0.6 Ma, corresponding to the lower Tanaka Formation, decreases from 0.5 m/kyr to 0.3 m/kyr, whereas the sedimentation rate decreases from 0.8 m/kyr to 0.5 m/kyr in the eastern region (OD-3). The Tanaka Formation around the Uemachi Upland is preserved in conformity only from the Ma3 layer to the Ma7 layer and is partially covered in an unconformity by the Ma12 layer. Suzuki (2016) estimated the sedimentation rate of the Miyakojima Formation to be 0.48 m/kyr using the deep drilling core data of Yoshikawa et al. (2000).

Uemachi fault zone

The Uemachi fault zone is a nearly N–S-striking fault system with a total length of 42 km, including the Butsunenji-yama fault, Uemachi fault, Nagai fault, Sakamoto fault, Kumedaike fault, Sakuragawa flexure, and Suminoe flexure. It is a reverse fault in which the eastern side of the fault zone has moved upward relative to the western side (Nakada et al. 1996a, 1996b, 1996c, 1996d; Okada and Togo 2000). It has been confirmed that the Uemachi fault zone has cut up to the Ma12 layer (Mitamura et al. 1994) and that the Holocene sediments near the surface are bent (Headquarters for Earthquake Research Promotion of MEXT 2004). The dip angle of the fault has been estimated to be 65°–70° (Headquarters for Earthquake Research Promotion of MEXT 2004), as inferred from topographical and geological features (Huzita and Kasama 1982) and the results of seismic reflection surveys (e.g., Sugiyama and Sangawa 1996; Sugiyama, 1997; Sugiyama et al. 2001, 2003). Based on the difference in the thickness of the Quaternary strata between the eastern and western sides of the Uemachi fault, the displacement rate of the Uemachi fault zone from 1.2 Ma to 0.6 Ma has been estimated to be ~0.3 m/kyr (Uchiyama et al. 2001) while that from 0.6 Ma to 0.15 Ma has been estimated to be ~0.4 m/kyr (Headquarters for Earthquake Research Promotion of MEXT 2001).

Ishiyama (2003), based on the seismic reflection survey of Yoshikawa et al. (1987), interpreted the bending structure of the reflection section as a fault-propagation fold. Given that the Uemachi fault has a dip angle of ~40° and that there is 1 km of slip along the fault, he demonstrated, using the trishear fault-propagation fold model of Allmendinger (1998), that the sedimentary structure of the Ma-1 layer at the lowermost part of the Tanaka Formation can be reproduced. Furthermore, because the Osaka Group is thinner at the Uemachi Upland and thicker eastward, Ishiyama (2003) suggested that the Uemachi fault zone may be a thin-skinned thrust converging to a low-angle detachment in the upper crust (less than 5 km; Fig. 1). Sato et al. (2009) interpreted the seismic reflection results to indicate that the Uemachi and Ikoma fault zones inclined to the east and that their low-angle detachment faults converge at a depth of approximately 10 km, which is much deeper than the depth suggested by Ishiyama (2003). Iwasaki (2016) evaluated the dip angle of the Uemachi fault using PLAXIS®, a finite element method ground analysis software. Given a displacement of several meters along the fault with a dip angle of ~30° and with the fault tip located 1,000 m below the surface, the inclination angle of the axial plane of the Sakuragawa flexure (65°–70°) can be reproduced.

Ikoma fault zone

The Ikoma fault zone is a nearly N–S-striking fault system with a total length of 38 km. It is composed of the Ikoma fault, Katano fault, Hirakata fault, Taguchi fault, and Konda fault (Nakada et al. 1996a, 1996b; Okada et al. 1996; Shimokawa et al. 1997; Sugiyama et al. 1999; Okada and Togo 2000). At the surface, the Ikoma fault is located ~12 km east of the Uemachi fault (Fig. 2a) The Ikoma fault is an east-side-up reverse fault located near the boundary between the Osaka Plain and the Ikoma Mountains, which consist of rocks of the Ryoke belt (Horike et al. 1995; Nakata et al. 1996a, 1996b; Shimokawa et al. 1997). Active faults are inferred from the fact that the low fault cliff that cuts the lower terrace runs alongside the Ikoma fault approximately 0.5–1 km to its west, with the location of the fault estimated along the slope transformation line that transitions from the mountain slope to the fan; the inferred active faults are thought to be more important in the Holocene than the fault along the mountain slope (Okada and Yagi 2019; Okada and Togo 2000). Using seismic reflection surveys at the Ikoma fault, Shimokawa et al. (1997) found that the Ikoma fault is inclined to the east at a moderate angle (approximately 30°–40°) below a depth of 400 m. Furthermore, they estimated the mean vertical displacement rate of the Ikoma fault system to be 0.5–1 m/kyr, as inferred from a trench survey. Ishiyama (2003) suggested that the high uplift rate of the Ikoma fault zone indicates thick-skinned trajectories. He also pointed out that the downward projection of the Uemachi fault zone soles into the Ikoma fault in the shallower portion of the crust, suggesting that the northern Uemachi fault zone and the Ikoma fault zone comprise a larger system of a west-verging active fold and thrust belt that accommodates E–W contraction within the upper crust and that the Uemachi fault zone is a leading edge of the thrust belt.

Numerical simulations

Method

Numerical simulations of a two-dimensional visco-elasto-plastic body were performed using the I2ELVIS code of Gerya and Yuen (2003, 2007) and Gerya (2010) for MATLAB®, with the deviatoric strain rate, $\dot{\varepsilon}_{ij}$, including three components:

$$\dot{\varepsilon}_{ij} = \dot{\varepsilon}_{ij(\mathrm{viscous})} + \dot{\varepsilon}_{ij(\mathrm{elastic})} + \dot{\varepsilon}_{ij(\mathrm{plastic})}, \quad (1)$$

where

$$\dot{\varepsilon}_{ij(\mathrm{viscous})} = \frac{1}{2\eta}\sigma_{ij}, \quad (1a)$$

$$\dot{\varepsilon}_{ij(\mathrm{elastic})} = \frac{1}{2G}\frac{\mathrm{D}\sigma_{ij}}{\mathrm{D}t}, \quad (1b)$$

$$\dot{\varepsilon}_{ij(\mathrm{plastic})} = 0 \text{ for } \sigma_{\mathrm{II}} < \sigma_{\mathrm{yield}},$$

$$\dot{\varepsilon}_{ij(\mathrm{plastic})} = \chi\frac{\sigma_{ij}}{2\sigma_{\mathrm{II}}} \text{ for } \sigma_{\mathrm{II}} = \sigma_{\mathrm{yield}}, \quad (1c)$$

where η is the effective viscosity, G is the shear modulus, $\mathrm{D}\sigma_{ij}/\mathrm{D}t$ is the objective co-rotational time derivative of the deviatoric stress component σ_{ij}, σ_{yield} is the plastic yield strength for a given rock, $\sigma_{II} = (1/2\sigma_{ij}\sigma_{ij})^{1/2}$ is the second deviatoric stress invariant, and χ is the plastic multiplier.

To yield an effective rheology, the Mohr–Coulomb law was simplified using the yield stress, σ_{yield}, criterion and implemented using a "Mohr–Coulomb viscosity", η_{MC}, as follows:

$$\eta_{MC} = \sigma_{yield}/(2\dot{\varepsilon}_{II}), \qquad (2)$$

where $\dot{\varepsilon}_{II} = (1/2\dot{\varepsilon}_{ij}\dot{\varepsilon}_{ij})^{1/2}$ is the second invariant of the strain rate tensor. The yield stress or plastic strength, σ_{yield}, of a rock generally depends on the mean stress on the solids, P, such that

$$\sigma_{yield} = C + \sin(\varphi)P, \qquad (3)$$

where C is the cohesion (the residual strength at pressure P = 0) and φ is the effective internal friction angle. The effective viscosity, η, is then defined using the following criterion:

$$\eta = \eta_{creep}, \text{ when } 2\dot{\varepsilon}_{II}\eta_{creep} < \sigma_{yield}, \quad (4a)$$

$$\eta = \eta_{MC}, \text{ when } 2\dot{\varepsilon}_{II}\eta_{creep} > \sigma_{yield}, \quad (4b)$$

where η_{creep} is the creep viscosity. The creep viscosity, depending on the stress and temperature, is defined by the following power law equation:

$$\eta_{creep} = 1/2\dot{\varepsilon}_{II}^{(1-n)/n} A^{-1/n} \exp(E/nRT), \quad (5)$$

where A is the pre-exponential factor [/(Pa^n·s)], E is the activation energy [J/mol], n is the stress exponent, T is the temperature [K], and R is the gas constant (8.314 J/(K·mol)).

In this study, for the sediment layer and the upper basement layer (i.e., the upper crust), we used the flow law parameters for wet quartz ($A = 4.0 \times 10^{-11.2}$/(MPa^n·s), including the effect of a constant water fugacity of 4 MPa at T = 473 K and P = 200

MPa, n = 4, and E = 135 kJ/mol; Hirth et al. 2001), and for the lower basement layer (i.e., the lower crust), we used the flow law parameters for wet plagioclase (An_{60}) (A = 1.0 × $10^{-1.5}$/(MPa^n·s), n = 3, and E = 235 kJ/mol; Rybacki and Dresen 2004). To incorporate the effect of strain weakening, it is modelled as a linear decrease of friction angle and cohesion between accumulated strain of ε_{II} = 0 and ε_{II} = 1. These lower and upper thresholds of strain for weakening activation and completion are similar to those in previous numerical investigations of strain weakening of crustal rocks (e.g., Allken et al 2012; Ruh et al. 2014; Döhmann et al. 2019). At ε_{II} <1, the cohesion and friction angle change linearly from the initial cohesion (C_i) and initial friction angle (φ_i) to the weakened cohesion (C_w) and weakened friction angle (φ_w), respectively, whereas at ε_{II} ≥1, the cohesion and friction angle are constant at C_w and φ_w. For the basement layer, C_i and C_w are 10^7 Pa and 10^6 Pa, respectively, and φ_i and φ_w are 44° (initial friction coefficient μ_i = ~0.7) and 30° (weakened friction coefficient μ_w = ~0.5), respectively. The value of the weakened friction coefficient corresponds to the experimentally determined values for phyllosilicate minerals, i.e., mica and chlorite (Morrow et al. 2000; Ikari et al. 2011), and the value for damage zones of the San Andreas fault

(Carpenter et al. 2015). For the sediment layer, according to Gerya et al. (2009) and Gerya (2010), the physical properties of $C_i = C_w = 10^6$ Pa, $\varphi_i = 14°$, and $\varphi_w = 6°$ are employed for the calculation. The shear modulus G of the upper basement layer and the sediment layer is 1.0×10^{10} Pa, and that of the lower basement layer is 2.5×10^{10} Pa.

An energy conservation law that does not consider internal heat generation can be expressed as

$$\rho C_p \frac{\mathrm{D}T}{\mathrm{D}t} = \frac{\partial q_x}{\partial x} + \frac{\partial q_z}{\partial z}, \quad (6a)$$

$$q_x = k\frac{\partial T}{\partial x},\ q_z = k\frac{\partial T}{\partial z}, \quad (6b)$$

where q_x and q_z are the conductive heat fluxes in the horizontal and vertical directions, respectively, t is time [s], ρ is the local density depending on the composition, C_p is the specific heat at constant pressure, and k is the thermal conductivity. According to Gerya et al. (2009) and Gerya (2010), the values of C_p for the sedimentary layer and the basement are set to 1,000 J/(kg·K) and the value of k [W/(m·K)] for the sedimentary layer and the upper basement is $0.64 + 807/(T + 77)$ while that for the lower basement is

1.18 + 474/(T + 77).

Conservation of mass is approximated by the incompressible time-dependent two-dimensional continuity equation:

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_z}{\partial z} = 0, \qquad (7)$$

where v_x and v_z are the horizontal and vertical components of the velocity vector, respectively.

The two-dimensional Stokes equations for creeping flow are

$$\frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xz}}{\partial z} = \frac{\partial P}{\partial x}, \qquad (8a)$$

$$\frac{\partial \sigma_{zx}}{\partial x} + \frac{\partial \sigma_{zz}}{\partial z} = \frac{\partial P}{\partial z} - \rho g, \qquad (8b)$$

where g is the gravitational acceleration (9.81 m/s^2). In this study, the densities of the sediment, upper basement, and lower basement are 2,600 kg/m^3, 2,700 kg/m^3, and 2,800 kg/m^3, respectively. A weak layer above the lithosphere ("sticky air", $\eta = 10^{18}$ Pa s, ρ =

1 kg/m^3, k = 300 W/(m·K), C_p = 3.0 × 10^6 J/(kg·K)) provides a free-surface-like condition, which is essential to transform crust or sediment naturally (Gerya and Yuen 2003, 2007; Gerya 2010).

The calculation domain in the initial state was 80 km × 35 km (Fig. 4). The size of the basement layers was 80 km × 30 km. All models used a finite-difference with marker-in-cell technique and were conducted on a fully staggered rectangular Eulerian grid with 4,480,000 markers. The grid spacing was irregular and was initially applied as an 800 × 210 grid of 80 km × 21 km in the upper domain and as an 800 × 60 grid of 80 km × 14 km in the lower domain. The grid spacing was recalculated at each time step to accommodate the horizontal shortening described below. The initial position of each marker was evenly set and then given random noise using the MATLAB® random number generator and, in some cases, we changed the random seed to evaluate the effect of the initial position of the markers. The absolute time step was 10 kyr. The calculation time was set to 3 Myr according to the fission track age (2.71 million years ago; Itihara et al. 1984) of the volcanic ash layer at the bottom of the Osaka Group.

We applied free-slip conditions at all boundaries and constant leftward and downward velocities at the right and bottom boundaries, respectively. Temperatures at the top and bottom boundaries were constant, i.e., 0 °C and 750 °C, respectively, and the right and left boundaries were adiabatic. An initial geothermal gradient of 25 °C/km (e.g., Okubo et al. 2005) was applied to the basement layer, whereas there was no temperature gradient in the sticky air layer (i.e., the temperature at the top of the basement layer was also 0 °C). Following the horizontal shortening strain rate (16–26 × 10^{-9}/yr) for central Japan of Wesnousky et al. (1982), the horizontal shortening rate was set to 2 m/kyr (Fig. 4). Therefore, at the end of the calculation (3 Myr), the 80-km-wide crust was shortened by 6 km. To conserve mass in the calculation domain, a changing downward velocity (from 0.88 m/kyr at 0 Myr to 1.02 m/kyr at 3 Myr) was applied at the bottom boundary.

To model preexisting faults, strained zones with a strain, ε_{II}, of 1, having the initial C_w and φ_w, and a yield stress or plastic strength weaker than the surrounding rock mass were applied in the basement layer as high-strained weak zones. In this study, we refer to such high-strained weak zones as “faults” or “fault zones”. Even though the width of a preexisting fault zone was set to 250 m, that is, 2.5 times wider than the grid space, the width of the zone follows the scaling law of the linear relationship between the width of the fault process zone and the fault length with a proportionality constant on the order of 10^{-2} (Vermilye and Scholz 1998).

To model sedimentation processes, we introduced an imposed sea level change that is consistent with the geologically observed sedimentation rate at the location of the OD-1 core site. At the initial state, the sea level was set to the boundary between the sticky air and the rock layer. At each time step, after all markers were moved according to the calculated velocity field, we changed the sea level and markers of the sticky air below the sea level were replaced with markers of the sedimentary layer. Surface erosion was not considered in this study.

We considered three types of cases with the simulation model: (1) cases without any preexisting fault zone; (2) cases with a single fault zone cutting the upper 12 km of the basement layer; and (3) cases with two preexisting fault zones corresponding to the Uemachi and Ikoma fault zones.

Results and discussion

Cases with no preexisting fault zone

In these cases, we did not consider sedimentation processes. In the cases without any preexisting fault, under the compressive stress field, the upper and lower basement layers deform homogeneously prior to 0.4–0.6 Myr, whereas after 0.4–0.6 Myr, in the upper basement layer, the strain is concentrated into narrow zones and localized high-strained zones, i.e., newly formed thrust faults that develop as a result of compression (Fig. 5). These cases demonstrate that simulation results with different random seeds at 3 Myr differ somewhat with respect to the dip directions of the newly formed thrust faults. The thrust faults grow downward from the surface because values of the yield stress σ_{yield} of a rock given by Eq. (3) are lower at shallower depths. They develop from the surface to the frictional–viscous transition zone, where the strain is accommodated primarily by frictional deformation with weakened friction (μ_w) relative to the viscous deformation (inset in Fig. 4). The dip angles of the faults are ~30° near the surface but bend gently near the frictional–viscous transition depth. These fault geometries (i.e., dip angles of the faults) are not explicitly “determined” by the code but form spontaneously during the propagation of the high-strained zone involving markers for which the yielding condition given by Eq. (3) is satisfied locally. When increasing the amount of horizontal shortening, the thickness of the early formed faults widens and becomes ~300–500 m at 3 Myr; however, the depth of the fault tips does not change significantly after ~2 Myr. In the lower basement layer, the strain is distributed to form large-scale folds instead of localized high-strained zones. The displacement along the faults leads to relative subsidence on the lower side of the faults or uplifting on the upper side. In general, the amount of vertical displacement decreases with distance from the faults. The blocks between faults with the same dip direction tilt antithetically (Fig. 5a), whereas the blocks between faults with opposite dip directions rise or sink vertically (Fig. 5b).

Cases with a single preexisting fault zone

In these cases, we did not consider sedimentation processes. In the cases with a single fault zone, we set the fault to have different dip angles (i.e., 15°, 30°, 45°, and 60°) at the surface location of x = 23 km. Based on the simulation results for the cases with no preexisting fault, because the local displacements in the strained zone in the deepest part of the upper basement layer and the lower basement layer are negligible, we considered the faults as only preexisting above 12 km. In this case, the effect of changing the random seed is smaller than in the cases without a preexisting fault and we only show the results with the random seed = 1 in Fig. 6. In all the cases, at the initial stage, strain localization occurs in the preexisting fault zone. The zone widens and becomes a source for newly generated spray fault zones. Displacement occurs along the preexisting faults in the upper basement layer for dip angles of 15°–45° but not 60° (Fig. 6). When the dip angle of the preexisting fault is 15° or 45°, the displacement along the preexisting fault is relatively small and new high-strained zones with dip angles of ~30° develop. The preexisting fault steepens by ~2° because of the rotation of the fault during horizontal shortening. When the dip angle of the preexisting fault is 30°, which is suitable for horizontal compression, the displacement along the fault is significant and the fault ultimately steepens by ~10° as a result of the horizontal shortening and the tilting of the upper basement layer. After the preexisting fault becomes steeper, new faults with dip angles of ~30° are formed. Conversely, when the dip angle of the preexisting fault is 60°, many new faults with dip angles of ~30° are formed and evenly displaced; in this case, the preexisting fault ultimately steepens by ~2°.

When the faults contain large amounts of clay minerals, the experimentally derived friction coefficient for the fault zone decreases to ~0.1–0.2 (Takahashi et al. 2007; Ikari et al. 2011). In fact, the frictional strength estimated for the weakest section of the San Andreas fault is ~0.1 but increases abruptly to a value of ~0.4–0.5 in the host sedimentary rocks (Carpenter et al. 2015). We performed a simulation for a single preexisting fault zone with lower friction values of μ_w = ~0.4 and ~0.3 (φ_w = 24° and 17°, respectively). Figure 7 shows the results of the effect of the different weakened friction values on the activity of a preexisting fault with a dip angle of 60°. In this case, the preexisting fault, which is not favorably oriented to the horizontal compressional stress field, is active under the condition of $\mu_w < 0.4$.

Cases with two preexisting fault zones

In these cases, we consider the sedimentation process. We set two faults, whose locations at the surface were x = 23 km (i.e., the western fault) and x = 40 km (i.e., the eastern fault). The western and eastern faults correspond to the Uemachi and Ikoma faults, respectively, which are separated by ~14 km at the surface in the final state. At the location equivalent to x = 18 km in the initial state, corresponding to the OD-1 core site, which is ~5 km west of the Uemachi fault, we set the mean sedimentation rate to deposit 900 m of the Miyakojima Formation and 650 m of the Tanaka and Namba formations (i.e., 0.5 m/kyr from 0 to 1.8 Myr, 0.6 m/kyr from 1.8 Myr to 2.6 Myr, and 0.4 m/kyr from 2.6 Myr to 3.0 Myr). Because Quaternary sediments are basically absent in the Ikoma Mountains (with a highest elevation of 642 m), the Ikoma Mountains may have been separated from the marine waters at 3 million years ago (Ma). However, because the elevation of the Ikoma Mountains at 3 Ma is unknown, the elevation of the Ikoma Mountains was assumed to be 0 m at 3 Ma (at 0 Myr in this study). According to the results for the cases with a single fault zone (Fig. 6), the optimal dip angle for the resolved shear stress is ~30° and the dip angles of the western and eastern faults should also be ~30°. Based on the geological interpretation, because the vertical displacement rate related to the Ikoma fault has been estimated to be much higher than that related to the Uemachi fault (Table 1), the dip angle of the western fault was set to differ slightly from the optimal value for the resolved shear stress, or the friction coefficients for the eastern fault were set to be lower than those for the western fault. Because previous studies have generally suggested that the dip angle of the Uemachi fault is steeper than that of the Ikoma fault (Fig. 1) and because the dip angle of the Uemachi fault was recently estimated to be ~40° (Kato and Ueda 2019), we applied dip angles of 40° and 30° to the western and eastern faults, respectively (Fig. 8a). In this case, the downward extension of the western fault nearly coincides with the epicenter of the 2018 northern Osaka earthquake.

We compared the simulation results and geologic structures proposed by Uchiyama et al. (2001, Fig. 3) based on the following points: (1) the large-scale variation in the thickness of the upper and uppermost sedimentary layers, corresponding to the Tanaka and Namba formations, in the area in-between the western and eastern faults, i.e., in east Osaka, especially at the location of the OD-9 core site (~2 km east from the Uemachi fault at the final state) in the Uemachi Upland and at the location of the OD-3 core site (~9 km east from the Uemachi fault at the final state) near the Onchi River, and (2) the vertical displacement rates after 1.8 Myr, corresponding to the sedimentation period of the Tanaka and Namba formations, estimated by the vertical differences between the highest point of the hanging wall basement and the lowest point of the foot wall basement near the western and eastern faults.

The displacement along the western and eastern faults leads to relative subsidence on the lower side of the faults so that sediment layers form in the relative subsidence regions (Fig. 8a). The surface of the basement slopes to the east as a whole, and an eastward thickening of the sedimentary layers throughout the Osaka Plain is observed. These layers result from the displacement along the eastern fault being much larger than that along the western fault (Table 1). The region west of the eastern fault is tilted eastward as a whole, even though displacement along the western fault and a difference in the sedimentary layer thickness between the areas on both sides of the western fault are observed. As shown in Fig. 8b, the vertical displacement rates related to the western and eastern faults increase monotonically prior to ~0.5 Myr. At the initial stage (<~0.5 Myr) of the simulation, the deformation does not concentrate along the entire fault; the deformation front moves downward along the fault, and then the entire fault deforms after ~0.5 Myr. Therefore, the monotonical increase in the displacement rate at the initial stage of the simulation results from the increase in the deforming length along the faults. The vertical displacement rate related to the eastern fault increases to ~1 m/kyr at ~1 Myr and becomes stable, whereas that along the western fault increases to ~0.3 m/kyr at ~0.5 Myr and then decreases. This implies that the deformation of the eastern fault is predominant after ~0.5 Myr, even though the deformation is more partitioned in the eastern fault, and then the western and eastern faults deform stably after ~1 Myr. The mean vertical displacement rates along the western and eastern faults for the last 1 Myr are ~0.07 m/kyr and ~0.8 m/kyr, respectively (Fig. 8b and Table 1). The mean displacement rate along the western fault is much slower than the inferred displacement rate along the Uemachi fault (~0.3–0.4 mm/yr; Uchiyama et al. 2001; Headquarters for Earthquake Research Promotion of MEXT 2001), whereas that along the eastern fault corresponds to the inferred displacement rate along the Ikoma fault (0.5–1 m/kyr; Shimokawa et al. 1997).

The depth of the basement surface is approximately 1,400 m at the location of the OD-9 core site in the Uemachi Upland and approximately 1,900 m at the location of the OD-3 core site near the Onchi River (Table 1). Both results are deeper than the geologically inferred values. The thickness of the sediment layer, corresponding to the

Tanaka and Namba formations, increases eastward; it is ~690-m thick near the Uemachi Upland and ~870-m thick near the Onchi River. Both results are thicker than the geologically inferred value. The sedimentation rate near the Onchi River is 0.7 m/kyr for the period of 1.8–2.4 Myr, corresponding to the lower Tanaka Formation, whereas that near the Uemachi Upland is 0.6 m/kyr for the same period (Fig. 8c); the former is consistent with geologically inferred values, but the latter is faster than the geologically inferred values (Table 1). However, although the sedimentation rate near the Onchi River hardly changes from 1.8 Myr to 2.4 Myr, the geologically inferred sedimentation rate decreases from 1.2 Ma to 0.6 Ma (Uchiyama et al. 2001). Consequently, in this case, the thicknesses of the upper and uppermost sedimentary layers corresponding to the lower Tanaka Formation and the upper Tanaka and Namba formations near the Uemachi Upland differ from the geological interpretation by ~160 m and ~280 m, respectively. The mean displacement rate along the western fault is smaller than that inferred from the geological observations. To resolve this discrepancy, the displacement rate along the western fault needs to be enhanced by decreasing the dip angle of the fault at least partially and/or by decreasing the friction coefficients. Accordingly, we evaluated gentler (30°) or curved western fault cases (Fig. 9). In the curved fault cases, the dip angle of the western fault is 40° at the surface and gradually decreases with depth. The dip angle of the deeper part (>7 km) of the western fault is constant and is 20° in the curved fault case and 15° in the more curved fault case.

Figure 9a illustrates the results of the case with the gentler dip angle (30°) for the western fault (the gentler Uemachi case). The displacement rate along the western fault is much higher than that in the case with the dip angle of the western fault set to 40°, whereas the displacement rate along the eastern fault is lower, implying that more deformation is partitioned to the western fault, decreasing the displacement rate along the eastern fault. Although the dip angle of the both faults is the same, the displacement rate along the western fault is lower than that along the eastern fault. It may be resulted from the effects of viscous flow near the bottom of the lower basement layer; when the adjacent, parallel preexisting faults are eastward dipping, the downward extension of the easternmost fault (i.e., the eastern fault in this case) develops as a lower crustal high-strained zone connecting to the basal viscous flow and then the strain localizes into the eastern fault (Additional file 1: Fig. S1). The eastern fault and its downward extension act as a bounding fault between the less deformed hanging wall and footwall blocks. A local uplift of the high-strained triangular zone of the sedimentary layers in between conjugate fault set propagated from the tip of the western fault is observed. The formation of the conjugate fault set may be resulted from the increase in the displacement rate and/or the decrease in the dip angle of the western fault. Based on a comparison with the geological interpretation, this gentler Uemachi case is more suitable; however, the difference between the geological interpretation and the simulation results is still large, especially the thickness of the uppermost sedimentary layer in the Uemachi Upland and the basement depth near the Onchi River (Table 1).

The results of the curved fault cases (the curved and more curved Uemachi cases) are illustrated in Fig. 9b–e. Nearly horizontal sedimentary layers form in west Osaka, whereas an eastward thickening of the sedimentary layers is observed in east Osaka. The surface of the basement in east Osaka slopes to the east, whereas that in west Osaka is nearly horizontal. These slopes result from the clockwise rotation of the east Osaka block arising from the displacement along the synthetic eastern and western faults. The degree of tilting toward the east of the east Osaka block in the curved western fault cases is larger than that in the non-curved cases; however, the difference in the degree of tilting between the curved and more curved cases is not significant. The thickness of the sediment layer in east Osaka in the more curved western fault case is larger than that in the less curved case (Table 1). Based on a comparison with the geological interpretation, the less curved case is the more suitable geometry for the present numerical model. In the curved fault case, the numerical results are consistent with the geological interpretation, even though the thickness of the uppermost sedimentary layer near the Uemachi Upland differs from the geological interpretation by ~170 m. This difference may be reduced by a higher vertical displacement rate along a fine-tuned, optimally curved western fault.

The difference in the thickness of the uppermost sedimentary layer between the simulation results and the geological interpretation may also be due to the effect of the changing sedimentation rate for the uppermost sedimentary layer. In this study, we set the mean sedimentation rate at the OD-1 core site to 0.4 m/kyr from 2.6 Myr to 3.0 Myr, even though the sedimentation rate from ~0.4 Ma to ~0.05 Ma actually decreased from 0.5 m/kyr to 0.2 m/kyr (Uchiyama et al. 2001). Furthermore, in the present simulation models, the subsidence rate in west Osaka decreases with distance from the western fault, even though, based on gravity and seismic reflection data (e.g., Nakagawa et al. 1996b; Iwabuchi 2000), the top of the basement deepens toward the west and the deepest part of the Osaka Bay is located ~27 km from the Uemachi fault, where the NNE–SSW-trending Osaka-wan fault is located. The Osaka-wan fault is an ESE-side-down reverse fault, having a mean vertical displacement rate of 0.5–0.6 m/kyr after 1 Ma (Yokokura et al. 1998). If the subsidence rate at the OD-1 site related to the activity of the Osaka-wan fault is higher than that used in the present models, the sedimentation rate in east Osaka will be lower because the rising rate of the sea level will be reduced. Further studies need to evaluate the effect of changing the sedimentation rate and vertical displacement along the Osaka-wan fault on the development of the geologic structures of the upper and uppermost sedimentary layers.

Evaluation of previously proposed fault geometry models

As previously mentioned, geometries of the deeper parts of the Uemachi and Ikoma faults have been proposed by Ishiyama (2003) and Sato et al. (2009). In this section, we evaluate their proposed Uemachi and Ikoma fault geometries.

Following Ishiyama (2003), we applied two faults initially in the basement layer with the western and eastern faults corresponding to the Uemachi and Ikoma faults, respectively (Fig. 10a). The dip angles of the western and eastern faults near the surface were 40° and 55°, respectively. The surface positions of the two faults were the same as in the cases with two preexisting fault zones. The eastern fault was linear, whereas the western fault had a downward convex curving part at a depth of ≥1.5 km, became nearly horizontal (2°) at a depth of 3.5 km, and then converged to the eastern fault at a depth of ~4 km. Because a preexisting weak fault with a dip angle of ~60° is active when the weakened friction coefficient (μ_w) is less than ~0.3, we performed numerical simulations for this case with μ_w = 0.3. In this case, even though the eastern fault (dip angle of 55°) was active, displacement related to the western fault (dip angle of 40°) is not significant and does not occur along the deeper (>3.5 km) and horizontal (<2°) parts of the fault. A newly formed fault with a dip angle of ~30° propagates from the shallower and steeper parts of the western fault, and displacement related to the Uemachi fault occurs along this newly formed fault. This result implies that a thin-skinned thrust converging to a low-angle detachment in the upper crust would not be active. The displacement along the eastern fault is significant but a large-scale downward fold, i.e., synform, is formed in the area east to the eastern fault, leading to the development of a sedimentary basin.

According to Sato et al. (2009), the western fault has a near surface dip angle of 50°, transitions to gently dipping at a depth of ≤7 km, and then converges to the eastern fault at a depth of ~9 km (Fig. 10b) with a dip angle of 4°. The dip angle of the eastern fault near the surface was set to 45°, and the eastern fault transitioned to gently dipping at a depth of 5 km. Below the converged depth, the dip angle of the eastern fault was set to 20°. The surface positions of the two faults were the same as in the cases with two preexisting fault zones. We performed numerical simulations for this case with $\mu_w = 0.4$. The simulation results are similar to the curved Uemachi case (Figs. 9b and 10b), which correspond roughly to the geological interpretation (Table 1). The displacements along the lower (>9 km) and horizontal (<4°) parts of the Uemachi fault are not significant, and the newly formed high-strained zones with a dip angle of ~30° propagate from the upper bound of the lower and horizontal parts (Additional file 1: Fig. S2). Therefore, the actual fault geometry of this model is similar to our curved Uemachi case at shallower depths (<~10 km).

Concluding remarks

Based on a simple numerical analysis, we can conclude that, when the friction coefficient of the faults is approximately 0.5, the dip angles of the Uemachi and Ikoma faults near the surface are ~30°–40° and the Uemachi fault has a downward convex curve at the bottom of the seismogenic zone but does not converge to the Ikoma fault. This implies that the downward extension of the Uemachi fault zone nearly coincides with the epicenter of the 2018 northern Osaka earthquake.

Abbreviations

MEXT: Ministry of Education, Culture, Sports, Science and Technology of Japan.

Author's contributions

HN, TO and KI performed the numerical simulations and evaluated the model results.

MM evaluated the geological features in the Osaka Plain to constrain the model results. HN and TO drafted mainly the manuscript, and all authors read and approved the final manuscript.

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Competing interests

The authors declare that they have no competing interests.

Availability of data and materials

All data generated or analyzed during this study are available from the corresponding author on reasonable request.

Consent for publication

Not applicable.

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Figure legends

Fig. 1 Summary of the inferred geometries of the Uemachi and Ikoma faults in the Osaka Plain illustrated on a schematic geological cross section (modified partially from Ishiyama 2003). Here, references 1, 2, 3, 4, 5 and 6 correspond to Headquarters for Earthquake Research Promotion of MEXT (2001, 2004), Ishiyama (2003), Director General for Disaster Management (2006), Sato et al. (2009), Iwasaki (2016) and Kato and Ueda (2016), respectively. The epicenter of the 2018 northern Osaka earthquake (EQ) proposed by Kato and Ueda (2019), located 12 km beneath the surface position 14 km from the Uemachi fault, is indicated by a star.

Fig. 2 **a** Locations of the drilling sites on the Osaka Plain referred to in this study. Faults and flexures of the Uemachi fault zone include the Uemachi fault (Um1), Sakuragawa flexure (Um2), Nagai fault (Um3), and Suminoe flexure (Um4), and those of the Ikoma fault zone include the Katano fault (Ik1), Hirakata fault (Ik2), and Ikoma fault (Ik3) (Headquarters for Earthquake Research Promotion of MEXT 2001, 2004). The inset indicates the location of the study area. **b** Columnar sections of the OD-1, TS, YU, OD-9, OT, OD-2, HA, and OD-3 drilling cores (partially modified from Uchiyama et al. (2001)).

Fig. 3 Schematic E–W geological section through the Osaka Plain (Uchiyama et al. 2001).

Fig. 4 Model geometry and boundary conditions. The inset shows the crustal strength profile at the initial state in the numerical model. $\dot{\varepsilon}$, μ_i, and μ_w are the strain rate, the initial friction coefficient, and the weakened friction coefficient, respectively. A changing downward velocity (from 0.88 m/kyr at 0 Myr to 1.02 m/kyr at 3 Myr) is applied at the bottom boundary.

Fig. 5 Results at 3 Myr for the numerical simulations with no preexisting fault zone. The newly formed faults (markers with a cumulative strain larger than 1) are illustrated in dark blue. **a** Case with the random seed number = 1. **b** Case with the random seed number = 3.

Fig. 6 Results at 3 Myr for numerical simulations with a single preexisting fault zone. The preexisting fault (PF) and the newly formed faults (markers with a cumulative strain larger than 1) are illustrated in dark blue. The weakened friction coefficient μ_w is ~0.5 (weakened friction angle ϕ_w = 30°). **a** Case with a dip angle of 15°. **b** Case with a dip angle of 30°. **c** Case with a dip angle of 45°. **d** Case with a dip angle of 60°.

Fig. 7 Effect of different weakened friction coefficients μ_w on the activity of a preexisting fault with a dip angle of 60°. **a** Case with μ_w = ~0.4 (weakened friction angle ϕ_w = 24°). **b** Case with μ_w = ~0.3 (ϕ_w = 17°). The preexisting fault (PF) and newly formed faults (markers with a cumulative strain larger than 1) are illustrated in dark blue.

Fig. 8 Results of the numerical simulation for a case with two preexisting fault zones with dip angles of 40° and 30°, corresponding to the Uemachi and Ikoma faults in the Osaka Plain, respectively. **a** The initial model geometry and time evolution. The epicenter of the 2018 northern Osaka earthquake (Kato and Ueda 2019), located 12 km beneath the surface position 14 km from the Uemachi fault, is indicated by a star. The lower panels show a close up view of a 60 km × 10 km domain including the western and eastern weak zones. The preexisting faults and newly formed faults (markers with a cumulative strain larger than 1) are illustrated in dark blue. **b** Time evolution of the vertical displacement rates along the western and eastern faults. **c** Time evolution of the sedimentation rates near the Uemachi Upland (i.e., at the location of the OD-9 core site) and the Onchi River (i.e., at the location of the OD-3 core site). The imposed sedimentation rates at the location of the OD-1 core site are also shown.

Fig. 9 Effect of changing the dip angle of the western fault (i.e., the Uemachi fault) on the development of the geologic structures. **a** The initial geometry (upper), final geometry (middle), and close up of the final geometry (lower) of the linear case with a dip angle of 30°. **b** The initial geometry (upper), final geometry (middle), and close up of the final geometry (lower) of the curved case. **c** The initial geometry (upper), final geometry (middle), and close up of the final geometry (lower) of the more curved case. The preexisting faults and newly formed faults (markers with a cumulative strain larger than 1) are illustrated in dark blue. The epicenter of the 2018 northern Osaka earthquake (Kato and Ueda 2019), located 12 km beneath the surface position 14 km from the Uemachi fault, is indicated by a star in the panels showing the result at 3 Myr. **d** Time evolution of the vertical displacement rates along the western and eastern faults for the curved case. **e** Time evolution of the sedimentation rates near the Uemachi Upland (i.e., at the location of the OD-9 core site) and the Onchi River (i.e., at the location of the OD-3 core site) for the curved case. The imposed sedimentation rates at the location of the OD-1 core site are also shown.

Fig. 10 Results of the numerical simulations for previously proposed fault geometries. **a** The initial geometry (upper), final geometry (middle), and close up of the final geometry (lower) for the fault geometry of Ishiyama (2003) with a weakened friction coefficient μ_w of ~0.3 (weakened friction angle ϕ_w = 17°). **b** The initial geometry (upper), final geometry (middle), and close up of the final geometry (lower) for the fault geometry of Sato et al. (2009) with μ_w = ~0.4 (ϕ_w = 24°). The preexisting faults and newly formed faults (markers with a cumulative strain larger than 1) are illustrated in dark blue. UmF and IkF represent the surface positions of the Uemachi and Ikoma faults, respectively.