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# Distribution of shortening landward and oceanward of the eastern Nankai trough due to the Izu–Ogasawara ridge collision

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**Abstract**

The convergence of the Philippine Sea and Eurasia plates is accommodated primarily by plate subduction along the Nankai trough west of about 137.3°E, but additionally by widely distributed onshore and offshore deformation to the east. This zone, with a total dimension of 200 km (parallel to the Nankai trough) by 350 km, has developed as one highly coupled region since the collision of the Izu peninsula with mainland Japan about 1–2 m.y. ago. The onshore faulting in central Japan is consistent with a WNW–ESE direction of maximum horizontal stress, while the oceanward deformation is driven by diffuse NW–SE shortening of the Izu–Ogasawara volcanic ridge which produces a kinematic discontinuity with respect to the Philippine Sea oceanic crust to its west. The strain release associated with the 1891 Nobi earthquake, and the onshore deformation as a whole, suggests that the Philippine Sea–Eurasia convergence is partially accommodated by onshore shortening between the Tsuruga Bay–Ise Bay Tectonic Line and the Izu peninsula.

## 1. Introduction

The recent tectonism of Japan is dominated by the westward subduction of the Pacific plate beneath northeast Japan and the northwestward subduction of the Philippine Sea plate below southwest Japan. The maximum horizontal stress is approximately parallel to the Pacific subduction vector (close to E–W) in northern Honshu and in central Japan north of the Median Tectonic Line (MTL, see Fig. 1); it is perpendicular to the Nankai trough (parallel to the Philippines Sea subduction vector) south of the MTL [1–3].

Fig. 1 shows that central Japan is a region of maximum topographic relief and, as noted by Wesnousky et al. [1], this increased relief appears to be directly related to the intensity of present intraplate deformation, between the ISTL (Itoigawa–Shizuoka Tectonic Line) and the TITL (Tsuruga Bay–Ise Bay Tectonic Line, see Fig. 3) north of the MTL. There, the axis of maximum shortening is E–W to WNW–ESE in spite of the proximity of the north-northwestward Philippine Sea plate subduction.

A simple quantitative discussion of the topographic relief helps to evaluate the excess of deformation in this area with respect to adjacent zones. Let us consider a series of great circle topographic profiles perpendicular to a small circle which follows

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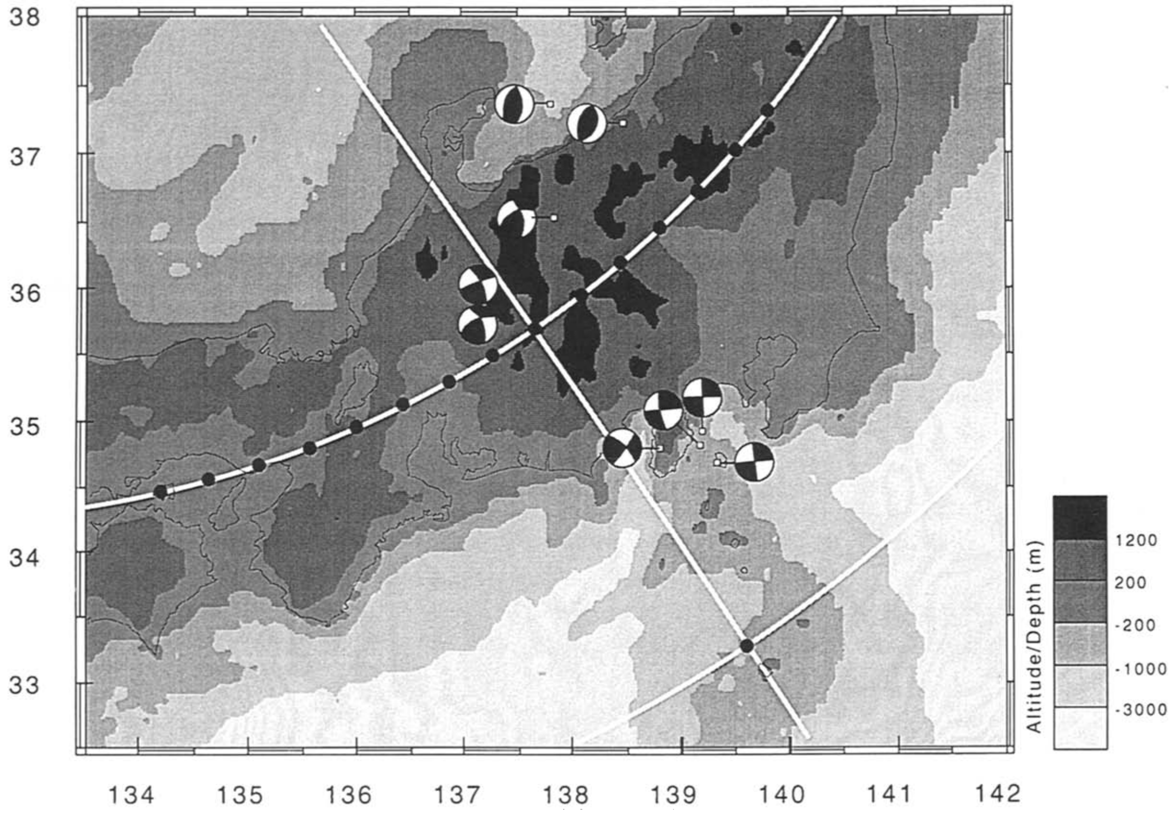
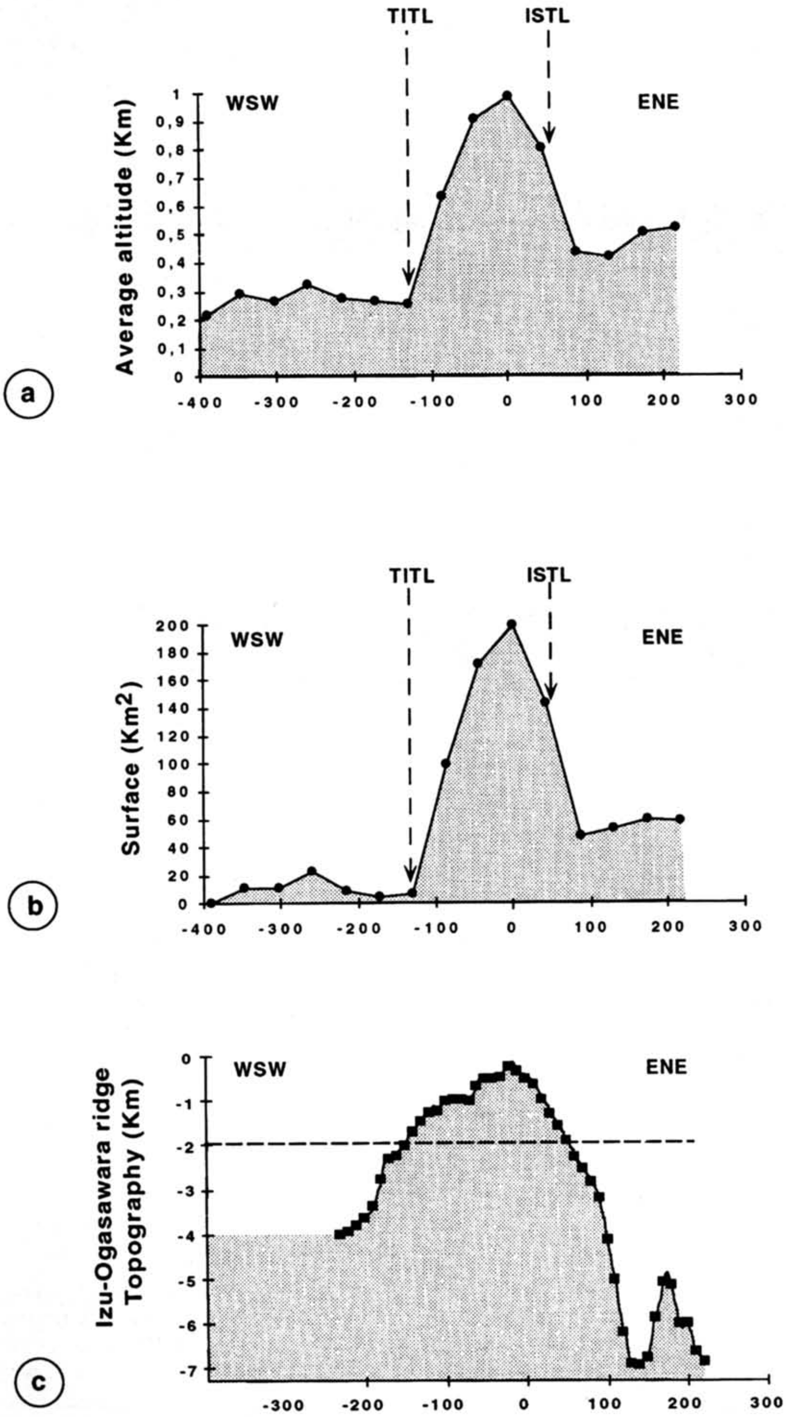


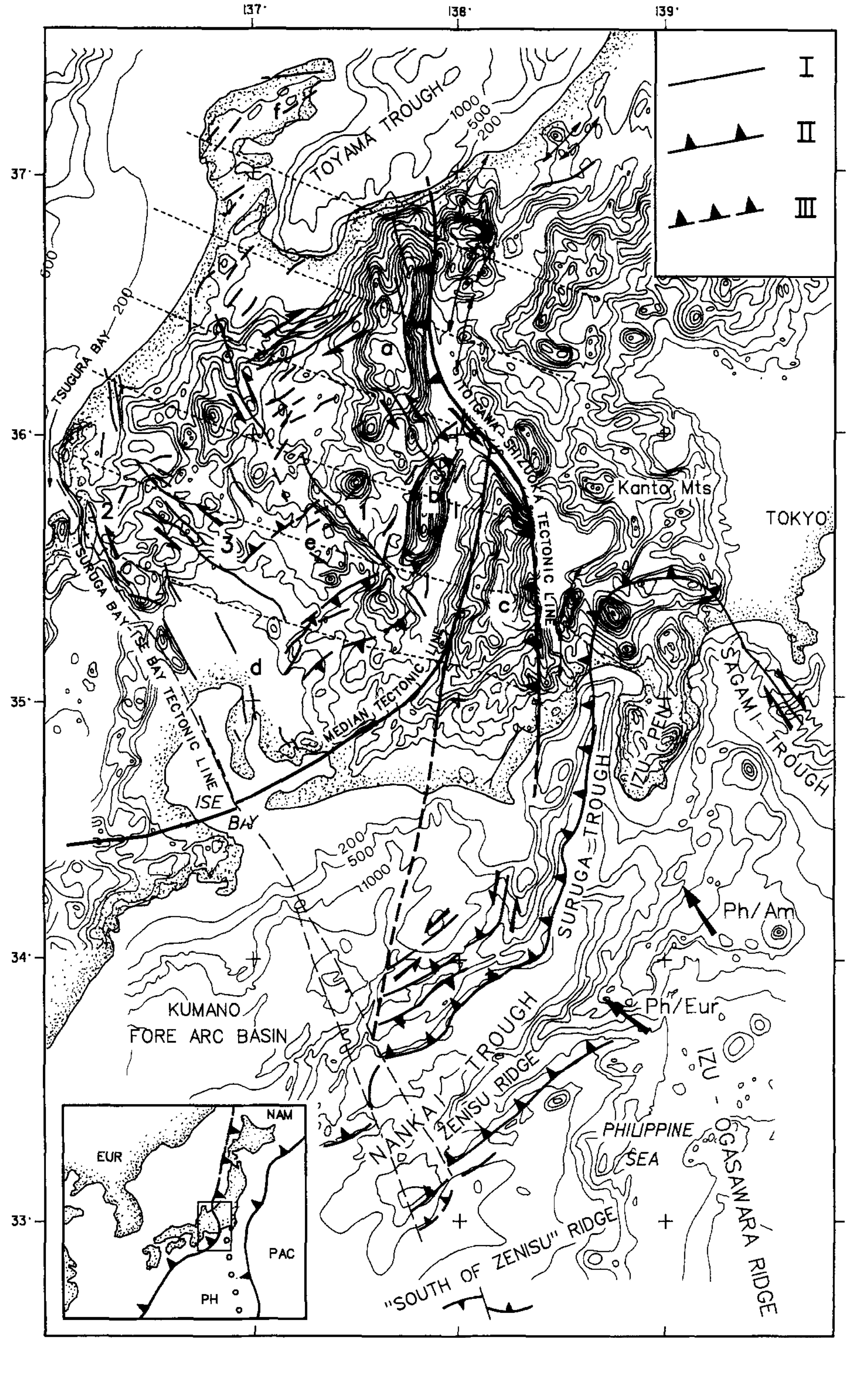

Fig. 1. Topographic map of central Japan with a few focal mechanisms for shallow earthquakes to represent the deformation. The topography is obtained from the ETOPO5 database. The small circle following the median line of the Japanese islands is the base line for Fig. 2a,b. The intersections of great circles perpendicular to it used in Fig. 2a and 2b are shown with dots. The great circle along which the topography is at a maximum is shown in full. Finally, the location of the topographic profile across the Izu-Ogasawara ridge shown in Fig. 2c is given.

approximately the median part of the Japanese islands from the western side of Fig. 1 to the east side. The average altitude of these profiles is plotted against distance in Fig. 2a. Altitude is quite constant at 270 m to the west of central Japan, then increases rapidly to about 1000 m above central Japan, and decreases to 400–500 m over northern Honshu. Taking the altitude at the western end (218 m) as the value, we then plot on Fig. 2b the surface above it for each of these profiles. If the topographic relief above this reference level is the result of tectonic shortening, which is a crude but reasonable first approximation, Fig. 2b shows the relative amount of shortening and confirms that it is at a maximum in central Japan over a width of 200 km.

East of the ISTL, the collision of the Izu–Ogasawara island arc with mainland Japan results in a sharp bend in the Nankai trough through the

Fig. 2. (a) Average altitudes of topographic profiles identified in Fig. 1 along distance from west to east. The profiles are taken from the shore of the Sea of Japan to that of the Philippine Sea or Pacific Ocean. Note the 200 km wide zone of high average topography coinciding with central Japan. (b) Surface (km$^2$) above the altitude of the westernmost profile (218 m) for the same topographic profiles as those of (a). The excess surface is approximately proportional to the average topography. (c) Topographic profile across the Izu–Ogasawara ridge. The location of the profile is identified in Fig. 1. The 2000 m depth level is identified. The significance of the similarity of this profile to the profile of (b) is discussed in the text.


TITL
ISTL
WSW
ENE
Average altitude (Km)
1
0,9
0,8
0,7
0,6
0,5
0,4
0,3
0,2
0,1
0
-400 -300 -200 -100 0 100 200 300
a
TITL
ISTL
WSW
ENE
Surface (Km²)
200
180
160
140
120
100
80
60
40
20
0
-400 -300 -200 -100 0 100 200 300
b
WSW
ENE
Izu-Ogasawara ridge Topography (Km)
0
-1
-2
-3
-4
-5
-6
-7
-300 -200 -100 0 100 200 300
c


137°
138°
139°
37°
36°
35°
34°
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I
II
III
TOYAMA TROUGH
1000
500
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TSUGURA BAY
ITOIGAWA SHIZUOKA TECTONIC LINE
MEDIAN TECTONIC LINE
SURUGA BAY ISE BAY TECTONIC LINE
ISE BAY
Kanto Mts
TOKYO
SAGAMI TROUGH
IZU PEN.
SURUGA TROUGH
Ph/Am
Ph/Eur
KUMANO FORE ARC BASIN
NANKAI TROUGH
ZENISU RIDGE
PHILIPPINE SEA
IZU - OGASAWARA RIDGE
"SOUTH OF ZENISU" RIDGE
EUR
NAM
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PH

Suruga and Sagami troughs around the Izu peninsula. Several authors have attributed the peculiarities of the central Japan deformation field to this collision (e.g., [4,5,1,6,7]).

To the southeast of the Nankai trough, active N–S to NW–SE shortening is manifested by the formation of intra-oceanic thrusts [8–11]. The intra-oceanic deformation is bounded to the east by the Izu–Ogasawara volcanic ridge and to the west by a line which marks the approximate seaward extent of the TITL. The Izu peninsula in the collision zone is being shortened in a NW–SE direction at a rate of at least 1 mm/yr [1]. Seismic activity shows that this shortening extends within the Izu–Ogasawara ridge to the south [2,12]. The whole northern part of the ridge is thus a zone of diffuse shortening, presumably because its crust is thick, hot and consequently weak. This area is an active volcanic ridge with a high heat flow of about 200 mW $m^{-2}$ near its axis and consequently the elastic thickness there is about 4× lower than within the adjacent oceanic floor [13]. The transition from diffuse shortening within the volcanic ridge to the concentrated subduction along the Nankai trough to the west produces a kinematic discontinuity along the ridge flank which is partly absorbed by intra-oceanic shortening [2,9–11].

Fig. 2c is a topographic profile across the Izu–Ogasawara ridge. It is remarkable that the distribution in adjacent central Japan of excess topographic relief (as shown in Fig. 2b), and presumably shortening, is proportional to the altitude of the Izu–Ogasawara ridge above a base level with a depth of 2000 m (see Fig. 2c). This coincidence is best explained if there is a causal relationship between the shortening of central Japan and the collision with it of the part of the Izu–Ogasawara ridge above the 2000 m depth.

In this paper, we propose that central Japan (i.e., both the onshore and offshore parts) forms a 200 km wide and 350 km long zone of strain distribution straddling the Nankai trough. We further propose that the distributed shortening in this zone is related to the diffuse shortening within the hot and weak collision zone to the southeast of central Japan.

## 2. Plate kinematics

The 305–310° direction of subduction of the Philippine Sea plate below the eastern Nankai trough was well established by the great Tonankai earthquake [14,15], which ruptured the eastern section of the trough (see Fig. 3). In the absence of space geodetic measurements, however, the magnitude of the velocity, on the other hand, could only be estimated indirectly. As a consequence, estimates vary between 2 and 4 cm/yr [16,2,17]. Recent VLBI estimates indicate that the 4 cm/yr value is more likely [18] and this is the solution proposed by Seno et al. [19].

Yoshioka et al. [20] have argued on the basis of geodetic modelling that Japan east of the ISTL belongs to the American plate, which would result in a 330° direction of subduction instead of 310°. However, in this paper, we only deal with a part that all authors attribute to the Eurasia plate. Thus, we ignore possible complications introduced by the presence of an America–Eurasia boundary.

## 3. Offshore deformation zone

Zenisu ridge is an elongate, elevated feature running roughly NE–SW about 60–80 km south of the edge of the Nankai accretionary prism over a length of about 150 km. This ridge appears to mark the prolongation of a morphological trend within the Izu–Ogasawara ridge [21]. Seismic [8], other geophysical [22,10,13] and submersible [9] data indicate that the ridge is the surface expression of compressive tectonism which has broken through the oceanic lithosphere and is still active today. A double thrust, with a 30° northward near-surface dip, bounds to the southeast the western part of the ridge and has resulted in a 3 km vertical basement offset [10]. The

Fig. 3. Distribution of active faults, tectonic lines and active thrusts in central Japan and the Nankai trough. Topographic and bathymetric contours are from [52]. *I* = active fault; *II* = thrust; *III* = blind thrust; *1* = Atera fault; *2* = Yanagase fault; *3* = Nukuni–Neodani–Umehara faults; *a* = Hida Range; *b* = Kiso Range; *c* = Akaishi Range; *d* = Nobi Plain; *e* = Mino–Hida Highland; *f* = Noto peninsula.

amount of shortening, based on this geometry, may be about 10 km. The deformation is obviously younger than the 23 m.y. old basement. On the basis of a mechanical model related to the flexure of the plate as it approaches the subduction zone [19], Lallemant et al. [10] estimated deformation as beginning at 1.25 m.y. for a 4 cm/yr rate. The rate of shortening would thus be about 20% of the subduction rate.

The deformation progressively dies out southwestward near 137.2°E. It also extends southward as incipient compressive deformation and is observed on the southern side of the 'South of Zenisu' ridge [10] (which is located 50 km south of Zenisu ridge—see Fig. 2). Northeastward, the eastern part of Zenisu ridge displays a more complex deformation pattern, with distributed shortening in addition to the deep crustal thrust system. Thus, the compressive deformation decreases both southwestward along the strike of the ridge over 150 km and southeastward perpendicular to its strike to a maximum distance of 120–150 km from the plate boundary. Chamot-Rooke and Le Pichon [19] have speculated that this deformation pattern is in steady state related to the plate subduction–collision context. However, this is not possible to test in the oceanic environment because older possible compressional features must have disappeared below the accretionary wedge.

The Suruga trough and eastern Nankai trough accommodate the greater part of the Philippine Sea–Eurasia relative motion east of the TITL. This is clear from both the pattern of present-day strain accumulation [20] as well as the patterns of crustal deformation and tsunamis following great historical offshore earthquakes [15]. The lack of a large interplate earthquake along the Suruga trough since 1854—following centuries of documented rupture roughly every 100–150 yr [23]—has led to the expectation of an impending great earthquake along the Suruga trough (the 'Tokai Earthquake') [15]. From a geological point of view, the Philippine Sea–Eurasia convergence is accommodated not only by the offshore shortening but also by onshore shortening. Mogi [24] suggested that strain release from the 1891 Nobi earthquake unloaded the Suruga trough enough to significantly retard its rupture relative to the neighbouring part of the Nankai trough, which last ruptured in a great earthquake in 1944. Pollitz and Sacks [25] gave quantitative support to this idea. The crustal shortening which accompanied the 1891 Nobi earthquake is part of a larger integrated onshore deformation zone whose primary characteristics have been determined by the Izu–Osagawara collision, as described in the following sections.

## 4. Central Japan deformation zone

The internal deformation of central Japan is accommodated by intense left-lateral faulting along north-northwest striking faults, minor amounts of right-lateral faulting on west-northwest striking planes accommodating block rotations [26], and thrust faulting along north-northeast trending faults bounding the three major ranges of the Japan Alps arranged en-echelon to the west of the ISTL. The thrust faulting appears along the eastern flanks of the Hida and Akaishi ranges, along the ISTL, and along both the eastern and western flanks of the Kiso Range. Significant reverse faulting appears to continue to the southwest of the Kiso Range, with a topographic expression which diminishes gradually towards the Nobi Plain. Pollitz and Sacks [25] deduced that this southwestward continuation of the zone of shortening occurs along a blind thrust fault buried to a depth of 5–10 km that has accommodated about 2 km of long-term reverse motion in the Quaternary. The amount of shortening accommodated by the other notable thrust zones is likely to be similar. The entire intraplate deformation pattern in central Japan is consistent with the WNW–ESE direction of maximum horizontal stress.

The central Japan deformation zone is bounded by three major tectonic lines, the ISTL to the east, the MTL to the south, and the TITL to the west. With the exception of a 40 km long segment to the west of the Akaishi Range which is seismically active, no significant activity occurred along the eastern part of the Median Tectonic Line in central Japan during the Quaternary [26].

The TITL [27,28] appears as a sharp discontinuity in both topography and gravity patterns [29,26]. It has accommodated significant left-lateral motion during the Quaternary, with 1 km of documented slip for the Yanagase fault segment (Fig. 3) [26]. In close proximity to the TITL is the Nukumi–Neodani–

Umehara fault system [26]. Many surface ruptures, with a total length of about 120 km, occurred along these segments during the great 1891 Nobi earthquake [30,31]. The Neodani fault is arranged in an en-echelon fashion and has a total length of about 60 km [31]. A total of about 3 km of left-lateral movement is estimated for the Quaternary [31,32]. Mikumo and Ando [33] deduced a southward extension of the Neodani fault based on levelling data from the Nobi earthquake. Large earthquakes in 1894 and 1945 occurred along this part of the fault, delineating a continuous fault zone from the Neodani fault to the MTL. The Neodani fault and its southward extension were designated the FNBB line by Kanaori et al. [26], who recognized the signature in the gravity pattern.

The ISTL is a strike-slip fault with a reverse component [34,35] which forms the western boundary of the Fossa Magna. The ISTL offsets the MTL in a sinistral sense by about 12 km [28]. The Upper Neogene formations of the Fossa Magna are affected by en-echelon folds with NE–SW trending axes that are curved in the vicinity of the fault and indicate left-lateral motion.

Two causes of active deformation have been identified: the collision of the Izu–Ogasawara island arc with mainland Japan, and strong interplate coupling with the Philippine Sea subducting slab which descends beneath central Japan at a shallow angle along the eastern Nankai trough and Suruga trough [36]. Compressional deformation of the southern Fossa Magna and the present-day Izu peninsula [4,5] began in the Late Miocene, shortly before the closure of the pre-collisional marginal basin. The initial collision of the Izu peninsula with Japan occurred about 1 to possibly 2 m.y. ago [34].

We previously discussed the NW–SE shortening of the hot and weak crust of the Izu–Ogasawara ridge. Immediately north of the Izu peninsula, volcanic activity related to the Pacific plate subduction has also raised the heat flow and made the lithosphere relatively weak. The lower strength and the unresolvably shallow dip of the Philippine Sea plate below Mt. Fuji result in complex distributed shortening along several thrust zones. The cusp-shaped collision zone leads to a fanning out of the maximum compressive stress direction away from the collision front [37]. We relate the unusually high rate of deformation of the lithosphere to the west of the ISTL over an area of 200 km (perpendicular to the convergence direction) by 200 km to the complex distributed shortening north of the Izu peninsula. We consequently suggest that the 4 cm/yr of Philippine Sea interplate convergence must be partitioned into an 'oceanic' component and an 'onshore' component east of the TITL.

One constraint on the degree of the partitioning follows from the Nobi earthquake, which apparently relieved a large fraction of the strain accumulation on the eastern Nankai trough and Suruga trough [38]. The occurrence of 7 m left-lateral slip along the Neodani fault [39], far from the Nankai trough, suggests an effective onshore 'slip accumulation' rate of several millimetres per year. The Atera fault at the southern end of the Hida Range [26,29] has the same sense of slip as the Neodani fault, and although it has not slipped seismically in historical times its total Quaternary slip is estimated to be about 5–7 km [26], and the average rate of horizontal displacement has been found to be 5 mm/yr [40]. If similar long-term slip rates are appropriate for the Neodani fault, possibly 1 cm/yr of slip accumulation may be transferred from the Nankai trough to mainland Japan east of the TITL. This is consistent with the back-slip rates inferred by Yoshioka et al. [20] for the shallow part of the downgoing Philippine Sea plate along the Suruga trough (these rates are about 1 cm/yr less than expected if both the overriding and downgoing plates were behaving rigidly). The convergence rate between the Izu peninsula and central Japan as estimated with VLBI [18] accords with the rates estimated by Yoshioka et al. [20].

We can compare this estimate of 1 cm/yr for the transfer of shortening from the Nankai trough to onshore Japan to the amount of shortening necessary to produce the observed excess topography described in the introduction. As a doubling of the thickness of a zero sea-level crust produces a 5 km altitude, the maximum 200 $km^2$ excess surface produced by shortening over central Japan corresponds to about 40 km of shortening (or 20 mm/yr over 2 m.y.), compared to 10 km and 5 mm/yr for the eastern profile (see Fig. 2b). The average excess topography over the 200 km wide central Japan region gives an average shortening rate of 12 mm/yr, similar to the estimate made above.

The geometry of crustal deformation associated with the 1891 Nobi earthquake may give us some insight into the extent of the strain partitioning. Pollitz and Sacks [41] summarised in their table 10 the components of a fault model for the 1891 earthquake involving several metres of slip along multiple strike-slip and reverse faults. The total displacement of the Japanese crust is defined as the coseismic displacement plus the post-seismic displacements following the relaxation of the subcrustal asthenosphere, this relaxation being a pronounced and well-recognized feature throughout Japan [42–45]. Pollitz and Sacks [25] have calculated the total displacement and strain fields at the limit of complete asthenospheric relaxation, which corresponds to a time effectively a few decades after the earthquake. These deformation fields are shown in Fig. 4. The displacement field (Fig. 4a) clearly shows that the Nobi earthquake released approximately 0.5 m of slip with respect to the eastern Nankai trough and Suruga trough, and that this release is sharply attenuated along a line which coincides roughly with the seaward extension of the TITL. This is reflected in the pattern of total strain release (Fig. 4b), which shows up to $2\text{–}4 \cdot 10^{-6}$/yr of strain release along the eastern Nankai trough and Suruga trough and which also attenuates rapidly west of the extension of the TITL.

The timing of the onshore deformation is available for several mountain ranges and faults. The history of sedimentation east of the Akaishi Range suggests that the latter has been rapidly uplifted since the Late Miocene [46,47]. Similarly, the Hida Range has been uplifting rapidly since the Middle Pleistocene [48,49], as attested to by the cooling history of an exceptionally young exposed granitoid pluton [50]. Differential uplift of the Mino–Hida Highland (southwest of the Atera fault) and the southwestern Kiso Range has occurred throughout the Quaternary [51], and extrapolation of the present vertical displacement rate of the Atera fault into the past yields a 0.8 m.y. age [52]. Erosion surfaces of Pliocene age are extensively developed in these mountains, suggesting a pause in activity between the Late Miocene and Middle Pleistocene–Present stages of deformation. This is correlated with the presence of a deep Early Pleistocene paleotrough between the Izu peninsula and central Japan [53] and supports the idea of a polyphase tectonic evolution for the Izu area [34]. The age of the latest stage of onshore deformation accords broadly with the time of the collision of Izu peninsula with mainland Japan.

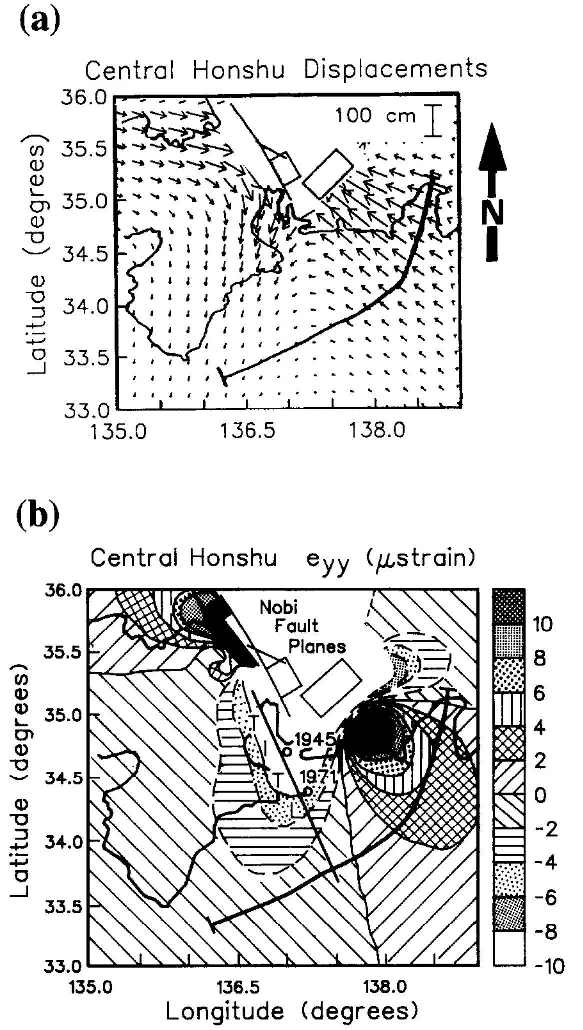


Fig. 4. Geographical pattern for (a) vector displacement and (b) strain release $e_{yy}$ due to the 1891 Nobi earthquake (defined as the coseismic + post-seismic strain release calculated at the boundary of complete relaxation). The $y$-axis is taken to be the direction N40°W. The heavy line represents the axis of the Nankai trough. Four rectangular fault planes (two vertical and two dipping fault planes) used to model the Nobi fault ruptures [41] are shown for reference. In (b), positive/negative values correspond to an increase in tension/compression. The epicentres of the $M = 7.1$ 1945 Mikawa and $M = 6.1$ 1971 Atumi–Oki earthquakes [55] and the oceanward extension of the Tsuruga Bay–Ise bay Tectonic Line are superimposed. Modified from [25].

## 5. Discussion

The offshore and onland zones taken together comprise a region of distributed deformation covering an area of 200 by 350 km. The fact that the TITL and its oceanward extension mark the western boundary of both the onshore and offshore deformation indicates that significant compressive stress concentration parallel to the Philippine Sea–Eurasia convergence direction occurs repeatedly in both the on- and offshore domains east of the TITL and its extension, but not to the west of it. Which factors are responsible for the concentrated stress buildups in the onshore and offshore regions?

We have previously identified the stress field generated by the incoming Izu block along a curved collision zone as the primary driving force for the onshore deformation, with strong coupling of the downgoing Philippine Sea plate as an additional secondary factor. Since the onshore deformation acts in such a way as to relieve a fraction of the stress buildup along the eastern Nankai trough (Fig. 4b), this would appear to inhibit rather than to promote the formation of the intra-oceanic thrusts. An overriding factor, however, is the internal deformation of the Izu peninsula and the Izu–Ogasawara ridge to the south. The N–S deformation of this island arc ridge, as previously discussed, leads to a kinematic discontinuity along the western flank of the ridge. The existence of this discontinuity explains the concentration of higher average stress in the oceanic crust around the eastern Nankai trough. If the onshore deformation rate were sufficiently rapid, stress concentration in the oceanic crust would not be high enough to produce the observed intra-oceanic shortening. The existence of the intra-oceanic thrust faults suggests that the rate of shortening of the Izu peninsula and Izu–Ogasawara ridge (projected along the interplate convergence direction) is comparable with the total rate of shortening represented by the onshore deformation in central Japan. Firm estimates of the rate of shortening of the volcanic ridge do not yet exist, but a rate of shortening (integrated from the collision zone to the area of the Zenisu ridge) comparable with that of central Japan is consistent with the higher heat flow and reduced strength of the volcanic ridge.

There is no clear structural or seismic evidence for the continuation of the TITL or FNBB south of the MTL, except that their suspected traces follow respective lines of saddles in the bathymetry (Fig. 3), and the trace of the MTL may bend in a zone of distributed left-lateral shear within Ise Bay (see the 'Active Faults in Japan' map in [54]). If the oceanward extension of the TITL does indeed have tectonic significance, it should accommodate left-lateral motion between the MTL and the Nankai trough and right-lateral motion between the Nankai trough and Zenisu ridge, as shortening occurs to the south of Zenisu ridge. In this manner, it would be expected to produce a left-lateral offset of the Nankai trough. Such an offset does appear along the edge of the accretionary wedge, and reaches about 10 km. If, for example, a 5 mm/yr average slip rate is appropriate at present and can be extrapolated back in time, the initiation of the offset could be dated to about 2 Ma. This roughly accords with the age of the onset of onshore deformation. The extent of the great 1944 earthquake was apparently bounded to the east by the seaward extension of the TITL [41]. This is interpreted as reflecting the retarding influence of strain release from the Nobi earthquake [38,25] (Fig. 4b) and illustrates the existence of long-term onshore strain partitioning.

## 6. Conclusions

The Philippine Sea–Eurasia interplate convergence in central Japan is accommodated along the major subduction zones (eastern Nankai trough and Suruga trough) and a broad zone of distributed onshore and offshore deformation that has a dimension of 200 by 350 km. Both the onshore and offshore domains are bounded to the west by the TITL and its southward extension. The present phase of onshore deformation is accommodated through rapid uplift, major left-lateral faulting and major reverse faulting, probably all initiated during the Middle Pleistocene about 2 m.y. ago. This date coincides with the initiation of the collision between mainland Japan and the Izu peninsula, which is presently undergoing NW–SE shortening in response to the collision. Active shortening along the Izu–Ogasawara ridge exists to the south. West of this volcanic ridge, intra-oceanic thrust faults appear along the Zenisu ridge and

'South of Zenisu' ridge about 60 km and 100 km , respectively south of the Nankai trough.

The long-term slip rates of the major onshore left-lateral faults suggest that perhaps 1 cm/yr of the possible 4 cm/yr interplate convergence rate is accommodated by the intraplate deformation of central Japan. We suggest that the main driving force of the onshore deformation is related to subduction–collision of the hot and weak Izu peninsula along a very shallow angle. All of the onshore faulting is consistent with failure in the presence of the stress field with a WNW–ESE trend of maximum horizontal stress which characterises all of central Japan north of the MTL.

The offshore deformation is driven by a kinematic discontinuity between the actively shortening Izu–Ogasawara volcanic ridge and the Philippine Sea oceanic crust to its west. The intra-oceanic thrusts which are presently active have an initiation age which appears to be in broad accord with the onset of the latest stage of onshore deformation. This could be interpreted as suggesting that the collision of the Izu peninsula with mainland Japan is fundamentally responsible for both the onshore and oceanic deformation. In other words, as high volcanic islands with thick, hot, weak crust arrive within the subduction zone, diffuse shortening will spread southward along the ridge and trigger the discrete intra-oceanic shortening to the west of that ridge.

The presence of relatively thick crust all the way from the Noto peninsula to the southern Izu–Ogasawara ridge is the most important factor responsible for the great length (350 km) of the zone of strain partitioning, while the width (200 km) represents the width of the Izu–Ogasawara ridge above the depth of 2000 m (compare Fig. 2a and Fig. 2c), presumably because above this depth the crust of the ridge is too thick to subduct easily.

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