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A model for the plate tectonic evolution of the east-central Pacific based on SEASAT investigations

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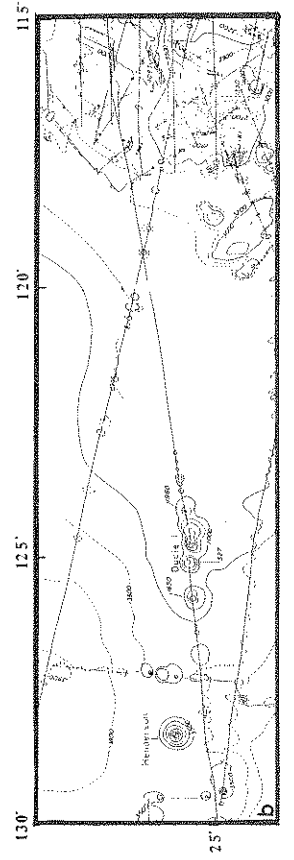
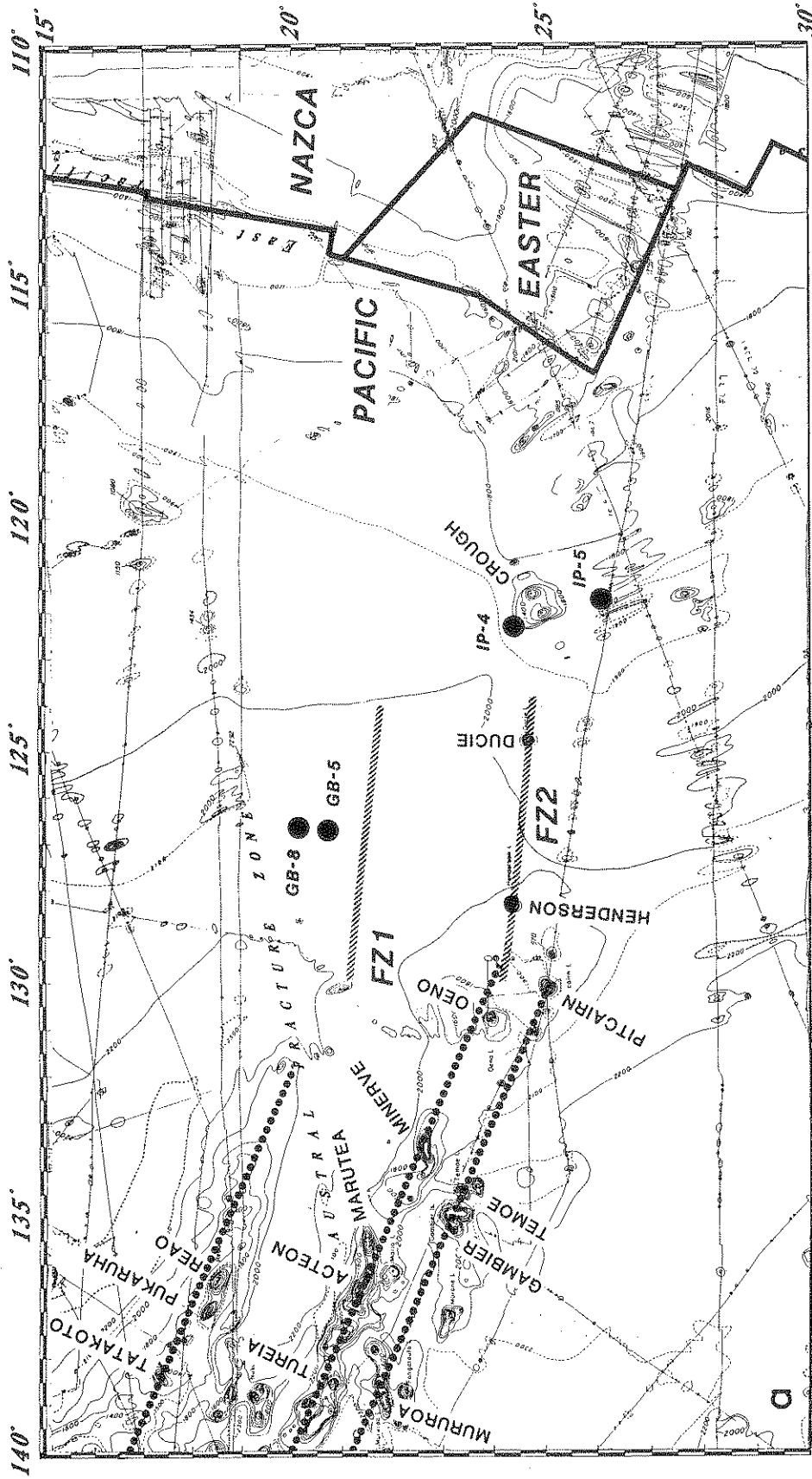
A complete SEASAT dataset is used to investigate the small-scale bathymetry of a 15° by 15° area of the east-central Pacific, northeast of Pitcairn Island, which was involved in the Miocene ridge jump and reorientation. Two fracture zones (FZ1 and FZ2) are mapped along 450 km, at an azimuth of N95°E, while the signature of the Austral Fracture Zone disappears east of 135°W. Several seamounts are discovered or confirmed along the Oeno-Henderson-Ducie lineament, and the degree of isostatic compensation of a number of islands at the eastern end of the Tuamotu group is studied: islands of the northern branch have practically no signatures in the geoid, while the southern ones must have formed on lithosphere at least 10 Ma old. On the basis of these constraints, we present a model of the evolution of spreading since 40 Ma B.P., in which the deactivation of the Austral Fracture Zone and the transfer of its offset to FZ1 in the Late Oligocene is interpreted as due to rift propagation, triggered by a northern Tuamotu hotspot; we propose that the Oeno-Ducie chain is the surface expression of a southern Tuamotu hotspot, deviated 15° in azimuth by the presence of a young and fresh fracture zone. Thus our model predicts that hotspots and ridge systems can have a moderate influence on each other's kinematics and surface expressions, which does not however extend to permanent trapping.

1. Introduction and background

Following the operation of the altimeter satellites GEOS-3 and SEASAT, the marine geoid is now well known over all oceanic areas, with a height accuracy of 10–30 cm and a horizontal resolution of 10–50 km. At wavelengths less than 500 km, geoid contours have been found to show a high level of correlation with seafloor topography, each type of tectonic feature (oceanic ridges and trenches, fracture zones, volcanic chains and seamounts) causing a specific signature in the geoid. The analysis of the shape, amplitude and wavelength of the corresponding geoid anomaly can yield information on internal densities, as well as on the state of isostatic compensation of the structures. Conversely, the systematic use of geoid anomalies derived from altimeter data may help

discover unsuspected bathymetric features. This approach has proven very effective especially in equatorial and southern oceanic areas where shiptrack coverage is often poor (e.g. [1]). Numerous uncharted seamounts have thus been detected either by visual inspection of the SEASAT data, or by applying matched filtering methods [2,3]. Other tectonic features have also been identified from their geoid signature; for example, Ruff and Cazenave [4] have recognized incipient subduction at the Hjort trench, south of New Zealand.

The present study focuses on an area of the Pacific plate centered around 22°S and 127°W, approximately 1000 km west of the East Pacific Rise, and covering about 1500 km × 1000 km. It includes the young volcanic island of Pitcairn, the atolls of Oeno, Henderson and Ducie (see Fig. 1), and also a site of intense intraplate seismicity



identified by Okal et al. [5] around 20.7°S and 126.8°W . This location, called “GB-5” or “Region C” by these authors, experienced a long seismic swarm during 1977–1979, featuring 97 recorded earthquakes with a maximum $m_b = 5.5$. After 4 years of quiescence, a magnitude $m_b = 6.0$ event took place 65 km to the north in July 1983, followed within two hours by an $m_b = 5.4$ aftershock. Both GB-5 and the epicenter of the 1983 events (GB-8) have been quiet ever since.

The extremely low density of shiptrack coverage in the area results in mostly blank bathymetric maps [6], and thus, no bathymetric information was available to Okal et al. [5] to suggest a character of weakness in the lithospheric plate and explain a preferential release of the seismicity at this site. In order to further investigate the possible existence of significant bathymetric features around GB-5, Sailor and Okal [7] examined SEASAT profiles over a small area, extending from 20°S to 22°S and 126°W to 128°W . They concluded that no large uncompensated seamount exists at the site; instead, they found evidence, on the six profiles examined, for a small fracture zone signature 70 km to the south, but due to their limited spatial coverage, they only mapped this new “Region C Fracture Zone” over a very short distance, on the order of 150 km. In the summer of 1980, a small scale reconnaissance in the vicinity of GB-5 using the SEABEAM-equipped N/O “Jean Charcot”, confirmed the absence of a large seamount structure, but revealed rugged seafloor morphology, with the prevailing tectonic direction oriented east-west, and relief of ± 400 m (J. Francheteau, personal communication, 1981).

The exact geological history of the lithosphere in the vicinity of seismic sites GB-5 and GB-8 is itself obscure, since it was generated during the early Miocene, a time when magnetic anomalies are few and broadly separated in age. At that time, the old Farallon Ridge (where the Pacific plate

was being generated) underwent a westward jump of some 500 km, and a clockwise reorientation of about 40° [8]. Based on a reconstruction of this episode by Mammerickx et al. [9], Okal and Bergeal [10] mapped the boundary, on the Pacific plate, dividing lithosphere generated at the old and new spreading systems. Within error bands due to the limited amount of available data, they found that GB-5 was indeed in the immediate vicinity of this boundary (which they called Jump (“J-”) line). Their simple model was, however, based on the assumption of an instantaneous and complete jump and reorientation; we will see that much more complex phenomena took place in the Region C area.

The purpose of the present paper is to carry out a more complete investigation of the tectonics of the area in order to put stronger constraints on the geological history of the Miocene Farallon Ridge jump. By considering a larger region (extending from 15°S to 30°S and 115°W to 130°W), and analyzing all available SEASAT profiles, we present evidence suggesting that the reorientation involved an intermediate stage of east-west spreading, and that the evolution of the Farallon Ridge just prior to the jump included an episode of rift propagation, probably controlled by the Easter Island hotspot. In addition, the study of the geoid in the vicinity of a number of islands in the eastern part of the Tuamotu archipelago provides data on their degree of compensation, and constraints on their formation.

2. SEASAT observations over the region ($15\text{--}30^{\circ}\text{S}$, $115\text{--}130^{\circ}\text{W}$)

Across a fracture zone, the geoid features a step-like behavior over distances of ~ 100 km. This offset (downward from the younger, shallower side to the older, deeper side) is due to the

Fig. 1. (a) Principal features of the study area, superimposed on bathymetry from Mammerickx et al. [6]. Full lines indicate present-day plate boundaries, full dots are relevant intraplate seismic epicenters. The dotted lines show the initial paths of hotspots HS1 and HS2, as well as the path of the Pitcairn hotspot. The striped lines show the extent of the fracture zones mapped in the present study. The names of seamounts and islands discussed in the text are outlined. (b) Close-up of the GEBCO bathymetric map [17]. Note substantial differences in the interpolation of shiptrack data, and absence of Crough Seamount.

files, and the corresponding residual geoid anomalies are presented in Fig. 3a and 3b. Data processing involved the removal of the global geoid GRIM-3 [18], up to degree and order 10 (corresponding to wavelengths greater than 4000 km, whose origin is unrelated to the surficial structures we are investigating), as well as the linear trend in individual tracks. A close examination of these height anomalies reveals two small fracture zones (hereafter FZ1 and FZ2): FZ1, extending from ~ 21°S, 130°W to ~ 22°S, 124.5°W; and FZ2, extending from ~ 24°S, 130°W to ~ 25°S, 124°W. FZ1 is the "Region C Fracture Zone" earlier

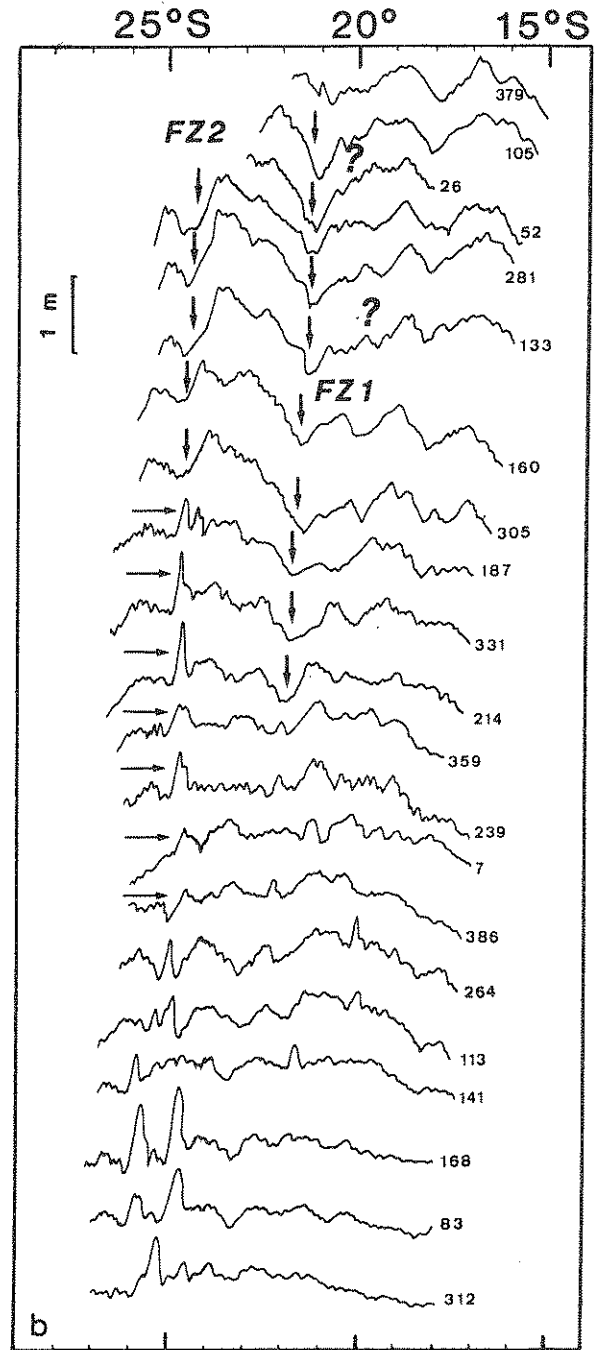
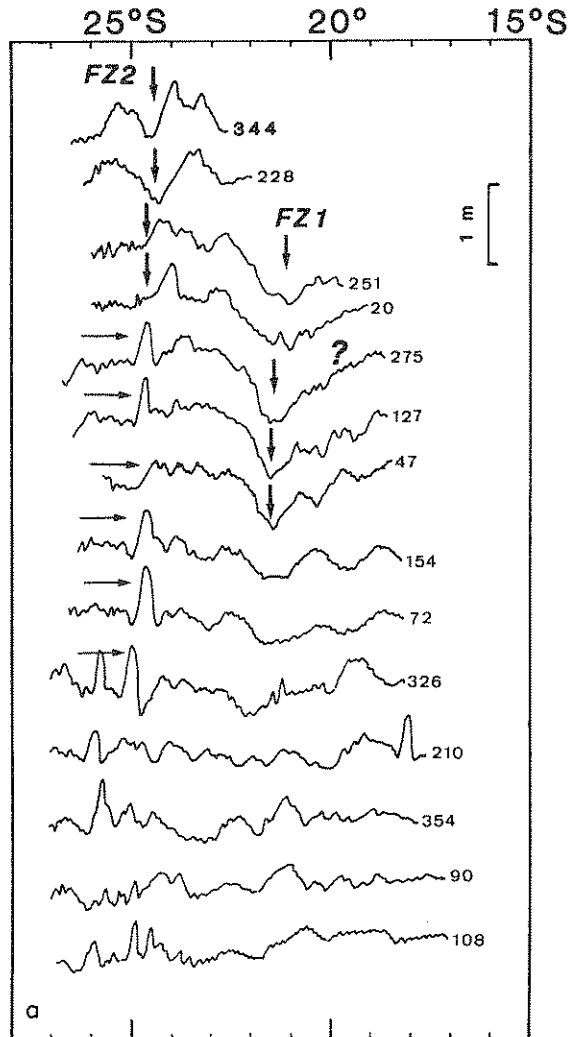


Fig. 3. (a) Geoid heights along selected ascending SEASAT tracks crossing study area. The vertical arrows point to the signatures of FZ1 and FZ2, while the horizontal ones indicate seamount signatures. Note also the absence of geoid anomalies at the expected location of the Austral FZ (question marks). Long-wavelength components have been filtered out. (b) Same as Fig. 3a, for descending tracks.

identified by Sailor and Okal [7]. FZ2, located approximately 300 km to the south, is parallel to FZ1 and coincides with the eastward lineament of the three atolls Oeno, Henderson and Ducie. The identification of the nature of FZ1 and FZ2 from their geoid signature is made difficult by the relatively small level of the signal. In particular, the along-track derivatives of the signal turned out to be too noisy for systematic use.

In the case of FZ1, the signature of the step at the fracture zone is partially masked in Fig. 3 by what must be the effect on the geoid of the bathymetric trough accompanying the fracture zone. However, the polarity of geoid offset associated with FZ1 is downward from south to north, a sense noted by Sailor and Okal and indicating a change from younger to older material. This is particularly clear on illumination images of the surface of the Pacific ([1, Fig. 9], and D.T. Sandwell, personal communication 1983). On the other hand, the offset along FZ2 is upwards to the north, suggesting that the material located between these two fracture zones is younger than the portions of lithosphere north of 20°S and south of 25°S, and requiring an indentation of the Pacific plate by the Farallon one. While the interpretation of individual profiles cannot always distinguish between fracture zones and seamounts, the good correlation existing between adjoining tracks (e.g., 52, 281, 133, 160, 305) appears typical of the signature of a fracture zone, although an alternate interpretation as an elongated plateau cannot be totally dismissed at this time.

FZ1 starts to be visible on profile 379, is well developed with an amplitude step of 0.5 m on profiles 105 to 305 (129°W to 126.5°W), then progressively fades out to become undetectable east of profile 214 (~ 124°W). Similarly, FZ2 gives a clear geoid signature between 129.5°W and 127°W (profiles 52 to 187), as well as around 124°W (profiles 7 and 386). The offset amplitude ranges from 0.5 to 0.8 m. As the lithosphere is about 20 Ma old in this region, this infers an age offset in the range 3–7 Ma, depending on the thermal structure of the plate in this area [15].

East of 126°W, the fracture zone signature due to FZ2 changes to seamount signature: the geoid profiles cross a complex of seamounts associated

with Ducie Island. Although the bathymetric chart of Mammerickx et al. [6] represents Ducie Island as an individual feature, we believe that the bathymetry in the vicinity of Ducie is very likely more correctly pictured by the GEBCO chart [17], as the geoid data indicate the presence of several seamounts (see profiles 331, 214, 239, 275, 127, 72 and 154). More to the east, profiles 264, 113 and 326 also present a seamount signature (~ 25°S, 122°W). Indeed, Mammerickx et al.'s [6] 1975 map shows a major edifice * topped by two seamounts reaching 1000 m b.s.l. at 25°S, 122.2°W and 24.8°S, 121.7°W. The flank of this structure was the site of a major normal faulting earthquake in 1955 [19]. Surprisingly, this feature is absent from the 1982 GEBCO maps (see Fig. 1b), but it is clearly present in the SEASAT geoid (Fig. 4). Still further east, SEASAT reveals another group of seamounts around 25.6°S, 121.5°W (profiles 326, 210 and 141). Additional seamount topography in the vicinity of seismic locality IP-5 (see Fig. 1) was reported by Sandwell [1], and is confirmed. Further east, in the immediate vicinity of the East Pacific Rise, several additional small seamounts are detected from their geoidal signature (profiles 168, 83, 312, 354 and 108). The geoid anomalies caused by these seamounts appear always very small (amplitude less than 1 m), and although the exact bathymetry is unknown, this suggests substantial isostatic compensation of these features, which was to be expected, since the plate supporting these loads is very young.

2.2. *The absence of the Austral Fracture Zone*

North of FZ1, SEASAT data clearly fail to detect any extension of the Austral Fracture Zone east of 135°W, as shown by question marks in Fig. 3a and 3b. A weak fracture zone signature at 18°S, 127–128.5°W is suggested from profiles 379, 105, 52 and 281, but the corresponding feature appears distinct from the expected location of the Austral FZ.

* Okal [19] has proposed to name this edifice after S. Thomas Crough.

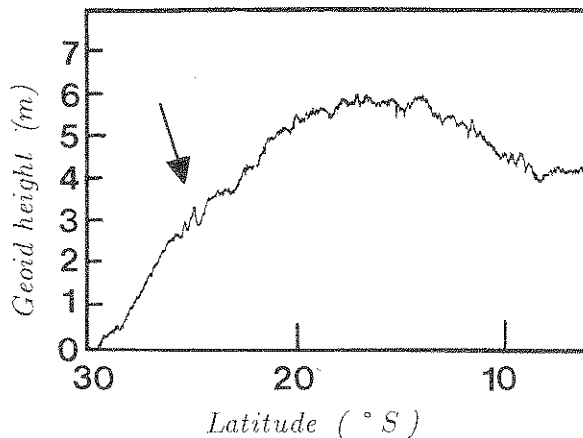


Fig. 4. Close-up of signature of Crough Seamount (arrow) on SEASAT track 113. This is raw data, from which long-wavelength components have not been removed.

2.3. Signature of a few Tuamotu Islands

The Tuamotus make up a large group of islands extending in a generally WNW-ESE direction for about 1500 km. To the southeast, and not structurally part of the Tuamotus, The Gambier–Mururoa–Duke and Gloucester–Hereheretue group has been interpreted as the trail of the Pitcairn hotspot by Duncan et al. [20]. The origin of the other islands is more obscure: presently they are all atolls and no basalts have been drilled from any of them. The Tuamotu archipelago sits on a broad plateau, in some areas 400 km across, continuous at the 2000-m isobath. This plateau does not show up as a massive feature in the geoid, indicating strong isostatic compensation; and anomalously thick crust (27 km) has been reported under Rangiroa [21]. All this evidence makes the Tuamotu system very comparable to the Iceland–Faeroe Ridge [22], or the Walvis Ridge [23], and suggests generation of the plateau by an on-ridge hotspot. DSDP Hole 318, 70 km northeast of Rangiroa, reached volcanogenic siltstones containing redeposited reef fossils dated 19–51 Ma [21]; this sets a minimum age of ~ 52 Ma for volcanism, which could have been on-ridge given that the plate is only 59 Ma in this area. However, the morphology of the chain suggests that several episodes of volcanism were involved in the creation of

the individual islands; furthermore, ^{40}Ar – ^{39}Ar dates obtained from dredged basalts show volcanic activity taking place 42 and 48 Ma B.P., at the nearby edge of the plateau west of Mataiva, where the plate age is 62 Ma, clearly requiring off-ridge volcanism [25]. Thus, a hotspot model for the origin of the Tuamotus requires at least two hotspots [26].

In the vicinity of our study area, the Tuamotus break into a northern segment (Tatakoto, Pukaruha, Reao) and a southern chain (Hao, Tureia, the Acteon Group, Muratea and the Minerve Reefs; see Fig. 5), a situation in line with a two-hotspot generation model. In order to study the time relationship of the proposed two hotspots, we have investigated the geoid in the immediate vicinity of a number of Eastern Tuamotu islands. While a SEASAT altimeter reflection exactly over an island or atoll always gives rise to a very sharp signal, the signature of an island or seamount on a profile missing its center by 20–40 km is a powerful discriminant on its state of compensation (e.g., [16]).

Fig. 5 shows SEASAT raw data along tracks 37, 117 and 268, sampling the neighborhood of the Gambier complex, and the Acteon and Marutea group in the south, as well as the islands of Pukaruha and Tatakoto in the north. Tracks 37 and 117 miss the exact location of Acteon and Marutea respectively by about 25 km. Tracks 37 and 268 similarly pass over the flanks of Pukaruha and Tatakoto at comparable distances from the atolls. While the southern atolls (Acteon) give a very sharp signal along these tracks, the northern ones have a barely detectable signature. This indicates that Acteon and Marutea are relatively uncompensated, while Pukaruha and Tatakoto are strongly compensated and must have formed on-ridge or in its immediate vicinity. Additional results similarly show that the southern islands of Hao, Amanu, Negonego, Menuhangi and Tureia are uncompensated, as well as Oeno, Ducie and Crough Seamount to the east; they must have formed off-ridge. No profiles come close enough to Henderson, Reao and the Minerve Reefs to provide useful data.

In order to obtain a more definite age relationship between the islands and the plate, we have

used the methods developed by Cazenave and Dominh [27], and bathymetry from Mammerickx et al. [6] and from the map series of the Institut Géographique National [28], to infer the flexural rigidity of the plate at the time of the islands' generation. It should be emphasized, however, that

these results are only tentative given the often poor bathymetric coverage, and our ignorance of any irregularity in the density structures inside the atolls. In the case of the northern islands Tatakoto and Pukaruha, the flexural rigidity D has to be less than 10^{20} N m, 2 orders of magnitude lower than

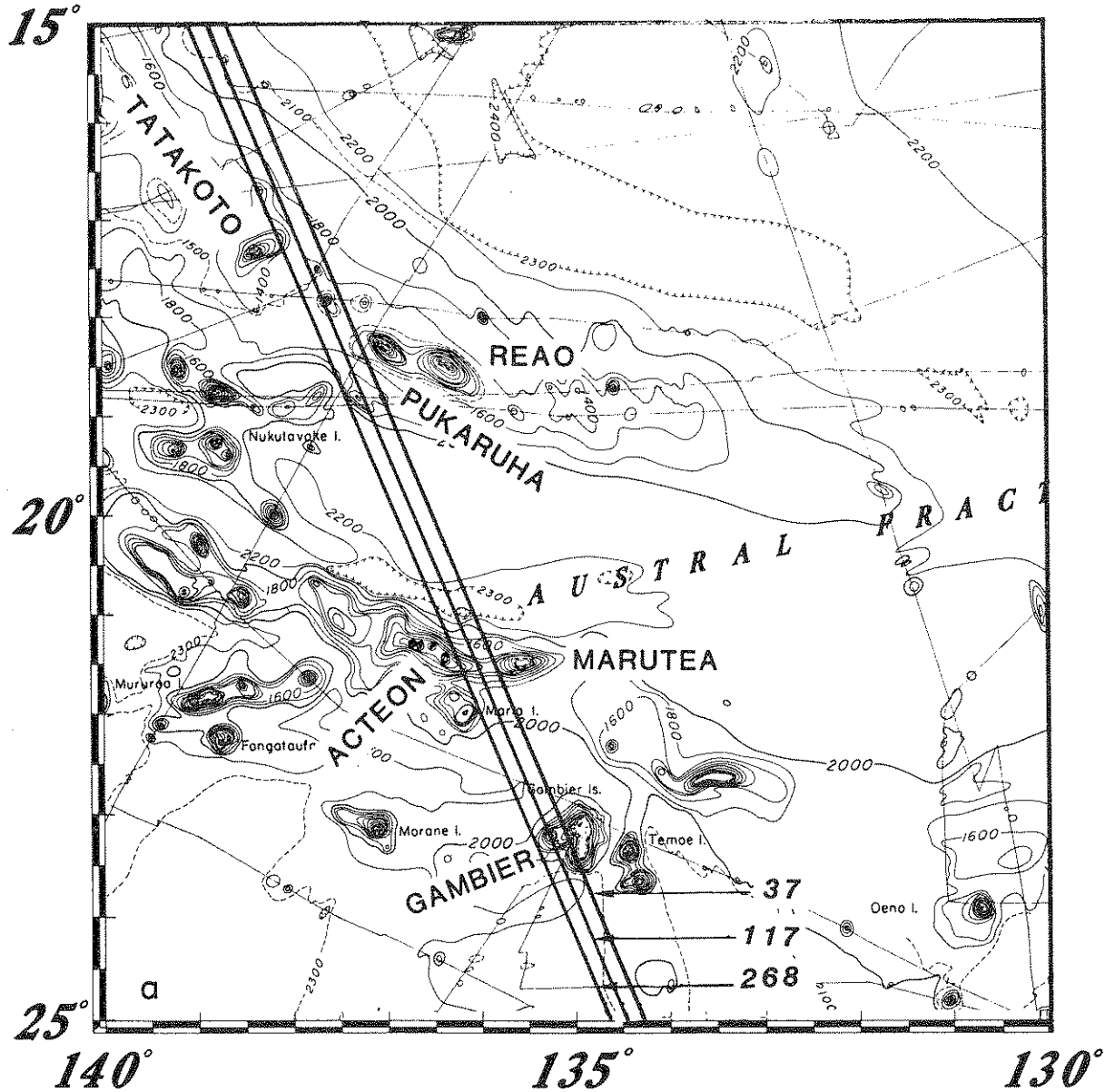


Fig. 5. (a) Map of Eastern Tuamotu islands, showing the northern plateau, with the islands of Tatakoto, Pukaruha and Reao, and the southern group including the atolls of Tureia, Acteon and Marutea, as well as the younger, volcanic Gambier complex. Superimposed are relevant SEASAT tracks; see text for details.

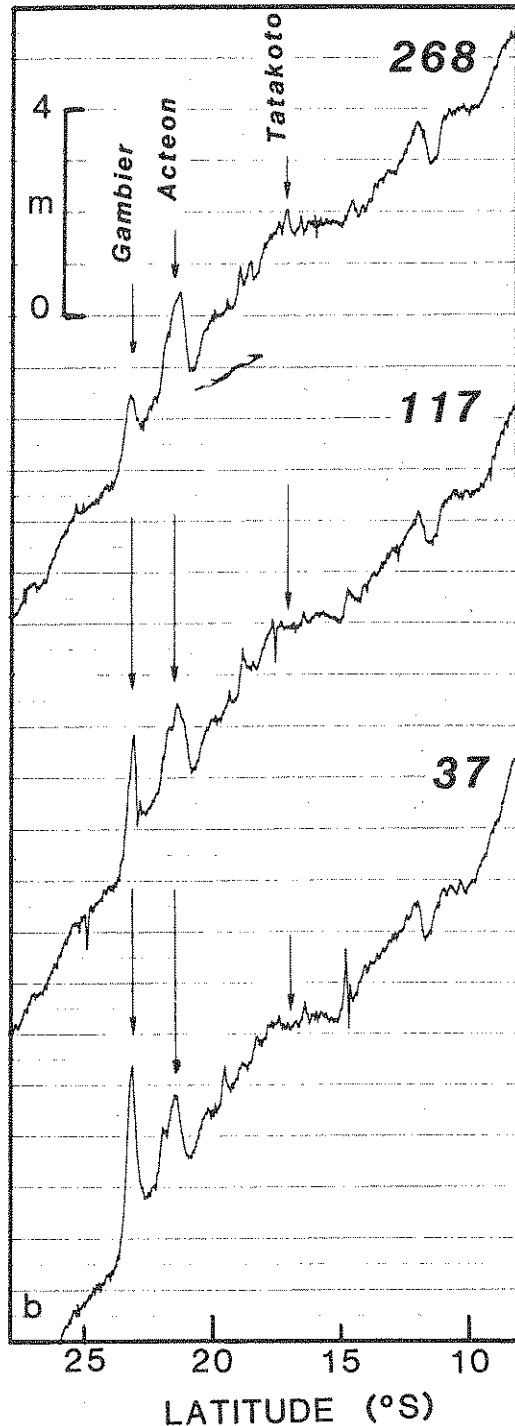


Fig. 5b. SEASAT data (unfiltered) along tracks 37, 117 and 268. Note strong signal of Gambier and Acteon groups, even for 117 and 37, which miss Acteon and Marutea by about 25

for islands developed on a mature plate [29], implying an elastic thickness T_e of at most 2.2 km. For the southern islands (e.g., Negonego), $D = 10^{21}$ N m and $T_e = 5$ km; for Crough Seamount, $D = (1-2) \times 10^{21}$ N m and $T_e = 5-6$ km. Corresponding values for the "flexural" age, or effective age of the plate at the time of loading [15,16], would be 1-2 Ma for the northern islands and 5-7 Ma for the southern ones. Flexural ages are not directly representative of the age of the plate at the time of loading, because of probable re-heating of the plate during island formation [30,31].

3. Tectonic discussion

3.1. Properties of the fracture zones

One of the most interesting properties of FZ1 and FZ2 is their orientation: both of them strike N95°E. This figure is intermediate between the azimuth of N70°E characteristic of the pre-jump Farallon regime, and the present N110°E orientation of spreading at the East Pacific Rise. Thus, the SEASAT data indicate that the reorientation of spreading in the East Pacific at the (present-day) latitudes 20-25°S was not instantaneous, but rather involved an intermediate regime oriented N95°E, during which the eastern plate (Farallon or Nazca) indented the Pacific one by several hundred kilometers. In the absence of magnetic anomaly data in the region, it is difficult to quantify the duration of this episode of spreading at an intermediate azimuth. Indeed, there is practically no way of constraining spreading rates between 18.5 and 12 Ma [9,32], and there exists documented evidence for significant and short-lived changes in spreading patterns in neighboring areas of the Pacific (e.g., [8]). An average (half) spreading rate of 7.4 cm/yr can nevertheless be obtained from anomalies 4, 5, and 6 south of FZ2; assuming symmetric spreading at the East Pacific Rise, this agrees with Handschumacher's [32] value of 16 cm/yr full rate; the Oligocene spreading rate

km. On the other hand, Tatakoto and Pukaruha are not detected in similar geometrical conditions (track 37 for Pukaruha and 268 for Tatakoto).

was substantially slower (~ 5 cm/yr). The 570 km along which FZ1 is documented then require a lifetime of at least 7.5 Ma for the FZ1 system. Similarly, the spreading system responsible for FZ2 lasted at least 8 Ma.

Another property of the two fracture zones is their apparent relationship with the Easter platelet. This feature has been recognized by Herron [33], and analyzed in detail most recently by Engeln and Stein [34]. FZ2 and its associated seamounts blend into the southwestern boundary of the platelet: although no clear geoid signature exists for FZ1 east of 124°W , the extrapolation of FZ1 to the east runs into the northeastern boundary of the platelet. In addition, the SEASAT signature of the northeastern boundary of the Easter platelet at 111°W is comparable to that of a fracture zone, with the youngest side to the north. The Easter platelet could have developed its present boundaries along lines of previous tectonic activity, such as the FZ1 and FZ2 lineaments.

3.2. The Oeno-Henderson-Ducie chain

Our investigation adds considerable data to the problem of the Oeno-Henderson-Ducie alignment. Namely, we have shown that seamounts aligned with the three islands exist, notably at 126°W , 123.5°W and 121.5°W . Additionally, we confirm the presence of a major edifice (Crough Seamount) at 122°W , in the vicinity of seismic epicenter 1P-4. We also reveal the existence of the underlying FZ2 fracture zone, running from 130°W to at least 124°W ; note that it *does not* go through Oeno (Fig. 1).

The exact origin of the three uninhabited islands is obscure. At present, they are small atolls. No detailed geological mapping has been carried out on them, but Henderson is known to be uplifted 30 m, a situation resulting from lithospheric flexure under the weight of the younger Pitcairn, 235 km to the southeast [29]. The islands must have volcanic cores, and their signature on the geoid shows that they have formed *intra*-plate, albeit on young crust. Their remarkable alignment and spacing suggest formation by a "plume" or "hotspot", which is likely to also have generated the southern branch of the Tuamotu islands

(Acteon, Marutea, and the Minerve Reefs).

A large earthquake occurred in 1955 in the immediate vicinity of Crough Seamount. Okal [19] has shown that its mechanism requires normal faulting, with tensional stress oriented horizontally at an azimuth of $\text{N}35^\circ\text{E}$. In addition, its focal mechanism requires a mantle hypocentral velocity. Tensional earthquakes are indeed current in very young oceanic lithosphere; their origin is not well understood, although several mechanisms for their occurrence have been proposed, (e.g., [35]). However, Wiens and Stein's [36] thorough catalogue shows that the 1955 event is by far larger than any other tensional event known in very young Pacific lithosphere. On the basis of a comparison with large normal fault earthquakes following documented volcanic activity at Deception Island, Antarctica, and Fernandina, Galápagos, Okal [19] has proposed that the 1955 event could represent a tectonic readjustment of the edifice of the seamount, following an episode of active volcanism. Such an interpretation must of course be confirmed by proper surveying, petrological and geochemical investigations of the seamount structure; furthermore, it would not necessarily mean that the present location of the hotspot is below Crough Seamount, since the activity could merely represent post-erosional volcanism, known to be current on Hawaiian volcanoes.

A striking feature of the chain is that its azimuth, from the center of Crough Seamount to Oeno, differs by 15° from that of the absolute motion of the plate, exemplified by the nearby Pitcairn-Gambier-Duke of Gloucester chain. While some islands chains (notably Hawaii) are locally curved away from the azimuth of absolute motion of the oceanic plate [37], this effect is not known to extend over lengths of 1000 km. West of Oeno, where FZ2 is absent, the trend of the chain is regular, and parallel to the Pitcairn-Gambier-Duke of Gloucester chain; thus we may be in a situation where the fracture zone has "trapped" the surficial expression of the hotspot, and deviated the lineament of the chain. This interpretation is at odds with the case of the Hawaiian chain, whose orientation is robust to the presence of major fracture zone systems (e.g., the Murray or

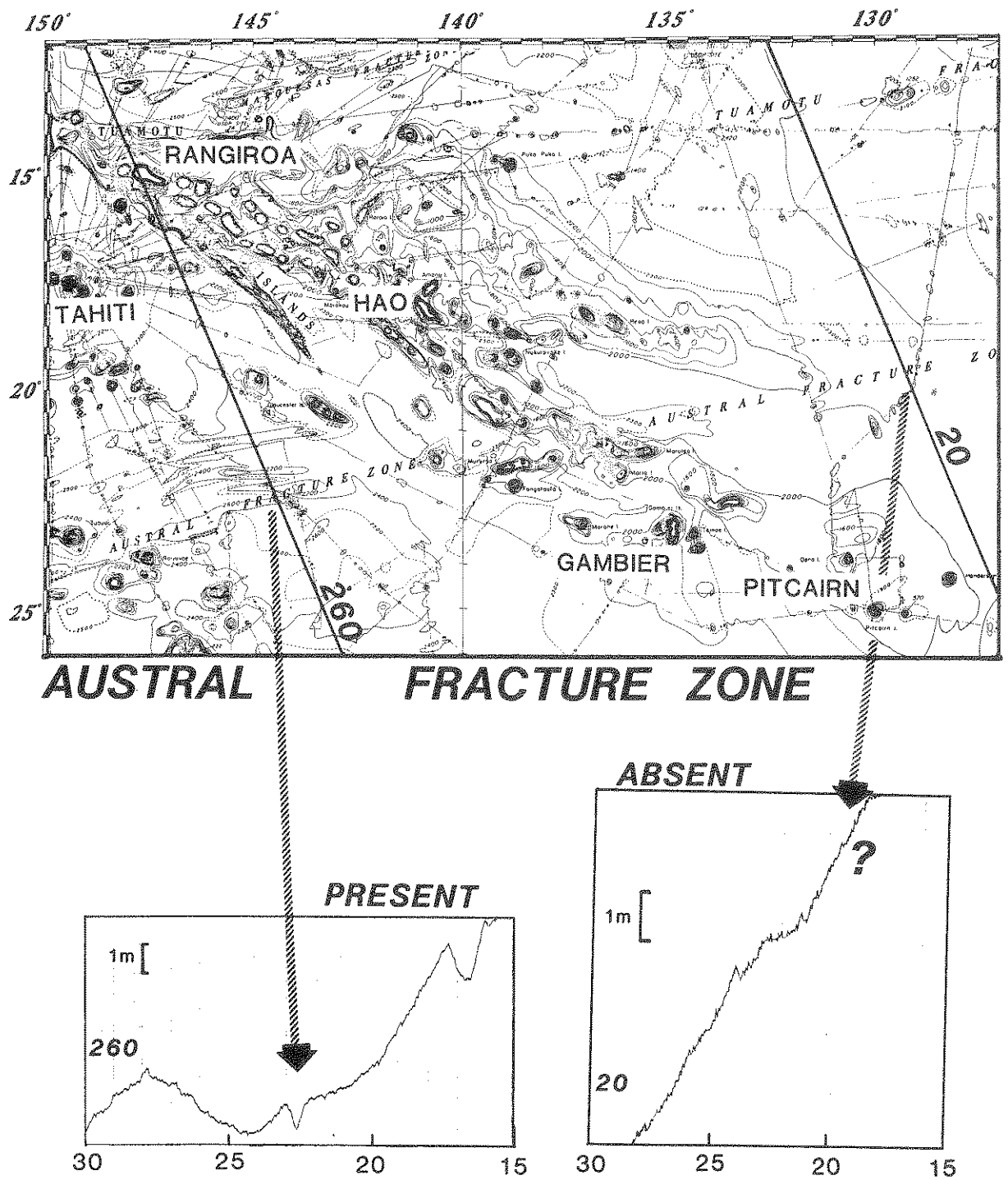


Fig. 6. Disappearance of Austral FZ signature east of 135°W. Top: Bathymetric map [6]; bottom: SEASAT data along tracks 20 and 260; arrows point from topographic location of crossing to corresponding point on SEASAT profile; note strong signal present on the western crossing, but absent from the eastern one (question mark), despite larger scale of figure.

Molokai); however, two major differences with Hawaii are the smaller discrepancy in azimuth between the hotspot chain and the fracture zone, and the much younger age of the fracture zone in the case of Oeno-Ducie. In particular, the lateral offset between Crough Seamount and its expected location in the axis of the Tureia-Oeno line, is only 300 km to the north. Lateral flow away from a hotspot has been proposed by Morgan [38] under young lithosphere over distances of 1000 km or more, and evidence to support his model is slowly building up [39,40]. An interaction between hotspot and fracture zone similar to the Oeno-Ducie situation, has been proposed at the Eltanin system, further south ([41]; and L.J. Henderson and R.G. Gordon, personal communication, 1984). FZ2, located in young, thin, and still hot lithosphere, would be particularly well suited to provide a preferential output for a hotspot. In addition, a disturbed and weak FZ2 could lend itself easier to ongoing or late-stage volcanism than a regular midplate location. We wish to emphasize that we present this interpretation as a speculation. It raises questions, which can be tested by surveys, and petrological and geochemical investigations.

3.3. The evolution of the Austral Fracture Zone

The Austral FZ is the well-documented remnant of a major right-lateral offset of the fossil Pacific-Farallon spreading center. Mammerickx et al. [9] have called the spreading segments immediately to the north and south of this TF the Mendoza and Roggeveen Ridges, respectively. West of our study area, the signature of the Austral FZ in both the bathymetry and the geoid is sharp and clean: for example, in the vicinity of 25°S and 143°W, in the neighborhood of seismic epicenter AU-3 [5] between the Gambier-Cloucester and Austral chains, the fracture zone is clearly expressed in worldwide maps of the geoid (e.g., [42]). Individual SEASAT tracks show a downwards step from south to north of 0.8 m (see Fig. 6).

In the area we investigated, however, we failed to observe the geoid signature of the Austral FZ at longitudes 125–135°W. Despite scarce bathymetric coverage at the expected location of the fracture zone, magnetic anomalies 13 and 7 are clearly

offset between 480 and 530 km; with Oligocene spreading rates of 5 cm/yr, this corresponds to an age gap of 10 Ma across the Austral FZ. In particular, the offset of anomaly 7 proves both the activity of the Mendoza and Roggeveen Ridges at that time (27 Ma), and the existence of an active transform feature between them. The targeted area at 20°S and 127°W is one of considerably younger age than the AU-3 neighborhood. Thermal boundary models of the plate based on a $t^{1/2}$ dependence of depth with age lead to a geoid height linear in t , and thus to a constant geoid step for a constant age gap [43]; on the basis of the observation of well-developed fracture zones, a “plate” model in which the geoid step for constant age gap increases with diminishing plate age is usually preferred [12,13,44]. Thus, the signature of the Austral FZ should *increase* eastwards. Since we do not observe it in the geoid, we must come to the conclusion that the pattern of the Farallon spreading center had changed at this latitude, even before the time of anomaly 7 (27 Ma), resulting in the deactivation of the Austral FZ, and the eventual activation of FZ1 in its place.

It is likely that this process occurred through rift propagation [45], the northern (Mendoza) ridge growing at the expense of the southern (Roggeveen) segment. Furthermore, we propose that the agent responsible for the initiation of the rift propagation may have been a hotspot presently in the vicinity of Sala-y-Gomez Island.

As mentioned above, the northern branch of the Tuamotu plateau was formed on (or in the immediate vicinity of) the Mendoza Ridge. While detailed models of its formation may vary [26,46], the general orientation of the Tuamotu plateau requires a slow northward motion of the Farallon Ridge system with respect to the mantle, or conversely a slow drift of the hotspot southward along the Mendoza Ridge; when it encountered the large Austral FZ offset, it became *intra*-plate, and generated discrete, uncompensated islands exclusively in the Farallon plate, resulting in the abrupt termination of the Tuamotu Plateau at the Austral FZ, at 20.5°S and 131.5°W [46].

While agreeing with the general concept of this model, we propose the following modification (see Fig. 7a, b): when the hotspot hit the Austral FZ, it

triggered rift propagation, resulting in the deactivation of the Austral FZ, and the eventual transfer of the offset to FZ1. The lithosphere located in the area where the Austral FZ is expected, at 20°S and 128°W, is a genuine piece of Pacific plate generated at the Mendoza Ridge, and the boundary between crust generated at the

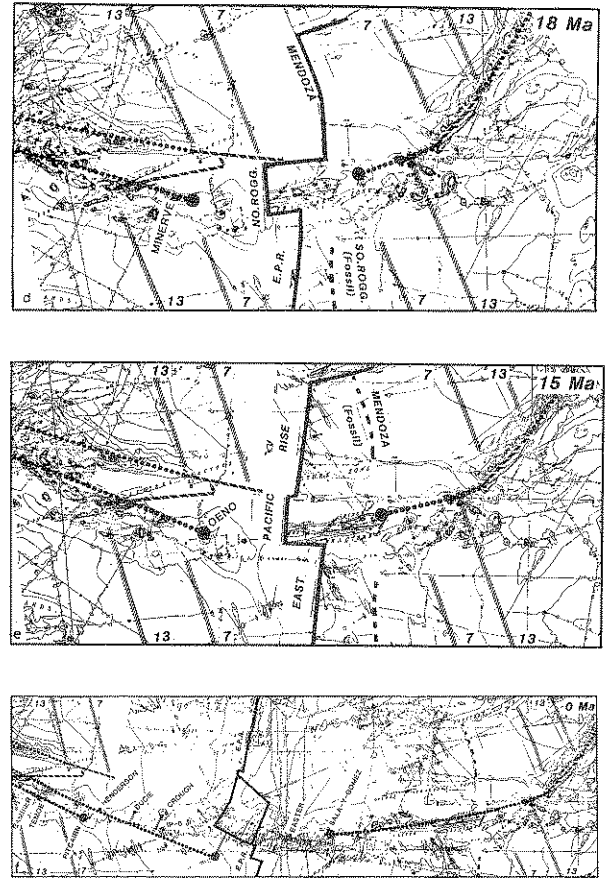
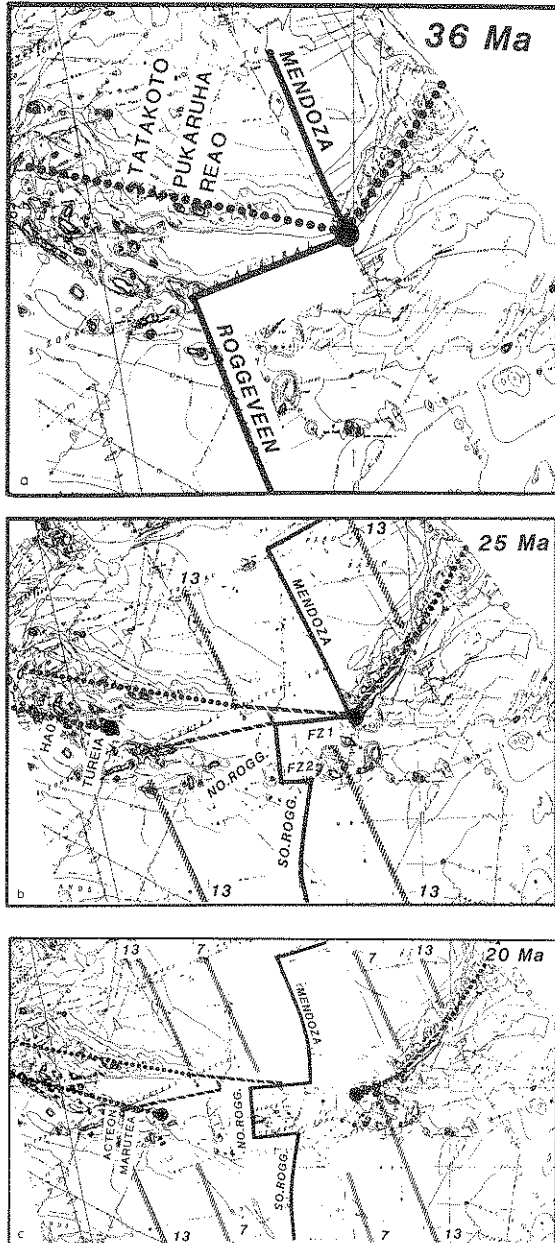


Fig. 7. (a) Reconstruction of spreading pattern at 36 Ma B.P. HS1 has reached the corner of the Mendoza Ridge and the Austral Transform Fault. Symbols common to Fig. 7a–f: solid lines: plate boundaries; dotted lines: hotspot paths; barber pole lines: pseudo-faults [45]; stippled lines: magnetic anomalies (only 13 and 7 are shown); dashed lines: fossil ridges; broken lines: hotspot leaks; large dots: positions of hotspots. Name of island or seamount outlined means formation took place since last snapshot. (b) Reconstruction of spreading pattern at 25 Ma B.P. The Mendoza Ridge has propagated south, and the Austral FZ offset has been transferred to FZ1. (c) Reconstruction of spreading pattern at 20 Ma B.P. HS1 has left the plate boundary; spreading is taking place at the intermediate azimuth. (d) Reconstruction of spreading pattern at 18 Ma B.P. The southern Roggeveen Ridge has jumped west into the southern section of the present-day East Pacific Rise; its reorientation is complete. (e) Reconstruction of spreading pattern at 15 Ma B.P. The Mendoza Ridge has jumped west into the northern section of the present-day East Pacific Rise; its reorientation is complete. HS2 has just created Oeno, and will start its interaction with FZ2. (f) Present-day location of plate boundaries and hotspots in the area, showing evolution of islands in the past 15 Ma. See text for more details.

Mendoza and Roggeveen Ridges is a “pseudo-fault” which should exist between the apparent termination of the Tuamotu plateau at 20.5°S and 131.5°W, and FZ1. Because this material was generated on-ridge, it is expected to be compensated and the pseudo-fault invisible in the geoid. At one point, the propagation eventually stopped, and the hotspot was transferred to the Farallon (or Nazca) plate. The reason for this change probably lies in the difference in velocities between the absolute motion of the Pacific plate and its accretion rate. Assuming that the spreading then stabilizes with FZ1 as an active transform, its timing can be worked out: the point at 130°W where FZ1 starts to be seen in the geoid marks the western end of the transform at the end of the rift propagating episode. Extrapolating from the mapped portion of anomaly 7 at a constant rate of 5 cm/yr, we date this point at 25 Ma, so that the rift propagation lasts a total of 11 Ma. The Mendoza Ridge then extended south to a point presently at 21.5°W, 126.2°W. FZ1 remains an active TF of the Farallon system for another 7–8 Ma. By 18 Ma B.P. (earlier south of FZ2), the ridge jumps 500 km to the west; the disappearance of the FZ1 signal to the east suggests that the jump was more important north of FZ1 than south of it, resulting in a much shortened offset.

This model explains the rugged bathymetry at Region C, evidenced by J. Francheteau’s small-scale bathymetric survey; the range of relief observed (± 400 m) is comparable to that charted along documented propagating rifts [17]. The model also explains the absence of major features in the geoid in the area, due to isostatic compensation of material generated on-ridge. It could also explain the seismic weakness of Region C, both in terms of the preferential development of seismicity at this location, and of the swarmlike behavior of the activity, concentrated at the relatively low $m_b = 5.5$ magnitude level. As for GB-8, located 65 km to the north, it is outside the pseudo-fault zone, and has more classical seismic properties: magnitude 6 seismicity and a main-shock aftershock pattern. Our model also explains why the fossil Roggeveen Ridge in the Nazca plate stops abruptly at latitude 30°S, as mentioned by Mammerickx et al. [9], since it was not active further north at the

time of the (later) ridge jump; the alternative explanation proposed by Okal and Bergeal [10] would have truncated an existing segment of the Roggeveen Ridge by a newly formed fracture zone of the new system during the jump, but would predict the existence of a well-developed Austral FZ right up to 127°W.

4. A model for the evolution of spreading in the Pacific since the Late Oligocene

In this section, we present a model for the evolution of our study area; Fig. 7a–f present snapshots of the spreading patterns and the position of the hotspots at a few critical times. These figures are superimposed on Mammerickx et al.’s [6] present-day bathymetry; however, the bathymetry of areas later subjected to ridge-jumping has been blanked out. For clarity, the names of islands and seamounts appear only once, in the first frame following their formation.

Around 40 Ma B.P., spreading was taking place along the Mendoza and Roggeveen Ridges at 5 cm/yr (half-rate); the Austral Transform Fault was a 500-km-long rightlateral offset of this system. The northern Tuamotu hotspot (HS1) was generating Pukaruha and Reao, at the most a few hundred kilometers away from the Mendoza Ridge. Assuming HS1 was fixed with respect to the Hawaiian hotspot, the motion of the Pacific plate over it was 7.9 cm/yr, 30° away from the direction of spreading, and thus HS1 was drifting southwards 3–4 cm/yr relative to the Mendoza Ridge. Thus, it reached the eastern end of the Austral Transform Fault about 36 Ma B.P. (see Fig. 7a), at a location corresponding to the mapped termination of the Tuamotu Ridge, at present-day 20°S, 131.5°W; the Austral Transform Fault then extended to present-day 21°S, 136°W, just north of Marutea and Acteon (none of these islands existed at that time).

An episode of rift propagation then started, during which the Mendoza Ridge grew south at the expense of the Roggeveen Ridge. It lasted about 11 Ma, at a propagating rate of 3 cm/yr, and at the same time involved a reorientation of the spreading and of the transform fault along

N95°E (the resulting pattern was then a mirror image of Hey's fig. 2 [45]). The propagation stopped 25 Ma B.P., and the spreading stabilized along a partially reoriented Mendoza Ridge, offset about 450 km by FZ1 (see Fig. 7b). Because of the reorientation, another TF developed along FZ2, breaking the Roggeveen in two: spreading rates at the southern segment started to accelerate (from 5 to 7 cm/yr) possibly through asymmetric spreading, but remained 5 cm/yr at the Mendoza Ridge and the northern part of the Roggeveen Ridge. FZ2 is well documented in the Nazca plate as the "Easter FZ" (see Fig. 7c). Since the Pacific plate was still moving faster (about 8 cm/yr) with respect to HS1, this hotspot eventually left the ridge system, and became off-ridge in the Farallon plate; it started generating seamounts of the "Sala-y-Gomez Ridge" (see Fig. 7c). This "ridge" clearly appears in the geoid as discrete seamounts [1], and must therefore have been formed off-ridge. The curved segments of ridges present in Fig. 7 as well as the difference of azimuth between FZ1 and transform faults to the north which may still be active in the original orientation, can be accounted for either through a gradient of asymmetry in the spreading (resulting in further curvature of the plate boundary with time), or the existence of smaller transform faults which were neither sufficiently offset nor long-lived to remain in the present bathymetric record, or the development of a short-lived platelet, which would be very hard to identify in the present record, since it would have formed during a magnetically quiet period. Another possibility would involve FZ1 as a leaky transform, a process recognized by Menard and Atwater [48] during spreading reorientations, and particularly likely in the vicinity of a hotspot.

Around 20 Ma B.P. the southern part of the Roggeveen Ridge jumped westward, possibly through a process involving an intermediate platelet, but preserved a substantial offset along FZ2 (see Fig. 7d). At 18 Ma, the Mendoza Ridge jumped westward 400 km, but the northern part of the Roggeveen Ridge (between FZ1 and FZ2) did not (see Fig. 7e). This reduced the offset of FZ1 to 100 km or less, so that its signature is absent from the geoid east of 124°W. The resulting straightened spreading system has essentially survived un-

til now as the "Oeste" Ridge or western boundary of the Easter platelet, although its activity has slowed considerably in the past 5 Ma (possibly earlier) with the growth of the "Este" system [34]. HS1 continued its activity along the Sala-y-Gomez chain, is presently in the immediate vicinity of Sala-y-Gomez Island, and generated Easter as a secondary type hotspot island [38].

Meanwhile, the southern Tuamotu hotspot (HS2) generated islands and seamounts, all off-ridge; because the Pacific plate was moving 8 cm/yr with respect to HS2, and was being accreted only 5 cm/yr at the ridge (7 cm/yr south of FZ2 after its activation), the hotspot moved closer to the ridge, has caught up with its motion north of FZ2, but not south of it. Its surface expression deviated when it encountered the inactive but fresh trace of FZ2, oriented only 15° from its trail, and it generated the Henderson-Ducie-Crough lineament. This interaction is shown by the small broken lines branching off HS2's trail in Fig. 7f. The flexural age of the plate at Crough Seamount is 5–7 Ma; at this location, the plate is 8 Ma north of FZ2, and 12 Ma south of it. It is improbable that this discrete seamount was generated along an active transform segment, where a continuous feature would be more likely, and stronger compensation would have taken place; thus, we use an average age of 10 Ma for the plate at Crough Seamount. This results in an age bracket of 3–5 Ma for Crough Seamount. In particular, an older age would require strong compensation, and would go against a possible continuation of volcanic activity to the present day. This estimate of 4 ± 1 Ma for Crough Seamount yields a present location for HS2 approximately 300 km south of the southwestern boundary of the Easter platelet (see Fig. 7f), in a poorly charted area, at the tip of a major free air gravity anomaly, suggestive of an area of deep upwelling [49]. From this model, one can work out the following ages for the islands formed by HS2: the first figure is the island's age, the second one the age of the plate on which it sits: Hao (29 Ma, 45–50 Ma); the Acteon group (23, 36); Marutea (22, 33); Oeno (16, 27); Henderson (13, 19) Ducie (8, 14); Crough Seamount (4, 10). The latter may still be active, in the form of post-erosional volcanism: a suspected eruption

would explain the large tensional earthquake of 1955.

Later, and independently, the Pitcairn hotspot created the Gambier and Pitcairn islands, loading from the latter resulting in the 30 m uplifting of Henderson. Finally, the existence of the two fractures, FZ2 and the fossil FZ1, set the stage for the later evolution of the Easter platelet: the prolongation of FZ1 may play the role of the inhibitor preventing further propagation of the East Rift to the north [34], while a possible excess of magma provided from HS2 through an asthenospheric "pipe" [38], would contribute to the initiation of spreading observed by the same authors along the southwestern boundary, itself none other than the prolongation and present-day expression of FZ2.

It is quite probable that the evolution of the Pacific-Farallon Ridge system into the East Pacific Rise was accompanied by more complex phenomena. Our proposed model attempts to provide a simple explanation of the existence of FZ1 and FZ2, as well as of the properties of the islands.

5. Conclusion

We have derived a model for the evolution of spreading during the Late Oligocene/Early Miocene in the east-central Pacific. In particular, we account for a stepwise reorientation of the ridge system, and confirm the time constraints on the ridge jump given by Mammerickx et al. [9]. We propose that the transfer of activity from the Austral FZ to the Region C FZ (FZ1) took place by rift propagation, itself triggered by the interaction of the northern Tuamotu hotspot and the fracture zone. We further propose that the islands of Oeno, Henderson and Ducie are part of the southern Tuamotu chain, and that the 15° deviation of their lineament from the Pacific plate's absolute motion was due to the interaction of this hotspot with FZ2, a feature now expressed as the southern boundary of the Easter platelet.

This raises the question of the interaction of the broad convection cells of the plate system (expressed by the ridges) with the narrower hotspot systems. The large number of active hotspots in

the vicinity of ridges has led some authors to speculate that they were capable of attracting ridges and "trapping" them. Crough's [31] model for thinning of the lithosphere under Hawaii provides one possible mechanism for this action, and Morgan [50] has proposed that Iceland indeed managed to trap the North Atlantic ridge through a succession of ridge jumps starting 35 Ma ago. On the other hand, many other hotspots have failed to do so; it is argued that some have actually left the ridge (e.g., Kerguelen). The evidence given in this paper indicates that a limited amount of interaction between hotspot and ridge did indeed take place during the Late Oligocene/Early Miocene, when HS1 distorted the pattern of spreading at the Mendoza Ridge. However, it later left the ridge, let it jump westward at least 400 km, and became a bona fide intraplate feature in the Nazca plate. Similarly, and because of a favorable geometry, the surficial expression of the path of HS2 was later deviated by FZ2. In both cases, however, the lateral extent over which the fracture zone-hotspot interaction took place is no larger than 500 km. If interpreted as representative of the lateral extent of an area of altered thermal and mechanical properties, the order of magnitude of this figure is in gross agreement with the 300 km diameter found by Hadley et al. [51] by teleseismic probing under Yellowstone; a similar figure has also been obtained by three-dimensional seismic imaging under Iceland [52].

Finally, we want to again stress the tentative and preliminary character of some of our assumptions. On the basis of our models, a number of predictions can be made concerning the absence of bathymetric expression of the Austral FZ east of 135°W, the expected bathymetry characteristic of propagated rifts between the mapped end of the northern Tuamotu plateau and Region C, the existence of FZ2, and the ages of Crough Seamount (3–5 Ma, with possible present-day activity). The present model is intended as a framework to conduct more detailed investigations of the history of spreading in the East Pacific. They will, however, have to rely on the acquisition of more marine geophysics, petrology, and geochemistry data in this important area.

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