A Plate Model for Jurassic to Recent Intraplate Volcanism in the Pacific Ocean Basin

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ABSTRACT

Reconstruction of the tectonic evolution of the Pacific basin indicates a direct relationship between intraplate volcanism and plate reorganisations, which suggests volcanism was controlled by fracturing and extension of the lithosphere. Middle Jurassic to Early Cretaceous intraplate volcanism included oceanic plateau formation at triple junctions (Shatsky Rise, western Mid Pacific Mountains) and a diffuse pattern of ocean island volcanism (Marcus Wake, Magellan seamounts) reflecting an absence of any well-defined stress field within the plate. The stress field changed in the Early Cretaceous when accretion of the Insular terrane to the North American Cordillera and the Median Tectonic arc to New Zealand, stalled migration of the Pacific- Farallon and Pacific-Phoenix ocean ridges, leading to the generation of the Ontong Java, Manahiki, Hikurangi and Hess Rise oceanic plateaus. Plate reorganisations in the Late Cretaceous resulted from the breakup of the Phoenix and Izanagi plates through collision of the Pacific-Phoenix ocean ridge with the southwest margin of the basin, and development of island arc-marginal basin systems in the northwest of the basin. The Pacific plate nonetheless remained largely bounded by spreading centres, and intraplate volcanism followed pre-existing lines of weakness in the plate fabric (Line Islands), or resulted from fractures generated by ocean ridge subduction beneath island arc systems (Emperor chain). The Pacific plate began to subduct under Asia in the Early Eocene from the record of accreted material along the Japanese margin. Further changes to the stress field at this time resulted from abandonment of the Kula-Pacific and Phoenix/North New Guinea-Pacific ocean ridges, and development of the Kamchatkan and Izu-Bonin-Mariana arcs, leading to generation of the Hawaiian chain as a propagating fracture. The final major change in the stress field occurred in the Late Oligocene as a result of breakup of the Farallon into the Cocos and Nazca plates, and caused a hiatus in Hawaiian volcanism, initiated the Sala y Gomez, Foundation, and Samoan chains, and terminated the Louisville chain. The correlations with tectonic events are compatible with shallow-source models for the origin of intraplate volcanism, and suggest that the three principal categories of volcanism, intraplate, arc and ocean ridge, all

arise from plate tectonic processes, unlike in plume models where intraplate volcanism is superimposed on plate tectonics.

Keywords: Pacific basin, Jurassic to Recent history, intraplate volcanism, accreted terranes, plate tectonics

INTRODUCTION

Before the mantle plume model was widely invoked, intraplate volcanism in the Pacific basin (Fig. 1) was attributed shallow mantle convection or propagating fractures induced by stresses acting on the plate (e.g. Jackson et al., 1972; Turcotte and Oxburgh, 1976; Jackson and Shaw, 1975; Bonatti and Harrison, 1976; Walcott, 1976). One of the most advanced explanations was the model of Jackson and Shaw (1975) who suggested volcanism along the Hawaiian chain reflected injection of magma into the lithosphere as the plate underwent rotation due to either changing forces along its boundaries or variations in coupling between the lithosphere and asthenosphere. Such models sought to understand intraplate volcanism within the framework of plate tectonics; however, with increasing popularity of the hotspot model, emphasis shifted to interpreting the intraplate volcanic record as the result of mantle plumes. Intraplate volcanism thus became imposed on plate tectonics by processes in the deep mantle or core-mantle boundary. However, after three decades of promotion of the plume model, there is little consensus with regard to the number of plumes, few examples of intraplate volcanism have been shown to conform to the predictions of the plume head-tail model, and many examples of intraplate volcanism can only be fit by invoking ad hoc variations to the plume model (e.g. Clouard and Bonneville, 2001; Anderson, 2005). Even then the plume model constitutes only a partial explanation for intraplate volcanism, as it cannot account for the vast majority of seamounts across the ocean basin (Natland and Winterer, 2005), or volcanism in areas of plate fracturing remote from postulated hotspots (Hirano et al., 2006). Instead of seeking alternative explanations only when a plume model cannot be fit, it should be asked whether plate tectonics or “plate model” (Anderson, 1998; Foulger and Natland, 2003; Foulger, this volume) could provide a comprehensive explanation for intraplate volcanism, rather than having two mechanisms for the generation of one category of volcanism.

In the plate model, subducted oceanic crust is remixed with the convecting mantle as in the statistical upper mantle assemblage (SUMA) model of Meibom and Anderson (2003) and streaky mantle model of Smith (2005), rather than being isolated in plume sources. The same recycled geochemical components are present as in plume models, but their distribution and sampling is different. Large regions of the shallow mantle are close to, or at the the peridotite solidus, and ocean island basalt-like melts are pervasive in the asthenosphere. The location of intraplate volcanism is controlled by extension or fracturing of plates which allows melt to be released (Anderson, 1998; Favela and Anderson, 2000; Foulger and Natland, 2003; Foulger, this volume; Stuart et al., this volume). The volume of melt is controlled by the size, percentage and orientation of streaks of recycled crust in the region of mantle being tapped (Meibom and Anderson, 2003; Maclennan, 2004; Smith, 2005). Fracturing is related to the stresses acting on the plate, which include thermal stress from cooling, loading stresses from the presence of topographic features, and stresses imposed from plate interactions along the margins of the plate (e.g. Hieronymus and Bercovici, 2000; Natland and Winterer, 2005). As boundary stresses may be transmitted great distances (e.g. Scotese et al., 1988), the plate model predicts a relationship between intraplate volcanism and the tectonic record. Correlations between the volcanic record and events around the basin margin have been suggested for the Hawaiian-Emperor chain (Jackson et al., 1972, Jackson and Shaw, 1975; Smith, 2003), and for oceanic plateaus in the mid-Cretaceous (Filatova, 1998; Tagami and Hasebe, 1999), although the mid-Cretaceous event has only been considered with the prior assumption of a plume origin. The aim of this paper is to examine how the entire record of intraplate volcanism in the Pacific basin could be interpreted in a plate model. Following Natland and Winterer (2005), the volcanic record has been divided into three stages, Middle Jurassic to Early Cretaceous, Late Cretaceous to Eocene, and Eocene to

Recent, and the relationship with plate configuration examined to determine the role of plate tectonic processes at each stage.

RECONSTRUCTION OF INTRAPLATE VOLCANISM IN THE PACIFIC BASIN

Previous plate reconstructions of the Pacific basin have mainly been based on the hotspot reference frame. The latter is not valid in a plate model because plumes are not required in the model, and volcanism can occur along pre-existing lines of weakness in the plate fabric, independent of the direction of plate movement. The tectonic history and record of intraplate volcanism in the basin has thus been reconstructed (Fig. 2) from the present day ocean floor topography (Smith and Sandwell, 1997), plate reconstructions based on seafloor magnetic anomalies (Scotese et al., 1988), paleomagnetic data for the Emperor chain and Ontong Java plateau, and the record of intraplate volcanism in accreted terranes around the basin margin.

MIDDLE JURASSIC TO EARLY CRETACEOUS

Plate Configuration at the Time of Formation of the Pacific Plate

The Pacific plate originated in the central paleo-Pacific basin between 175 and 170 Ma in the Middle Jurassic, soon after the initial stages of opening of the Atlantic Ocean (Bartolini and Larson, 2001) and coincident with the closure of the Mongol-Okhotsk basin in the west of the basin (Nokleberg et al., 2000). The Pacific plate was bordered by the Izanagi, Farallon, and Phoenix plates to the west, east and south, respectively, although the orientations of ridge systems at this stage are speculative (Fig. 2a). In the Middle Jurassic, the Izanagi, Farallon and Phoenix plates were of approximately equal size, but the occurrence of Tethyan faunas in older accreted oceanic assemblages such as the Cache Creek terrane of western Canada (e.g. Orchard et al., 2001) (Fig. 1), implies an earlier plate configuration dominated by a large eastward-subducting Farallon plate to allow for transport of terranes across the paleo-Pacific basin.

At the time of formation of the Pacific plate, the Farallon plate subducted beneath the Insular (super)terrane to form the Gravina-Nutzotin-Gambier arc in the northeast of the basin (Nokleberg et al., 2000). The Insular terrane is a composite of the Alexander terrane which comprises crust rifted from eastern Gondwana in the mid-Paleozoic (Gehrels and Saleey, 1987), and Late Paleozoic-Mesozoic arc strata overlain by intraplate basalts (Karmutsen-Nikolai Formation) of the Wrangell terrane (Fig. 1). In plate models the Karmutsen-Nikolai Formation can be interpreted as an arc-rift assemblage (Barker et al., 1989). The Insular terrane was separated from North America by the Bridge River Ocean, which began to close in the Middle Jurassic as a result of westward movement of North America on opening of the Atlantic. The existence of an island arc-marginal basin system to the south of the Insular terrane in the Middle Jurassic (Fig. 2a), is suggested by strata in the Vizcaino peninsula of western Mexico (Moore, 1985). Strata in the Guerrero terrane of Mexico and the Western Cordillera of Colombia indicate the arc system off Central America was well established by the Late Jurassic-Early Cretaceous (Tardy et al., 1994). However, the Vizcaino-Guerrero arc system is unlikely to have extended south of the latitude of the Arica elbow in South America, where subduction of the Farallon plate beaneath the continental margin began to generate the Early Andean Magmatic Province in the Middle Jurassic (e.g. Oliveros et al., 2006).

The Guerrero-Vizcaino arc was separated from North and Central America by oceanic crust of the Arperos basin (Freydier et al., 1996) (Figs. 2a,b). The relationship of the Vizcaino-Guerrero arc to the Insular terranes is unknown, on account of uncertainties in the width of the Arperos basin and the paleolatitude of the Insular terrane. The debate regarding the latter has been summarised by Cowan et al. (1997): paleomagnetic evidence suggests accretion of the Insular terrane at a latitude of 30^0 to 40^0 N, but a lack of evidence for suitable fault systems along which subsequent northward displacement could have could have taken place, has led to suggestions of accretion at present day latitudes. The position for the Insular terrane illustrated in figure 2 follows the model of Butler et al. (2001) which advocates moderate post accretion latitudinal displacements of 1000 km, and results in the Insular terrane lying approximately 10^0 north of the

Guerrero arc. In the Arperos basin, the intraplate record begins in the Early Cretaceous, with intraplate basaltic rocks in central Mexico interpreted as fragments of accreted ocean islands or an aseismic ridge (Ortiz and Martinez, 1993; Freydier et al., 1996). Early Cretaceous basaltic rocks of the Amaime Formation of Colombia, and Pallatanga and San Juan groups of western Ecuador (Lapierre et al., 2000; Kerr et al., 2002), and in the Nicasio terrane of the western United States (Silberling et al., 1987), may also have originated in the Arperos basin (Fig. 2b).

Plate reconstructions for the northwest Pacific basin in the Middle Jurassic to Early Cretaceous have portrayed the Farallon-Izanagi ridge migrating northwards along the continental margin of Asia, followed by the accretion of oceanic plateaus, seamounts and microcontinent blocks from the Izanagi plate (e.g. Maruyama and Seno, 1986; Isozaki et al., 1990; Kinoshita, 1995). The microcontinents include the crustal blocks bordering the South China Sea (Reed Bank, Macclesfield Bank, Palawan), the Kurosegawa and Abukuma-South Kitakami belts of Japan, and the Tahin terrane of Sikhote-Alin, which were derived from the Tethyan region in the Late Paleozoic (Maruyama et al., 1989; Metcalfe, 1996). However, evidence for accreted intra-oceanic arc remnants in Hokkaido (Oku-Niikappu complex), suggests the northwest Pacific basin was more like the present day Philippine Sea in containing a series of arc and marginal basin systems (Ueda and Miyashita, 2005). The latter model is supported by the occurrence of blocks of volcanic rocks of arc origin in the mid-Cretaceous Yuli belt accretionary complex of Taiwan (Sun et al., 1998), and by Jurassic-Early Cretaceous arc volcanic rocks on the Academy of Sciences Rise and in the Kvakhon terrane along the eastern margin of the Okhotsk block (Watson and Fujita, 1985). The Academy of Sciences Rise-Kvakhon arc must have lain at low latitudes on account of Tethyan faunas in Late Jurassic to Early Cretaceous limestones which cap guyots in the Nabilsky terrane (Nokleberg et al., 2000) along the suture between the Eurasian continent and the Okhotsk block. The southwest continuation of the Academy of Sciences Rise-Kvakhon arc is suggested to be the Oku-Niikappu arc of Ueda and Miyashita (2005), remnants of which were accreted to Hokkaido in the mid-Cretaceous. Intraplate volcanism within the Middle-Late Jurassic marginal basins is represented by basalts in the Taukha terrane of Sikhote-Alin, Tamba terrane of Japan, and Anvina terrane of Sakhalin (Nokleberg et al., 2000; Koizumi and Ishiwatari, 2006) (Fig. 2a).

The terrane record on New Zealand also suggests the existence of an island arc-marginal basin system along the southwest margin of the Pacific basin in the Middle Jurassic to Early Cretaceous (Bradshaw, 1989). Remnants of the Mesozoic arc, which is thought to have been constructed along the western margin of an earlier Permian arc (Brook Street arc; Bishop et al., 1985), are found along the Median Tectonic Zone of New Zealand (Kimbrough et al., 1993; Muir et al., 1995).

Oceanic Plateaus and Diffuse Island Chains

The oldest intraplate volcanism on the Pacific plate is represented by the Early Cretaceous Shatsky Rise and western Mid-Pacific Mountains oceanic plateaus, and seamounts in the Magellan and northern Marshal groups (Fig. 2b). The Shatsky Rise formed in conjunction with a triple junction between the Pacific, Izanagi and Farallon plates over a time period of 16 million years between 143 and 127 Ma (Nakanishi et al., 1999; Sager, 2005), and the chemistry of the volcanism is MORB-like (Mahoney et al., 2005). The Mid-Pacific Mountains has a complex morphology suggesting it is comprised of a series of ridges and seamounts (Winterer and Sager, 1995; Natland and Winterer, 2005). Radiometric ages for the Mid-Pacific Mountains range from 128 to 73 Ma (Clouard and Bonneville, 2005) indicating a protracted evolution. As only the central and eastern parts have been dated, volcanism may have overlapped formation of the Shatsky Rise, and have formed in a triple junction setting with respect to the Pacific-Farallon and Pacific-Phoenix ridges (Fig. 2b). Splayed ridges on the western (Early Cretaceous or older) and eastern (Late Cretaceous) section of the Mid-Pacific Mountains have no counterpart in current intraplate volcanism in the Pacific basin, and their origin is unknown (Natland and Winterer, 2005). Linear east-west, and northwest-southeast trending age progressions have been suggested for the Magellan and Marshall seamounts in plume models (e.g. Koppers et al., 2001). However, other studies have noted a lack of a consistent paleolatitudinal pattern as would be expected from a plume origin, and have argued that the volcanism has no systematic age distribution as a result of the plate lacking a well-defined stress field (Sager et al., 1993; Natland and Winterer, 2005).

Basaltic sequences of the Kamuikotan Complex in the Late Jurassic-Early Cretaceous Sorachi terrane of Japan have been suggested to represent a conjugate oceanic plateau to the Shatsky Rise (Kimura et al., 1994; Kimura, 1997). However, the Sorachi terrane had accreted to the continental margin by 119 Ma (Kimura et al., 1994), which would require a very high rate of plate movement of 25 cm yr^{-1} for transport from a central Pacific location. An origin for basalts in the Sorachi terrane as a result of fore-arc rifting has also been proposed from the presence of arc volcanic rocks within the terrane (Takashima et al., 2002). The position of the Sorachi terrane adjacent to the Oku-Niikappu complex raises the possibility of a relationship to the rifting of the Oku-Niikappu arc described by Ueda and Miyashita (2005) (Fig. 2b). An oceanic plateau (Okhotsk arch) has also been suggested to underlie the north-central Sea of Okhotsk (Watson and Fujita, 1985), and could potentially be linked with volcanism in the Sorachi terrane as part of rifting of the arc system separating the Jurassic marginal basins from the Izanagi plate.

Formation of the Ontong Java plateau also likely occurred in conjunction with an ocean ridge system (Hussong et al., 1979). Lineations on the plateau have been suggested to correspond to both southwest-northeast (Winterer, 1976; Winterer and Nakanishi, 1995; Gladczenko et al., 1997) and northwest-southeast (Neal et al., 1997) trending ridge axes. When plume models were developed for the plateau, age data were available for only three drill sites and appeared to indicate formation of the main plateau from 124 to 121 Ma, followed by volcanism on the eastern lobe at 90 Ma (Mahoney et al., 1993; Neal et al., 1997). The narrow range of ages, along with correlations with sites in nearby basins and sections on Malatia and Santa Isobel islands, were used to infer rapid eruption (e.g. Tarduno et al., 1991; Coffin and Gahagan, 1995); however, if the ridge system had been orientated northwest-southeast, the sampling localities would have lain close to the ridge axis and formation of the plateau could have taken place over a longer timescale

(Smith, 2003). Recent Ar-Ar dating of new drill sites from ODP Leg 192, discussed in Fitton et al. (2004), has yielded ages of 122-105 Ma. Although the younger ages have been suggested to be an artefact of analysis, which would also negate the 90 Ma episode of volcanism (Fitton et al., 2004), paleontological evidence suggests eruption continued to 112 Ma. The model suggested in figure 2c is thus of eruption of both the main plateau and eastern lobe over northwest-southeast and east-west trending segments of the Pacific-Phoenix ridge, over a time span of approximately 10 m.y.

The Hess Rise (110-100 Ma) also formed on an ocean ridge (Pacific-Farallon), and has been suggested to result from confinement of ridge upwelling by transform faulting along the Shatsky and Mendocino fracture zones (Windom et al., 1981; Rea and Dixon, 1983). The tectonic setting of the Manahiki plateau (123-110 Ma) is uncertain, although the structure of the plateau has been related to a jump of the Pacific-Farallon-Phoenix (Tongevara) triple junction, which intersected the plateau during its early stages of formation (Joseph et al., 1993; Viso et al., 2005). Implied in such models is a plume origin whereby the plateau was already forming as a result of hotspot activity at the time of the triple junction jump. An eastern conjugate to the Manahiki plateau (Mollendo ridge) was postulated to have subducted under South America from a Late Cretaceous magmatic gap in the Peruvian Andes (Soler et al., 1989). Similarly, the Hikurangi plateau, which lies off the New Zealand margin (Mortimer and Parkinson, 1996), has been suggested as a southern conjugate of the Manahiki plateau (Sutherland and Hollis, 2001). Rifting of the Manahiki and Hikurangi plateaus is considered to have occurred along the Osbourn Trough, a slow spreading ridge that now lies nearly equidistant from the plateaus in the southwest Pacific (Billen and Stock, 2000). Recent models have suggested the Ontong Java-Manahiki-Hikurangi plateaus as part of a single large igneous province, which was rifted at approximately 112 Ma by a triple junction west of the Osbourn Trough (Taylor, 2006; Worthington et al., 2006).

Mid-Cretaceous Plate Reorganisation

The generation of oceanic plateaus in the Early Cretaceous coincided with a protracted phase of plate reorganisation along the eastern and southwestern margins of the Pacific basin. Accretion of the Insular terrane to western North America began in the Late Jurassic (McClelland et al., 1992) and involved closure of basins now represented by the Hozameen, Bridge River and Methow-Tyaughton terranes (Monger and Berg, 1987). Final closure of marginal basins did not occur until around 100 Ma, however, from the timing of arc activity in the Spences Bridge Group along the western margin of North America (Thorkelson and Smith, 1989). Closure of the Arperos basin also took place in the Early Cretaceous, with the island arc system along the west of the basin accreting to form western Mexico and Central America (Fig. 2d). The oldest volcanism in the Gravina-Nutzotin-Gambier arc on the Insular terrane arc occurred at 110 Ma (Nokleberg et al., 2000), with subduction re-establishing under the continental margin to generate the Coast Plutonic Complex, Sierra Nevada, Salinia and Peninsular Ranges batholiths in the Late Cretaceous (e.g. Butler et al., 2001). Subduction of the Farallon plate was therefore disrupted between 110 and 100 Ma, corresponding to the timing of formation of the Hess Rise. Similarly, accretion of the island arc system in the southwest of the Pacific basin to the New Zealand margin (Rangitata orogeny), began in the Late Jurassic and terminated during the Early Cretaceous (Muir et al., 1995). The interruption of subduction of the Phoenix plate from 130 to 110 Ma as a result of the accretion event (Spörli and Balance, 1989), corresponds to the timing of formation of the Ontong Java, Manahiki and Hikurangi plateaus.

In contrast to the re-organisations along the northeast and southwest margins, there was a greater continuity of tectonic regime in the northwest and southeast Pacific basin. The South American and Asian margins both experienced strike-slip faulting in the mid-Cretaceous, but subduction continued beneath arc systems in the northwest Pacific basin and along the Peru-Chile margin (Watson and Fujita, 1985; Taira et al., 1989; Osozawa, 1998; Scheuber and Gonzalez, 1999). The evidence from the North American and New Zealand margins thus suggests formation of the mid-Cretaceous oceanic plateaus may have been linked to the interruption of subduction causing migration of ocean ridge systems bounding the subducting plate to slow or stall. Large volumes of melt could potentially be generated from entrainment of large streaks of recycled crust in the convecting mantle into the ocean ridge upwelling (Meibom and Anderson, 2003; Smith, 2005; Korenaga, 2005). The slowing of ridge migration would result in the excess melt no longer being accommodated by the formation of normal oceanic crust, thereby leading to the generation of an oceanic plateau. The generation of the Ontong Java, Manahiki, Hikurangi and Hess Rise plateaus may therefore mark the first major change in the stress field of the Pacific plate. Pinning of the Pacific plate by the jamming of subduction zones along the northeast and southwest margins of the basin would also explain the lack of movement of the plate suggested by paleomagnetic data, rather than having to invoke pinning of the plate by plumes as suggested by Tarduno and Sager (2002).

LATE CRETACEOUS TO EARLY EOCENE

The Fate of the Pacific-Phoenix Ridge and Onset of Subduction of the Pacific Plate

In the mid-Cretaceous, the Farallon plate separated into north and south parts along the Mendocino fracture zone, and the Pacific-Farallon South ridge became orientated north-south. The Pacific plate extended to the southeast by spreading along the Osbourn Trough and/or a series of ocean ridge jumps (Joseph et al., 1993; Worthington et al., 2006). Expansion of the Pacific plate westwards from the Early Cretaceous reconstruction would also have resulted in the Izanagi-Pacific ridge colliding with the margin of Asia, and the onset of subduction of the Pacific plate. The latter event was suggested to have occurred around 83 Ma by Uyeda and Miyashiro (1974), from extrapolation of the Kula-Farallon ridge configuration of Larson and Chase (1972). The fate of the Pacific-Phoenix ridge was not considered by Uyeda and Miyashiro (1974), but presumably must have migrated southwards along the western Pacific margin in such models. Subduction of the Pacific-Phoenix ridge was suggested to occur at 105 Ma in the southwest of the basin, followed by rifting of Zealandia (New Zealand, Campbell Plateau, Chatham Rise, Lord Howe Rise, and Norfolk Ridge) from eastern Gondwana in the early Late Cretaceous (Bradshaw, 1989). Other models suggest the Pacific-Phoenix ridge was abandoned close to the continental margin, with rifting of Zealandia resulting from the subducted slab of the Phoenix plate and overlying crustal blocks being captured by, and moving with the Pacific plate (Luyendyk, 1995). The western margin of the basin may thus have been characterised by transform faulting or extension, possibly with ocean ridge systems separating the Pacific, Australian, and Indian Ocean plates (Wells, 1989; Coney, 1990).

Models for subduction of the Pacific plate in the Late Cretaceous are consistent with evidence from accreted fragments in subduction zone assemblages in Japan and Sakhalin, for the passage of the Izanagi/Kula-Pacific ocean ridge system northwards along the continental margin from 95 to 40 Ma (e.g. Kimura et al., 1992; Kinoshita, 1995). However, a second ridge system was proposed by Osozawa (1992, 1997) to have intersected the Kyushu margin at 50 Ma and migrated southwards. This second ridge was suggested to be the North New Guinea-Pacific ridge along the northern margin of the North New Guinea plate. The latter plate was hypothesized from the Cenozoic tectonics of the Philippine Sea by Seno and Maruyama (1984), on account of inferred northerly motion of the Pacific plate in the hotspot reference frame being incompatible with subduction beneath the northern margin of the Philippine plate in the Early Eocene. The argument regarding the motion of the Pacific plate does not apply in the plate model, but the record of convergence in Papua New Guinea and the Philippines supports the existence of the North New Guinea plate. Northern New Guinea represents an island arc beneath which seafloor north of southern New Guinea was subducted in the Late Cretaceous to Early Miocene (Seno and Maruyama, 1984; Cullen and Pigott, 1989). Similarly, the eastern Philippines and Daito ridge of the Philippine plate represent a Late Cretaceous to Paleocene arc, to which the western Philippines were accreted from the Australian plate in the Early Eocene (Honza and Fujioka, 2004). Locating such arcs along the southern margin of the Pacific plate would have resulted in accretion to southern Asia. The occurrence of widespread boninitic volcanism along the Izu-Bonin-Mariana arc in the Early Eocene is also consistent with the presence of a ocean ridge system in the western Pacific supplying young, thermally anomalous oceanic crust parallel to the arc system during its early stages of evolution. Without the North New Guinea plate, a mantle plume has to be invoked to provide the thermal anomaly as in the model of Macpherson and Hall (2001).

The name of the North New Guinea plate reflects its geographical location, but implies it represents a new plate formed after the Pacific plate had begun to subduct along the western margin of the basin. Rather than invoking a new plate, it is suggested that the North New Guinea plate was a remnant of the Phoenix plate (Fig. 2d). The ocean ridge migration sequence along the Japanese margin may thus reflect the collision of Izanagi-Pacific-Phoenix triple junction with the continental margin, with the implication that subduction of the Pacific plate under Asia did not begin until the Early Eocene (Osozawa, 1992) (Fig. 2e,f). The model also raises the possibility of a link between the tectonic evolution of the North New Guinea plate and rifting of Zealandia. Opening of the New Caledonia basin occurred around 90 Ma, with separation of the Norfolk Ridge from the Lord Howe Rise (Sdrolias et al., 2003). If the spreading system in the New Caledonia basin was an extension of the North New Guinea-Pacific ridge, and the New Caledonia basin was linked with the Osbourn Trough by faulting across the southern Norfolk Ridge, the Pacific plate may have been bounded almost entirely by ocean ridge systems in the early Late Cretaceous (Fig. 2d). The Pacific plate would thus have continued to lack a well-defined stress direction as in the Middle Jurassic-Early Cretaceous, hence the continuation of diffuse ocean island volcanism in the Marcus Wake, Magellan, and Marshall islands through the early Late Cretaceous.

Late Cretaceous ocean island volcanism in the Line Islands chain was more linear than in the Marcus Wake, Magellan and Marshall islands, but lacked any linear age progression and was cut by several cross trends (Natland and Winterer, 2005). Similarities in the orientation of the Line Islands chain to earlier spreading centres on the Pacific-Farallon ridge were noted by Winterer (1976), and the volcanism can be explained by reactivation of earlier lines of weakness in the plate fabric (Davis et al., 2002). A prominent lineation orientated N32^{0}W in the central, mid-

Cretaceous section of the Mid-Pacific Mountains, parallels the Line Islands trend and is also interpreted to result from reactivation of an earlier abandoned spreading centre. Limited dating of volcanism in the eastern Mid-Pacific Mountains suggests volcanism may in part have been orthogonal to the Pacific-Farallon South ridge, and thus related to thermal contraction of the oceanic lithosphere (Natland and Winterer, 2005).

Early Campanian Plate Reorganisation and Breakup of the Izanagi Plate

A major plate reorganisation occurred in the Pacific basin at 84-82 Ma (Early Campanian), which included splitting of the Kula plate from the Farallon North plate (Woods and Davies, 1982). Volcanism along the Emperor chain was initiated at this time from dating of Detroit seamount (Keller et al., 1995), which suggests an age of greater than 81 Ma for Meiji seamount at the northern tip of the chain (Fig. 3). In hotspot models, the orientation of the Emperor chain is used to infer northward motion of the Pacific plate between 81 and 50 Ma. Such motion has been attributed to northward subduction of the Kula plate faster than spreading on the Kula-Pacific ridge, imparting the motion of the Kula plate to the Pacific plate (Lonsdale, 1988a). The Kula plate as depicted by Lonsdale (1988a), includes the Izanagi plate in the northwest of the Pacific basin, and therein lies a difficulty as reconstructions of the northwest Pacific basin from the terrane accretion record, suggest the Izanagi plate began to fragment with development of the Kronotskaya and Achaivayam-Valaginskaya island arcs between 89-83 Ma (Konstantinovskaia, 2000, 2001) (Fig. 2d). The timing corresponds to final docking of accreted terranes now under the Sea of Okhotsk (Soloviev et al., 2006), such that breakup of the Izanagi plate may have been precipitated by the plate reorganisation along the basin margin (Fig. 2e). However, regardless of the cause of breakup, the reduction in size of the Izanagi plate in conjunction with subduction of young oceanic crust beneath the Kronotskaya arc, would likely have slowed spreading on the Pacific-Izanagi ridge, making it unlikely that motion of the Pacific plate was controlled by plates in the north of the basin.

A further difficulty with hotspot models is that paleomagnetic evidence indicates formation of the Emperor chain at least 15^0 north of the present day location of the supposed Hawaiian plume (Tarduno and Cottrell, 1997; Sager, 2002, Doubrovine and Tarduno, 2004). The paleomagnetic position of 43^0N for Detroit seamount (Sager, 2002), lies close to the position of the Pacific-Farallon-Izanagi triple junction in the plate reconstruction of Rea and Dixon (1983). The proximity of initial Emperor magmatism to the Kula-Pacific ridge has been suggested from other plate reconstructions (Mamerickx and Sharman, 1988), guyot morphology (Caplan-Auerbach et al., 2000), and the geochemical signatures of the oldest lavas in the Emperor chain (Keller et al., 2000; Regelous et al., 2003). Meiji seamount may thus have been generated in a triple junction setting between the Pacific, Izanagi and Farallon North plates, with volcanism interrupted by the plate reorganisation at 84-82 Ma such that the edifice did not reach the size attained by other oceanic plateaus (Smith, 2003).

The Pacific-Kula/Izanagi ridge configuration pertaining to the initiation of the Emperor chain has been refined by Norton (this volume), who suggests the triple junction associated with formation of Meiji seamount jumped to a position northeast of the Hess Rise, with the Emperor Trough acting as a sinistral transform fault between the old and new triple junctions (Fig. 2d). The Pacific-Kula/Izanagi spreading centre then re-orientated east-west and began to collide with the Kronotskaya arc. Following the reconstructions of Konstantinovskaya (2001) and Norton (this volume), the Emperor chain is envisaged to result from tearing of the Pacific plate, due to a section of the Pacific-Kula/Izanagi ridge becoming trapped outboard of the Kronotskaya arc in the east of the basin. The Pacific plate would therefore have subducted under the western Kronotskaya arc, whereas the Izanagi/Kula plate subducted under the eastern section of the arc (Fig. 2e). Weakening of the Pacific plate by lithospheric loading from Meiji seamount may also have facilitated initiation of the Emperor chain. Ocean island fragments in the subduction assemblage of the Kronotskaya arc (Watson and Fujita, 1985) may be related to similar plate fracturing by convergent margin geometry. The timing of volcanism in the Kronotskiy terrane is constrained only from Late Cretaceous to Oligocene. However, as the island fragments now lie at the same latitude as Meiji and accreted around 10 Ma (Fig. 3), they may have lain approximately 900 km to the west of Meiji from current plate velocities and convergence trends, and are speculated to have been related to subduction of a more-westerly segment of the Kula-Pacific ridge segment beneath the Kronotskaya arc (Fig. 2e).

Major plate reorganisations also took place in the south of the Pacific basin in the mid Late Cretaceous, including the abandonment of the Osbourn Trough following accretion of the Hikurangi plateau to the Chatham Rise (Worthington et al., 2006). As a result of this event, the Pacific plate extended southwards by the capture of the Hikurangi and Moa plates between the Osbourn Trough and Antarctica at approximately 82 Ma (Sutherland and Hollis, 2001) (Fig. 2d). The Tasman Sea also began to open at 85-80 Ma by spreading along the Tasman ridge which was linked to the Pacific-Antarctic ridge by transform faulting (Kamp, 1986; Schellart et al., 2006). The Pacific plate may have been bounded by arc systems along the New Guinea-New Zealand margin at this time; however, if the Tasman ridge continued into the North New Guinea-Pacific ridge, Zealandia may have belonged to the Pacific plate until return of the crustal blocks to the Australian plate on abandonment of the Tasman ridge in the Early Eocene (Fig. 2e,f).

The Louisville chain has been related to the Eltanin fracture zone in several models (Hayes and Ewing, 1971; Larson and Chase, 1972; Watts et al., 1988). Seafloor magnetic anomalies suggest formation of the fracture zone as a result of the Early Campanian plate re-organisations (Fig. 4). Only Eocene to Oligocene volcanism along the chain extrapolates directly into the fracture zone, but the proximity of the Louisville chain to the Osbourn Trough (Figs. 1, 4), raises the possibility that early volcanism along the chain was related to the triple junction responsible for rifting of the Ontong Java and Manahiki plateaus in the model of Worthington et al. (2006) (Fig. 2d). As for Meiji seamount, the ocean ridge system may have been abandoned during the early stages of Louisville volcanism, which may then have propagated to the southeast toward fracture zones along the Pacific-Antarctic ocean ridge close to the Chatham Rise, prior to reorganisations along the spreading centre in the Early Eocene which resulted in control by fracturing along the Eltanin fracture zone.

Farallon and Caribbean Plates

Late Cretaceous to Early Eocene intraplate volcanism on the Farallon and Caribbean plates includes Paleocene seamounts of the Crescent terrane of the northwestern USA (Duncan, 1982), the Caribbean plateau, and oceanic assemblages in the Western Cordillera and coastal provinces of Colombia and Ecuador (e.g. Kerr et al., 1997; Reynaud et al., 1999). Seamounts of the Crescent terrane were generated when the Kula-Farallon and Pacific-Farallon ridges were in close proximity to the North America, and have been suggested to result from rifting related to migration of the Kula-Farallon ridge along the continental margin (Babcock et al., 1992). The origin of the Caribbean plateau has generally been considered to involve generation over the Galapagos hotspot at approximately 90 Ma, followed by collision of the plateau with the east-dipping Caribbean Great Arc some ten million years later (e.g. Kerr et al., 1997; Sinton et al., 1998). Subduction then has to switch to the eastern margin of the plateau as it moves between North and South America. However, difficulties with the accretionary model for the Caribbean plateau were pointed out by James (2006), who argues for an insitu origin. The latter is adopted in this study, with the Caribbean plateau depicted as forming in a back-arc setting behind the Caribbean Great Arc (Fig. 2d).

Accreted intraplate volcanic rocks in coastal Ecuador (Piñón Formation), on Gorgona Island, and in the Serranía de Baudó province of Colombia were originally considered part of the Caribbean province, but were later attributed to the Sala y Gomez hotspot (e.g. Kerr and Tarney, 2005). The occurrence of komatiites on Gorgona Island has been considered strong evidence for a plume origin (e.g. Arndt et al., 1997). However, generation of komatiite melts under hydrous melting regimes in supra-subduction zone environments similar to those invoked for boninites, has been proposed by Parman and Grove (2005). Volcanism in the Gorgona province, as for the

Kamuikotan Complex of the Sorachi terrane, and the Karmutsen-Nikolai Formation of the Wrangell terrane, may thus have a common origin related to arc-rifting rather than core-mantle boundary processes.

EARLY EOCENE TO RECENT

Early Eocene Plate Reorganisation

The bend between the Hawaiian and Emperor chains has been considered to mark a change in motion of the Pacific plate from northerly to west-northwest in plume models, in response to either the collision of India with Asia (e.g. Clague and Dalrymple, 1989), or the onset of subduction of the Pacific plate beneath the Philippine plate (Seno and Maruyama, 1984). However, it has been argued that Tethyan events would have little effect on motion of the Pacific plate (Richards and Lithgow-Bertellini, 1996), and ages of 49-47 Ma for the earliest boninitic magmatism in the Izu-Bonin-Mariana arc (e.g. Cosca et al., 1998), appeared to indicate the onset of subduction beneath the Philippine plate before the earlier accepted age of 43 Ma for the Hawaiian-Emperor bend. Other studies pointed out a lack of corresponding tectonic events around the basin margin that would have been expected had plate motion changed at 43 Ma (Herron, 1972; Norton, 1995). Interactions with the Philippine plate have been reconsidered in plume models, from re-evaluations of the age of Hawaiian-Emperor bend. An age of ~47 Ma for the bend was suggested by Sharp and Clague (1999) corresponding to the age of Daikakuji seamount which may be considered the westernmost expression of the Hawaiian chain, but was later increased to 50 Ma (Sharp and Clague, 2006) from extrapolation of the age of 47.9 Ma for post-shield alkaline lavas on Kimmei seamount. The uncertainties in the timing reflect a degree of non-linearity in the age progressions, along with sampling of only the later stages of volcanism on the edifices. However, even if the initiation of the Hawaiian-Emperor bend and the Izu-Bonin-Mariana arc were contemporaneous, slab pull and ridge push forces would be unlikely to change rapidly enough to cause a change in plate direction (Favela and Anderson, 2000). Rather, the continuity of magnetic lineations along the Pacific-Farallon ridge (Foulger and Anderson, 2005), suggests there was no change in Pacific plate motion at either 43 or 50 Ma.

The stress field of a plate would however respond rapidly to a change in plate interactions. In a plate model, the development of the Izu-Bonin-Mariana arc, which was likely associated with abandonment of the North New Guinea-Pacific ridge (Fig. 2f), represents only one of three major tectonic changes in the mid Eocene. In addition to the potential Early Eocene onset of subduction of the Pacific plate under the continental margin of Asia (Fig. 2e,f), the stress field of the Pacific plate may also have changed as a result of plate interactions in the northwest Pacific basin (Smith, 2003; Shapiro et al., 2006). At approximately 46 Ma, subduction of the the Iruney-Vatuna basin ceased and the polarity of subduction under the Acahivayam-Valaginskaya flipped as as a result of collision of the arc with the continental margin (Konstantinovskaia, 2000; Shapiro et al., 2006) (Fig. 2e). The change in subduction polarity also coincides with, or followed soon after the cessation of activity on the Izanagi/Kula-Pacific ridge (Byrne, 1979). Thereafter the Izanagi and Kula remnants became part of the Pacific plate which subducted to form the central Kamchatkan volcanic belt. The effect of the Early Eocene events along the margins of Kamchatka and the Philippine plate would have been to change the stress field in the west of the Pacific basin from compressional- to extensional- perpendicular to plate motion, which would have allowed formation of the Hawaiian chain as a propagating fracture (Stuart et al., this volume).

Late Oligocene and Miocene-Pliocene Plate Reorganisations

After the Early Eocene plate reorganisation, volcanism along the Hawaiian chain was sparse until Midway Island at 28 Ma, and then shows a hiatus between 27 and 21 Ma west of the Murray fracture zone (Norton, 1995) (Fig. 3). Volcanism then shifts to slightly different trend with a different pole of rotation (Epp, 1984; Koppers et al., 2001) until approximately 5 Ma. The increase in volcanic output at the Murray fracture zone may indicate structural control, and the hiatus which precedes it coincides with Late Oligocene basin-wide tectonic events including the division of the Farallon into Cocos and Nazca plates at 25-23 Ma (Tebbens and Cande, 1997; Lonsdale, 2005), reorganisations along the Antarctic-Pacific ridge (Kamp, 1991), and the initial opening of the Japan Sea (28 Ma; Jolivet et al., 1995). The bathymetry of the Hawaiian chain shows a further break between the islands of Nihoa (7 Ma) and Kauai (5 Ma), which corresponds to the timing of plate re-organisations along the Fiji margin (Cox and Engebretson, 1985). The Late Oligocene and Miocene-Pliocene events have similar significance in other island chains on the Pacific plate. Volcanism along the Samoan chain (23-0 Ma) lies in a region of deformation of the Pacific plate opposite where the margin changes from subduction along the Tonga trench to a transform boundary along the Fijian margin (Natland and Winterer, 2005). The Cook and Foundation chains were initiated during the Late Oligocene event, and volcanism in the Austral and Marquesas islands began during the Miocene-Pliocene event (Natland and Winterer, 2005).

Volcanism along the Louisville chain also shows a marked decrease in output around 25 Ma (Géli et al., 1998). In plume models, the present day Louisville hotspot is considered to lie beneath a seamount at 50^0 S 139^0 W (Lonsdale, 1988b), such that the chain appears to curve away from the Eltanin fracture zone. However, a single seamount should not be considered evidence for a plume, and although the seafloor topography map of Smith and Sandwell (1997) shows few seamounts in the region, there are others at comparable distances from the Pacific-Antarctica ridge which have no relationship to any island chain in the region, such as the seamount at 60^0 35’ S 129^0 W south of the Udintsev fracture zone. Volcanism along the Louisville chain is thus suggested to have been terminated by the Late Oligocene plate reorganisations, with the sporadic formation of small seamounts since then being related to axial ridge processes or reactivation of other fracture zones in the region.

The fission of the Farallon plate into Cocos and Nazca plates was accompanied by re-orientation of the East Pacific Rise orthogonal to the orientation of subduction. This re-organisation of spreading centres provides strong evidence for the transmission of stress from the convergent margin across the plate (Natland and Winterer, 2005). Relationships between convergent margin geometry and intraplate features have also been suggested for the Nazca plate (Isacks and Barazangi, 1977; Anderson, 1998; Favela and Anderson, 2000; Smith, 2003). Between 10^0 and 40^0 S, the convergent margin is characterised by two regions of flat (15^0 dip) subduction, marked by an absence of Quaternary volcanism in southern Peru and central Chile (Isacks and Barazangi, 1977) (Fig. 5). The Juan Fernandez ridge was noted to coincide with a change in slab dip at 33^0 S, and the Nazca Ridge and Sala y Gomez chains to extrapolate to changes in slab geometry at 15^0 and 28^0 S. In conventional models, the assumption of a plume origin for the volcanism has resulted in the changes in slab geometry being attributed to subduction of topographic features associated with the intraplate volcanism (e.g. Yáñez et al., 2002). Such a relationship may apply with the Nazca ridge, as the volcanism was terminated by the Late Oligocene plate re-organisation. However, the slab flexure at 33^0S cannot result from subduction of the Juan Fernandez ridge as the older section of the ridge curves northeast and intersects the convergent margin at 31^0 S (Fig. 5). Similarly, if the Sala y Gomez chain was initiated as a result of the Late Oligocene plate re-organisation, the age progression along the chain, suggests it has yet to undergo subduction, such that the chain cannot be causing the change in slab geometry. The region between 25^0 and 28^0 S to which the Sala y Gomez chain extrapolates, is characterised by a change in curvature of the slab from concave upward to convex upward (Cahill and Isacks, 1992). The zone of slab deformation would thus be more extensive than the abrupt flexuring of the slab at 15^0 and 33^0 S. Transmission of stress from warping of the slab may thus explanation the diffuse distribution of volcanism along the Sala y Gomez chain compared to the more linear Juan Fernandez chain.

CONCLUSIONS

Intraplate volcanism shows a relationship to plate tectonic events and lithospheric architecture throughout the history of the Pacific ocean basin, which suggests that plate tectonics controls the volcanism, rather than the volcanism controlling plate interactions as proposed for some stages of basin evolution in plume models. In the Middle Jurassic to Early Cretaceous, the accreted terrane record indicates that extensive island arc-marginal basin systems existed in the northwest, northeast and southwest margins of the basin. Major plate tectonic events such as the opening of the Atlantic caused tectonic reorganisations within the marginal basin systems, such that stresses were not transmitted directly to the Pacific plate. The Pacific plate thus grew symmetrically with no well-defined stress direction until the mid Early Cretaceous, such that apart from generation of oceanic plateaus by focussing of melts at triple junctions, intraplate volcanism on the plate lacked any linear trend or age progression. Basins along the northeast and southwest of the Pacific basin closed in the Early Cretaceous, halting the subduction of the Farallon and Phoenix plates. Stresses were transmitted across these plates, slowing the migration of spreading on the Pacific-Farallon and Pacific-Phoenix ridges, leading to the generation of the Hess Rise, Ontong Java, and Manahiki-Hikurangi oceanic plateaus.

Modelling of plate configurations for Late Cretaceous-Early Eocene is critically dependent on the timing of the onset of subduction of the Pacific plate beneath Asia. Conventional interpretations that the latter event occurred around 83 Ma should be re-evaluated as the fate of the Pacific-Phoenix ridge was not considered. If the Pacific-Phoenix ridge subducted beneath the western Pacific margin in the Late Cretaceous, the North New Guinea plate has to be invoked to explain the Cenozoic tectonic evolution of the western Pacific. Alternatively, following the interpretation of Osozawa (1992, 1997) that two ocean ridges migrated along the Japanese margin in the Late Cretaceous-Paleogene, an alternative hypothesis can be constructed whereby a fragment of the Phoenix plate was trapped in the western Pacific, where it subsequently became the North New Guinea plate. In this second, and preferred model, subduction along the western margin of the Pacific plate did not commence until the Early Eocene. Other complexities arise from the uncertainties in the South Pacific plate configuration, and the possible relationship of volcanism along the Louisville chain to ocean ridge systems west of the Osbourn Trough. The Late Cretaceous-Early Eocene nonetheless appears to have been characterised by three contrasting styles of ocean island volcanism: diffuse volcanism in the Marcus-Wake, Magellan and Marshall islands suggests the plate continued to lack any well-defined stress field until at least the Early Campanian plate reorganisation, linear non-age progressive volcanism in the Line Islands followed pre-existing lines of weakness in the plate fabric, and the Emperor chain may have been related to tearing of the Pacific plate as a result of subduction geometry under island arc systems in the northwest of the basin.

The Pacific plate attained its modern configuration in the Early Eocene as a result of the onset of subduction under Asia, accretion of arc systems along the Kamchatkan margin and the abandonment of the Kula-Pacific ocean ridge, and initiation of the Izu-Bonin-Mariana arc and the demise of the Pacific-North New Guinea ocean ridge. The latter two events changed the stress field in the western Pacific from compressional to extensional, giving rise to the Hawaiian chain as a propagating fracture. Subsequent ocean island volcanism was linear as a result of the development of a principal stress direction from subduction of the Pacific plate, but shows a strong relationship with Late Oligocene and Miocene-Pliocene plate re-organisations. The key to understanding the volcanic record is thus in reconstruction of tectonic history of the basin rather than modelling of hypothetical plume events.

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Figure Captions

1. Map of the Pacific basin showing the location of ocean island volcanism and oceanic plateaus on the present day seafloor and in accreted terranes around the basin margin. Abbreviations; plates: AN Antarctica, AU Australia-India, CB Caribbean, CO Cocos, EU Eurasia, NA North America, NZ Nazca, PA Pacific, PH Philippine SA South America; submerged plateaus: CH Chatham Rise, CP Campbell plateau, LHR Lord Howe Rise, NR Norfolk Ridge; basins: NCB New Caledonia basin; ocean ridges: EPR East Pacific Rise, JdF Juan de Fuca, OT Osbourn Trough; fracture zones: Efz Eltanin fracture zone. Seamounts, island chains, plateaus and rises with ages in Ma from Clouard and Bonneville (2005) except where indicated: AC Austral-Cook (29-0), BR Benham Rise (~49), CL Caroline (14-1), CE Cobb-Eickelberg (26-2), CN Carnegie Ridge (15-0), CS Cocos Ridge (20-0), EM Emperor (81-46), FO Foundation (21-2), HA Hawaiian (46-0), HP Hikurangi plateau (>85; Mortimer and Parkinson, 1996), HR Hess Rise (111-87), JF Juan Fernandez (0-4; Baker et al., 1987), JS Japanese seamounts (108-71), KB Kodiak-Bowie (25-0), LI Line islands (128-35), LV Louisville (77-25), MA Magellan seamounts (129-74), MG Marshall Gilbert (138-56), MJ Meiji (>81), ML

Magellan Rise (135-100), MN Manihiki Plateau (123-110), MP Mid-Pacific Mountains (>128-73), MQ Marquesas (6-1), MU Musicians (98-64), MW Marcus Wake (120-78), NR Nazca Ridge (>25; O'Connor et al., 1995), OJ Ontong Java (122-105; Fitton et al., 2004), PG Pitcairn-Gambier (11-0), SA Samoa (14-5), SG Sala y Gomez (22-0; O'Connor et al., 1995), SO Society (5-0), SR Shatsky Rise (143-129), TU Tuamotu (>50-20). Accretionary terranes/assemblages containing ocean island/plateau material with ages of active intraplate volcanism (compiled from Monger and Berg, 1987; Silberling et al., 1987; Taira et al., 1989; Sun et al., 1998; Reynaud et al., 1999; Nokleberg et al., 2000): AI Amaime (160-130), AA Anvina (185-110), AR Arperos (150-110), CC Cache Creek (340-200), CR Crescent (62-49), GO Gorgona-Serranía de Baudó (88-73), KY Kronotskiy (83-25?) NB Nabilsky (185-110), NC Nicasio (140-135), PI Piñón (>95-80), PL, SJ Pallatanga-San Juan (~123), SH Shimanto (132-65), SY Sorachi-Yezo (163-147), TA Taukha (170-150), TO Tokoro (142-110), WR Wrangell (232-228), YU Yuli (>110-97).

2. Plate reconstruction of the Pacific basin from the Middle Jurassic to the mid-Eocene. Active intraplate volcanism is shown in black. Ocean ridges shown by solid lines are defined by magnetic anomalies; ocean ridges indicated by dashed lines are speculative. Reconstructions are based on Scotese et al. (1988), with modifications as indicated for each timeframe. Plates: AN Antarctic, CB Caribbean, FA Farallon (n = north, s = south), HK Hikurangi, IV Iruney-Vatuna, IZ Izanagi, KU Kula, MB Marie Byrd, MO Moa, NG North New Guinea, PA Pacific, PH Philippine, PX Phoenix. Ocean ridges: OT Osbourn Trough, PA-MN Pacific-Manahiki, TR Tasman ridge, VO Vetlovka. Ocean island/seamounts (triangles): AA Anvina, CR Crescent, EM Emperor, HA Hawaiian, JS Japanese seamounts, LI Line Islands, LV Louisville, KY Kronotskiy, MA Magellan, MG Marshall-Gilbert, MU Musicians, MW Marcus-Wake, NB Nabilsky, NC Nicasio, SH Shimanto, TA Taukha, TM Tamba, TO Tokoro, TU Tuamotu, YU Yuli. Oceanic plateaus (shaded = active, unshaded = inactive), CH Chichibu, CP Caribbean, GO Gorgona-Serranía de Baudó, HP Hikurangi, HR Hess Rise, OJ Ontong Java, MJ Meiji, ML Magellan, MN Manahiki, MP Mid Pacific Mountains, PA

Pallatanga, PI Piñón, SJ San Juan, SR Shatsky Rise, SY Sorachi. Crustal blocks: AK Abukuma-South Kitakami, CH Chatham Rise, IT Insular terrane, JA Japan/Sikhote- Alin, KG Kurosegawa, LHR Lord Howe Rise, OK Okhotia (ocean floor and arc composite; dashed line indicates the extent of crust considered part of the Okhotsk block), NR Norfolk Ridge, RB Reed Rank, TH Tahin. Arcs: AC Academy of Sciences, AV Achivayam-Valaginskaya, CA Central America, CGA Caribbean Great Arc, EP East Philippines-Daito Ridge, GU Guerrero, IBM Izu-Bonin-Mariana, KR Kronotskaya, KV Kvakhon, MC Macuchi, MT Median Tectonic, ON Oku-Niikappu, VZ Vizcaino.

(a) Middle Jurassic (~170 Ma). The Pacific plate formed in the centre of the paleo-Pacific basin. Island arcs fringed much of the basin (after Maruyama et al., 1989; Muir et al., 1995; Nokleberg et al., 2000). Intraplate volcanism was limited to ocean island/seamount volcanism in marginal basins in the northwest of the Pacific basin. The sequence of terrane accretion that led to the formation of continental crust of Japan and Sikhote-Alin, follows the reconstruction of Osozawa (1998) for the continental margin of Asia before strike-slip displacement of terranes in the mid-Cretaceous.

(b) Late Jurassic-Early Cretaceous (150-130 Ma). The Pacific plate attained a rhombic outline. The Shatsky Rise and western Mid-Pacific Mountains formed in conjunction with triple junctions along the Pacific-Izanagi and Pacific-Farallon ocean ridges. Island arc systems in the northwest of the basin underwent rifting, leading to the generation of the Sorachi oceanic plateau. Development of island arc systems to the west of Central America, lead to formation of the Arperos basin (Freydier et al., 1996).

(c) Mid-Cretaceous (120-100 Ma). Oceanic plateau formation along the northeast and south/southwest margins of the Pacific plate coincided with plate re-organisations along the adjacent basin margins as a result of accretion of the Insular terrane to North America and Median Tectonic arc to New Zealand. The position of the Manahiki plateau is based on Joseph et al. (1993). The latitude for the Ontong Java plateau is based on Riisager et al. (2002).

(d) Late Cretaceous (84-82 Ma). Major plate re-organisations took place in the north and south of the Pacific basin, with fragmentation of the Izanagi plate and the extension of the Pacific plate southwards on abandonment of spreading on the Osbourn Trough. The plate configuration along the Antarctic margin is from Larter et al. (2002). Remnants of the Phoenix plate were trapped in the west and south of the Pacific basin, where they became the North New Guinea plate and part of the Antarctic plate, respectively. The Phoenix/North New Guinea-Pacific ridge is suggested to have continued into the New Caledonia basin, with the Pacific plate boundary extending across the Norfolk ridge to the Osbourn Trough in the early Late Cretaceous. The boundary between the Australian and North New Guinea plates follows Honza and Fujioka (2004). The plate configuration in the northwest of the basin has been drawn to fit the ocean ridge migration sequence along the Japanese margin described by Osozawa (1997).

(e) Paleocene (~59 Ma). The North New Guinea-Pacific ridge approached the continental margin of Asia, and possibly linked with the Tasman ridge between Australia and the Lord Howe Rise. The Pacific-Kula ridge began to subduct under island arc systems in the northwest of the basin (configuration after Konstantinovskaya, 2000, 2001), leading to generation of the Emperor chain as a propagating fracture.

(f) Mid-Eocene (~46 Ma) illustrating the plate configuration immediately after the Early Eocene reorganisation. The Hawaiian chain evolves as a propagating fracture, whereas volcanism along the Louisville chain is controlled by development of the Eltanin fracture zone (Efz). The outline of the Philippine and North New Guinea plates are based on Hall (2002).

3. Morphological relationship of the Hawaiian-Emperor chain (bathymetric outline according to the 4 km isobath) to events around the Pacific basin margin. Age dates (in Ma) are from Keller et al. (1995), Clouard and Bonneville (2005), and Sharp and Clague (2006). Abbreviations: KY Kronotskiy terrane.

4. Map showing the relationship of volcanism along the Louisville ridge to the Eltanin fracture zone (After Smith and Sandwell, 1997; Vlastelic et al., 1998). Magnetic lineations are from Müller et al. (1997), age dates (in Ma) are from and Clouard and Bonneville (2005). Plates: PA Pacific, AN Antarctica.

5. Relationship of intraplate volcanic features on the Nazca plate to slab geometry between 10^0 and 40^0 S along the South American margin. Quaternary continental arc volcanism is marked by black triangles. The position of the subducted slab at 150 km depth follows Isacks and Barazangi (1977), modified following Cahill and Isacks (1992). Dashed line indicates the trace of the Juan Fernandez ridge as depicted by Yáñez et al. (2002). Ages are shown for the Sala y Gomez and Nazca ridges in Ma (O'Connor et al., 1995). Plates: AN Antarctica, NZ Nazca , PA Pacific, SA South America. Ocean ridges: CR Chile Rise, EPR East Pacific Rise.

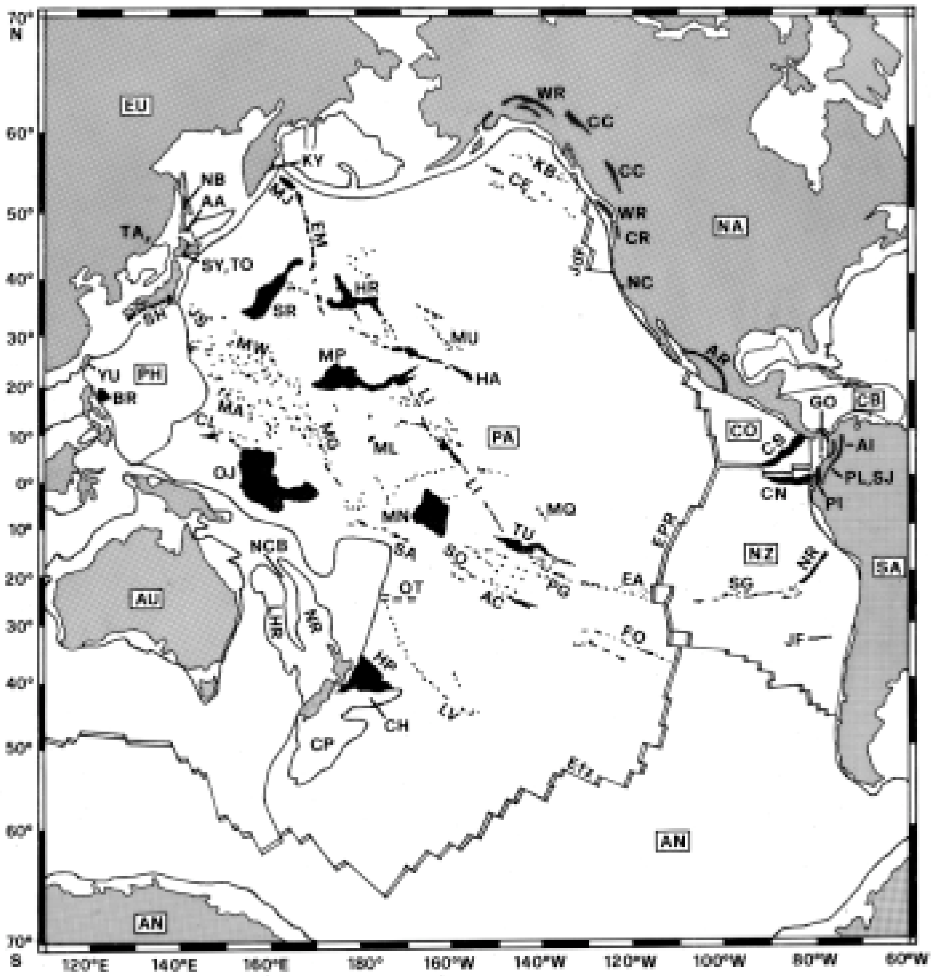

EU
NB
AA
TA
SY, TO
SH
KY
MJ
EM
SR
HR
MW
MP
MU
HA
PH
YU
BR
MA
CL
MG
ML
PA
OJ
LI
MN
MQ
TU
SA
SO
PG
AC
OT
NCB
LHR
NR
HP
CH
CP
LV
AU
AN
WR
CC
KB
CE
CR
NA
NC
AR
CO
GO
CB
AI
PL, SJ
CN
PI
CS
EPR
NZ
EA
SG
FO
JF
70° N
60°
50°
40°
30°
20°
10°
0°
10°
20°
30°
40°
50°
60°
70° S
120°E
140°E
160°E
180°
160°W
140°W
120°W
100°W
80°W
60°W

Fig. 1

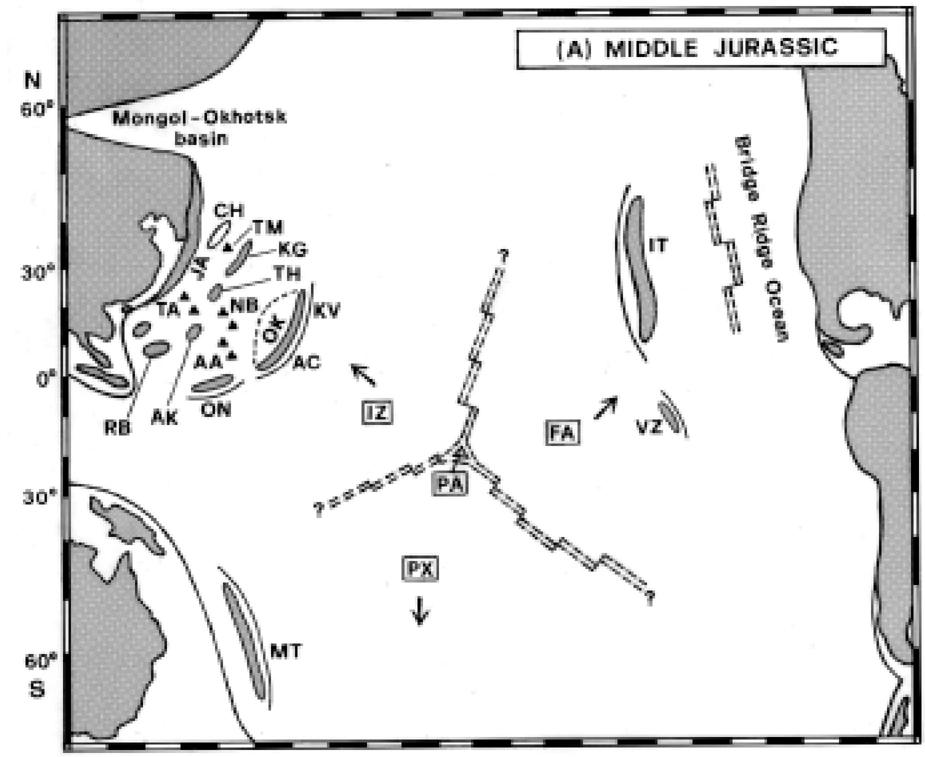

(A) MIDDLE JURASSIC
N
60°
30°
0°
30°
60°
S
Mongol-Okhotsk basin
CH
TM
KG
TH
JA
TA
NB
OK
KV
AA
AC
RB
AK
ON
IZ
PA
PX
FA
VZ
IT
MT
Bridge Ridge Ocean

Fig. 2a

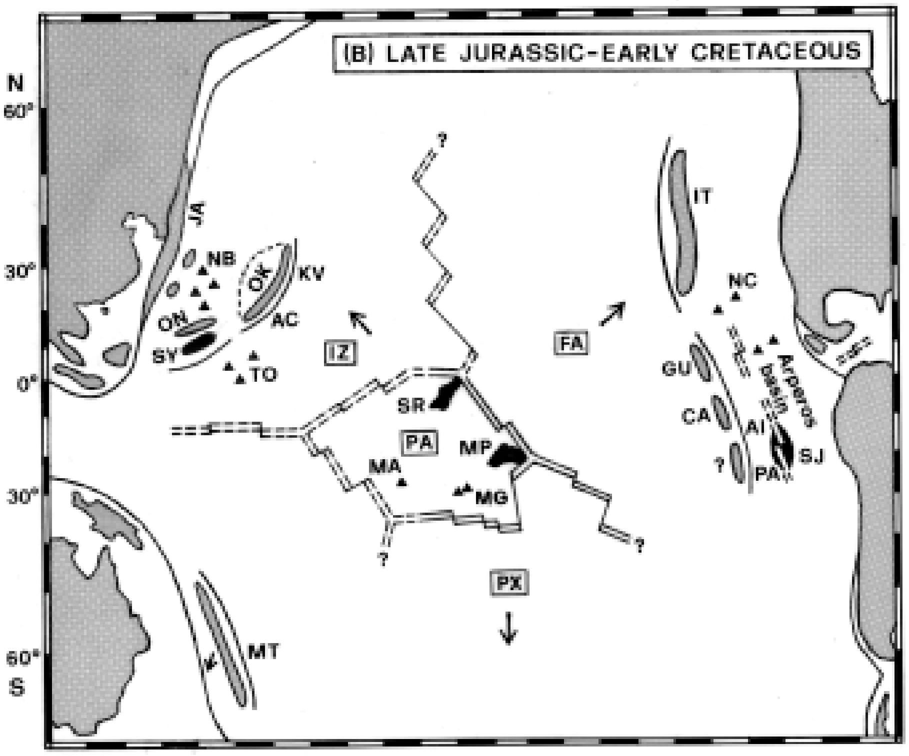

(B) LATE JURASSIC-EARLY CRETACEOUS
N
60°
30°
0°
30°
60°
S
JA
NB
OK
KV
ON
AC
SY
TO
IZ
SR
PA
MP
MA
MG
PX
MT
FA
IT
NC
GU
CA
AI
PA
SJ
Arperos basin
?

Fig. 2b

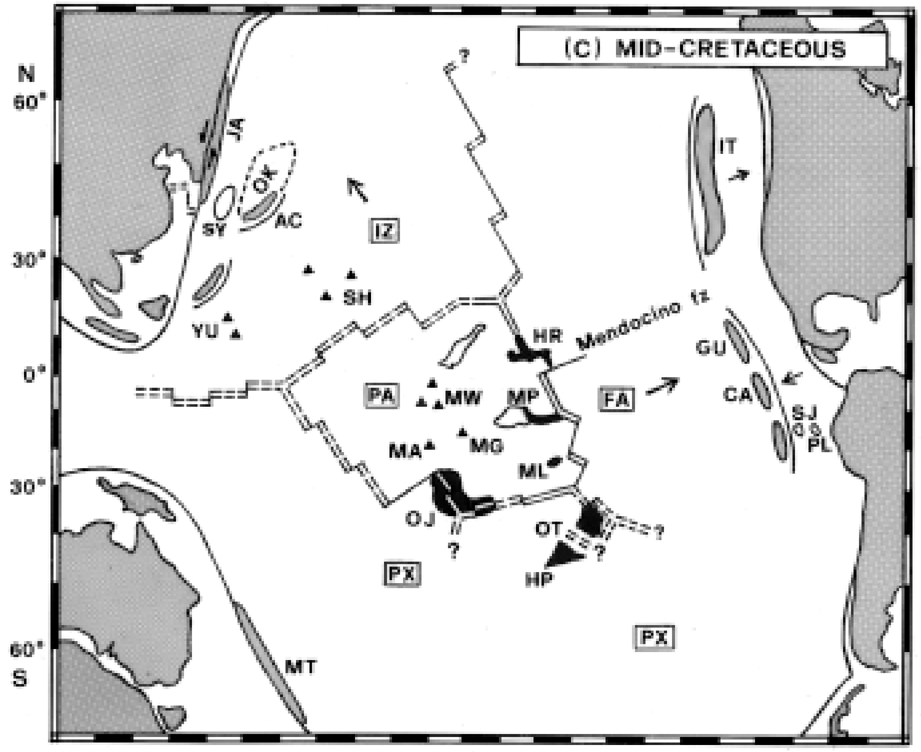

(C) MID-CRETACEOUS
N
60°
30°
0°
30°
60°
S
JA
OK
SY
AC
IZ
SH
YU
PA
MW
MP
HR
Mendocino fz
MA
MG
ML
OJ
OT
HP
PX
PX
FA
IT
GU
CA
SJ
PL
MT

Fig. 2c

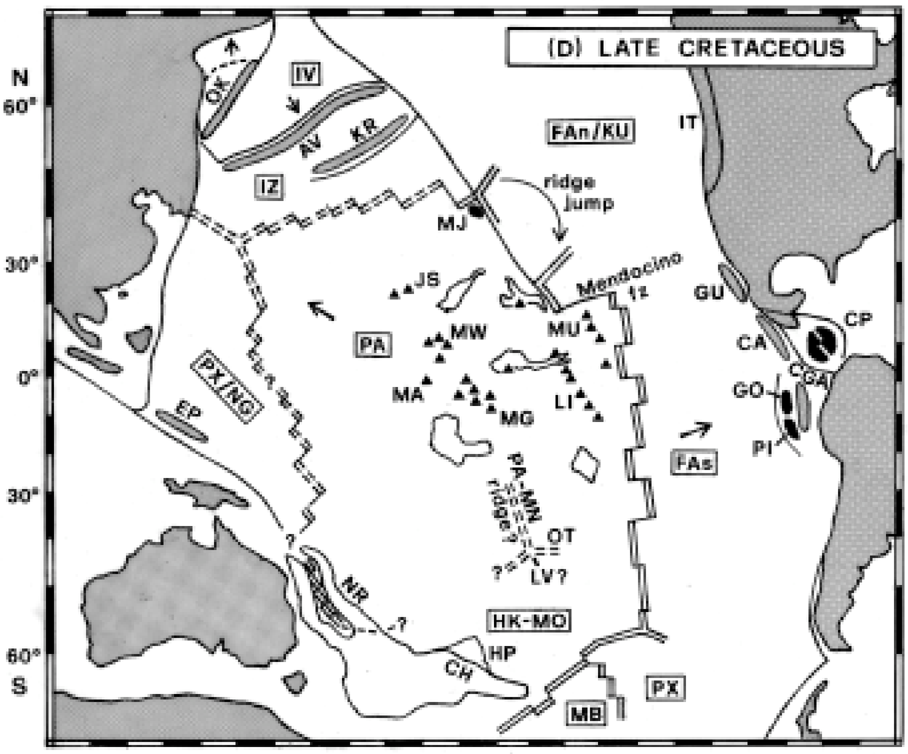

(D) LATE CRETACEOUS
N
60°
30°
0°
30°
60°
S
OK
IV
AV
KR
IZ
FAn/KU
ridge
jump
MJ
IT
Mendocino
fz
JS
CA
MU
GU
PA
MW
CA
CP
PX/NG
MA
MG
LI
GO
CGA
EP
PI
FAs
PA-MN
ridge?
OT
LV?
NR
HK-MO
HP
CH
MB
PX

Fig. 2d

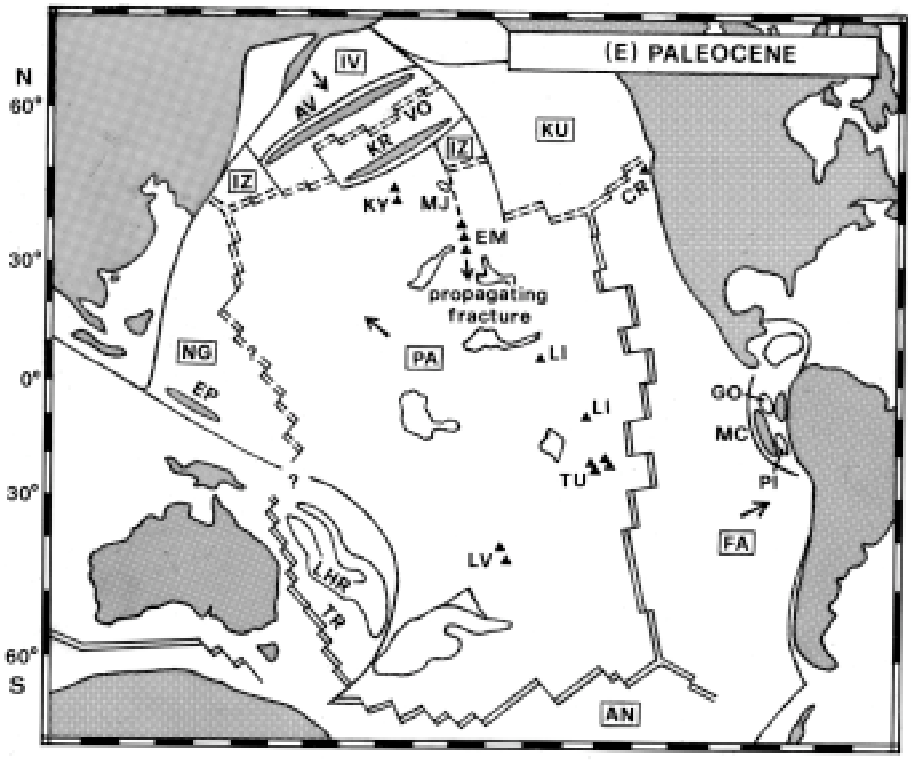
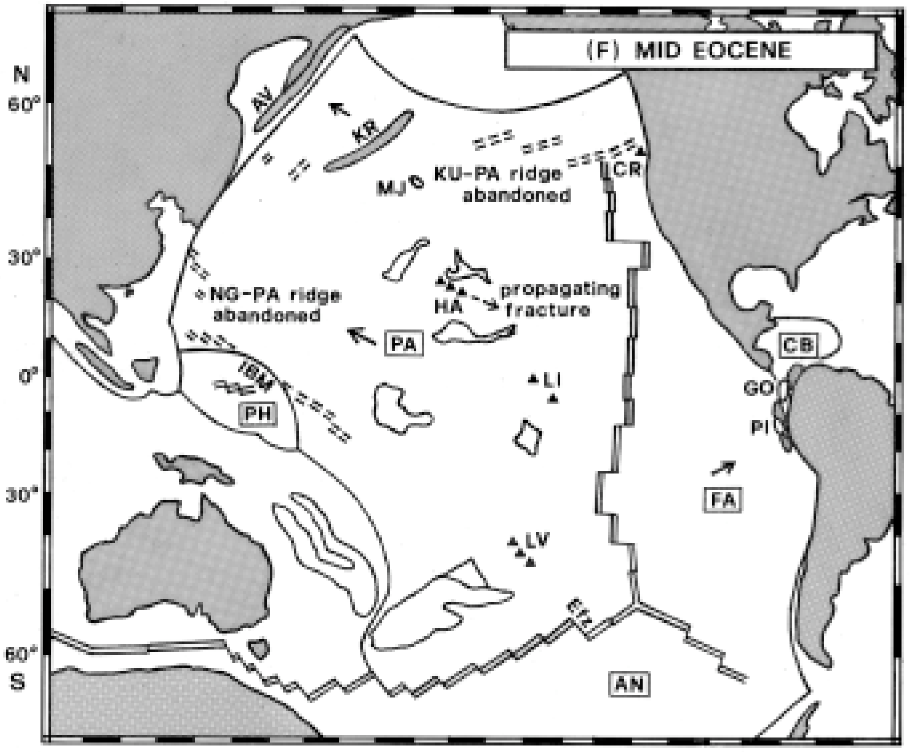

(E) PALEOCENE
N
60°
30°
0°
30°
60°
S
IV
AV
VO
KR
IZ
KU
IZ
KY
MJ
EM
CR
propagating
fracture
NG
PA
LI
EP
LI
GO
MC
TU
?
PI
FA
LV
LHR
TR
AN

Fig. 2e



(F) MID EOCENE
N
60°
30°
0°
30°
60°
S
AV
KR
MJ
KU-PA ridge
abandoned
CR
NG-PA ridge
abandoned
HA
propagating
fracture
PA
IBM
PH
LI
LV
CB
GO
PI
FA
AN

Fig. 2f

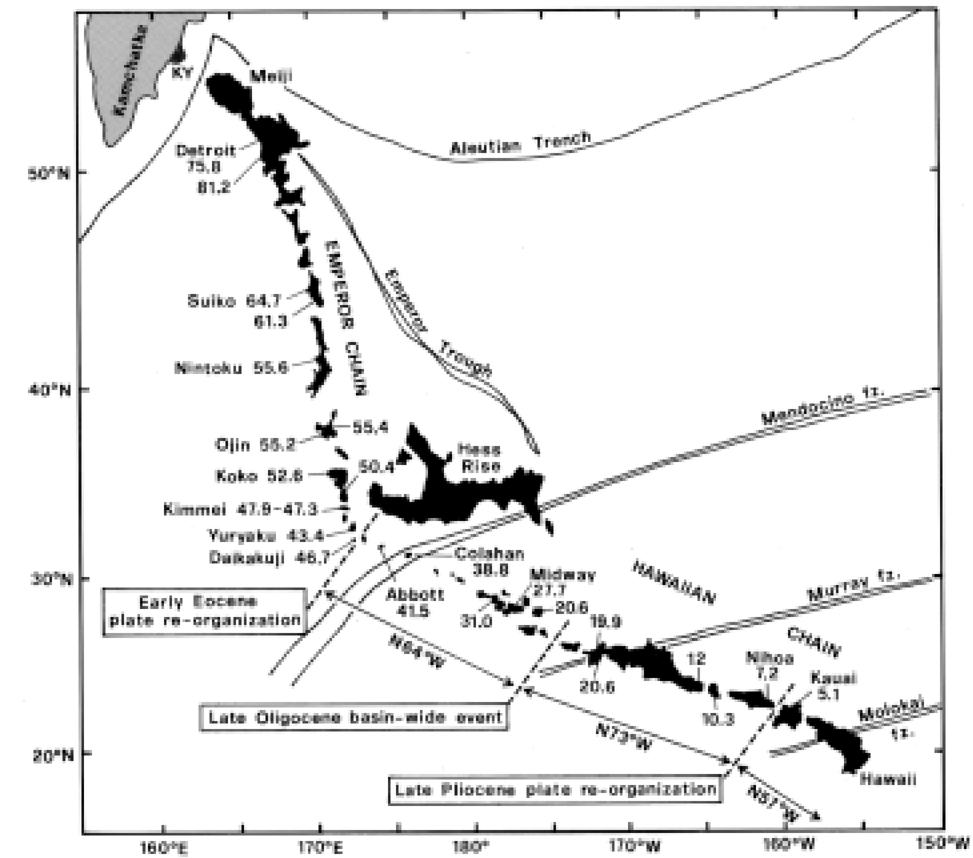
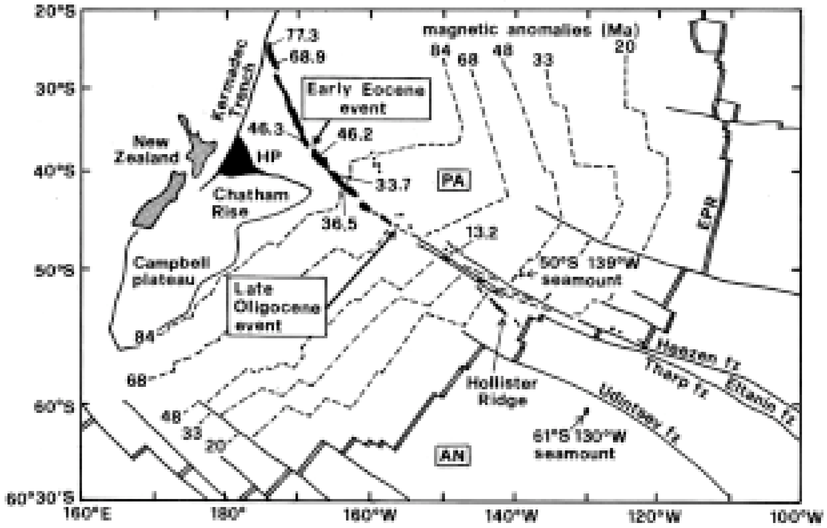
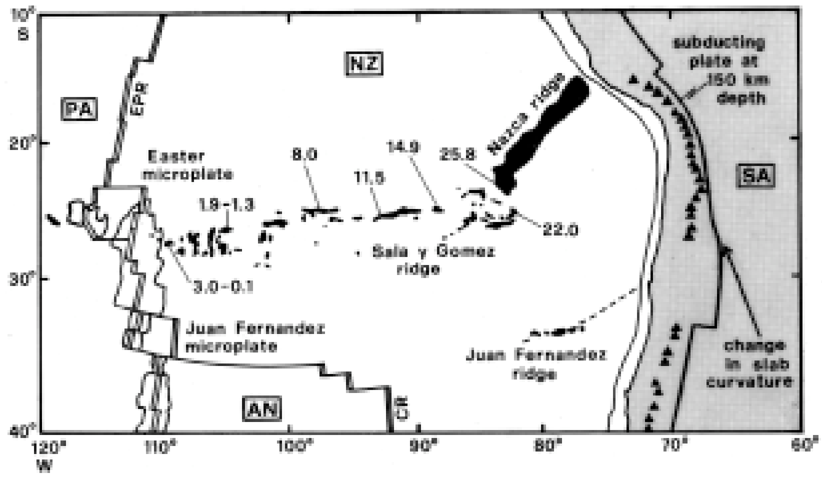

Kamchatka
KY
Meiji
Aleutian Trench
Detroit
75.8
81.2
EMPEROR CHAIN
Emperor Trough
Suiko 64.7
61.3
Nintoku 55.6
55.4
Ojin 55.2
Koko 52.6
50.4
Hess Rise
Kimmei 47.9-47.3
Yuryaku 43.4
Daikakuji 46.7
Colahan
38.8
Abbott
41.5
Midway
27.7
20.6
31.0
Mendocino fz.
HAWAIIAN
CHAIN
Murray fz.
19.9
20.6
12
10.3
Nihoa
7.2
Kauai
5.1
Molokai fz.
Hawaii
Early Eocene plate re-organization
Late Oligocene basin-wide event
Late Pliocene plate re-organization
N64°W
N73°W
N57°W
50°N
40°N
30°N
20°N
160°E
170°E
180°
170°W
160°W
150°W

Fig. 3



magnetic anomalies (Ma)
84
68
48
33
20
77.3
68.9
Kermadec Trench
Early Eocene event
46.3
46.2
New Zealand
HP
33.7
PA
Chatham Rise
36.5
13.2
Campbell plateau
50°S 139°W seamount
Late Oligocene event
EPR
84
68
Hazen fz
Tharp fz
Eltanin fz
Hollister Ridge
48
33
20
Udintsev fz
61°S 130°W seamount
AN
20°S
30°S
40°S
50°S
60°S
60°30'S
160°E
180°
160°W
140°W
120°W
100°W

Fig. 4



10°
S
20°
30°
40°
120°
W
110°
100°
90°
80°
70°
60°
PA
NZ
SA
AN
EPR
CR
Easter
microplate
Nazca ridge
8.0
14.9
25.8
11.5
1.8-1.3
22.0
Sala y Gomez
ridge
3.0-0.1
Juan Fernandez
microplate
Juan Fernandez
ridge
subducting
plate at
150 km
depth
change
in slab
curvature

Fig. 5