SO 224 Manihiki II – Leg 1

DAS MANIHIKI PLATEAU – ENTSTEHUNG, AUFBAU UND AUSWIRKUNGEN OZEANISCHER PLATEAUS UND PLEISTOZÄNE DYNAMIK DES WESTPAZIFISCHES WARMWASSERPOOLS

Abschlussbericht

03G0224

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G. Uenzelmann-Neben, K. Gohl (AWI)

Inhaltsverzeichnis

I.1 Aufgabenstellung

Im Forschungsprojekt Manihiki II – Leg 1 wurden umfangreiche geophysikalische Untersuchungen im Bereich des Manihiki Plateaus durchgeführt, um Struktur und Aufbau der Kruste und des obersten Mantels des Plateaus detailliert zu erfassen. Ziel der Analyse dieser Daten waren Rückschlüsse auf die petrologische Zusammensetzung (ultra-mafisch, mafisch, serpentenisiert etc.) der Kruste und des obersten Mantels. Diese Parameter bilden gerade mit den vulkanisch-geochemischen Untersuchungen (So 192 und So 225) eine zwingende Grundlage, um den petrologisch-magmatischen Aufbau des Plateaus, die Tiefen- und Horizontalverteilung und damit eine Volumenabschätzung und Altersabfolge der Intrusiva und Extrusiva zu ermöglichen. Weiterhin wurden Struktur und Verteilung der Sedimentbedeckung untersucht, um die Entwicklung des Plateaus nach seiner initialen Entstehung und seine Rolle für den Wassermassenfluß im zentralen Pazifik zu verstehen. Die durchgeführten Arbeiten sind Teil des Verbundvorhabens "Manihiki II" und somit thematisch eng verbunden mit dem Projekt Manihiki II – Leg 2 (03G0225A).

Die geophysikalischen Hauptziele waren

1) eine Charakterisierung der Kruste und des oberen Mantel des Manihiki Plateaus, wobei wir uns auf die beiden größten Subplateaus (Western Plateaus und High Plateau) konzentrierten. Es sollten mögliche Unterschiede in Mächtigkeit und P-Wellengeschwindigkeitsverteilung zwischen dem High Plateau und den Western Plateaus identifiziert werden sowie die Einbettung der Krustenstruktur des Plateaus in die umliegenden Becken.
2) Überprüfung der Struktur des Danger Island Troughs als ein Riftsystem, das vor der Bildung ozeanischer Kruste abgestorben ist.
3) die tektonisch-magmatische Charakterisierung des Süd- und des Ostrands (Manihiki Scarp) des Manihiki Plateaus, um den Separationsprozess des Manihiki Plateaus vom Hikurangi Plateau zu dokumentieren. Weiterhin sollte die Bildung des Manihiki Scarps über die Kombination der seismischen und vulkanologisch-geochemischen Untersuchungen rekonstruiert werden.
4) eine Identifikation vollständig ungestörter Sedimentsequenzen, zur Festlegung möglicher Bohrlokationen. IODP proposal 630 soll überarbeitet, aktualisiert und erneut vorgelegt werden. Die dafür auszuwählenden Bohrlokationen müssen so ausgewählt werden, dass kretazische Sedimente und das Basement erbohrt und beprobt werden können. Es ist sicherzustellen, dass diese möglichen Bohrlokationen nicht durch zu dünne, auskeilende kretazische Sequenzen und Erosivstrukturen beeinträchtigt werden.

Durch Kombination der Ergebnisse dieser geophysikalischen Arbeiten mit denen von SO193 MANIHIKI und den Ergebnissen des Projektes Manihiki II – Leg 2 (03G0225A) sollen grundlegende neue Erkenntnisse über die Bildung und Entwicklung ozeanischer LIPs und damit assoziierter Mantelprozesse erarbeitet werden.

I.2 Voraussetzungen

Die tektonische Entwicklung des Manihiki Plateaus ist Gegenstand kontroverser Diskussionen. Es ist unklar, ob diese Large Igneous Province (LIPs) gemeinsam mit dem Ontong Java Plateau und dem Hikurangi Plateau in einem gemeinsamen „Superplume"-Ereignis entstanden ist (Greater Ontong Java Plateau Event; Coffin and Eldholm, 1993; Ingle and Coffin, 2004; Taylor, 2006b). Andere Untersuchungen erbrachten Hinweise darauf, dass das Hikurangi Plateau gemeinsam mit dem Manihiki Plateau entstand (z.B. Billen and Stock, 2000b; Downey et al., 2007; Hoernle et al., 2004). Bis vor Kurzem resultierten die Kenntnisse über das Manihiki Plateau im Wesentlichen aus einigen Ausfahrten (darunter SO35 und SO67) und einem DSDP-Leg (Site 317) in den 1960er und 70er Jahren sowie einer japanischen Ausfahrt 2003 mit R/V HAKUHO MARU (z.B. Ingle et al., 2007)). Im Mai/Juni 2007 wurden von der Arbeitsgruppe Hoernle während SO193 umfangreiche Kartierungen und erstmalig eine Beprobung aller großen geomorphologischen Einheiten des Manihiki Plateaus durchgeführt. Ergebnisse deuten auf eine deutlich komplexere geodynamische Entwicklung des Manihiki Plateaus hin als bisher angenommen (Coffin et al., 2007; Hauff et al., 2008; Hoernle et al., 2008; Werner and Hauff, 2007; Werner et al., 2007). Seismische Daten lagen bisher nur von den älteren Ausfahrten vor und wurden in Winterer et al. (1974a) zusammenfassend dargestellt. Dabei handelt es sich um seismische Einkanaldaten aus den 60er und 70er Jahren, die als Analogschriebe vorliegen. Beiersdorf et al. (1995b) stellen weiterhin 3.5 kHz und 20 kHz Daten, ebenfalls Analogschriebe, vor. Mit den von Winterer et al. (1974a) und Beiersdorf et al. (1995b) vorgestellten Daten ist es aber leider nicht gelungen, die Sedimentpakete vollständig zu durchdringen, da hierfür stärkere Quellen wie GI-guns erforderlich sind. Über das Projekt Manihiki II wurden nun zwei refraktionsseismische und 28 reflexionsseismische Profile im Gebiet des Manihiki Plateaus erfasst, welche eine Untersuchung der Krustenstruktur und damit Hinweise auf den Entstehungsprozess sowie Pfade des Strömungssystems in diesem Gebiet ermöglichen.

I.3 Planung und Ablauf des Vorhabens

Der Ablauf des Vorhabens entsprach im Wesentlichen der von uns im Antrag vorgeschlagenen Arbeits-, Zeit- und Aufgabenplanung.

Die Vermessungsarbeiten auf der Expedition So 224 (11.10. – 17.11.2012) verliefen außerordentlich erfolgreich. Insgesamt wurden ca 4200 km an reflexionsseismischen und ca 1000 km an refraktions/weitwinkelseismischen Daten gewonnen. Für die refraktionsseismischen Profile wurden zweimal 33 Stationen ausgebracht. Bathymetrische und Parasound Daten wurden parallel zu den seismischen Daten gesammelt. Diese letzteren zwei Datensätze wurden weiterhin für die Planungen während der Expedition So 225 (Manihiki II – Leg 2) genutzt. Der Verlauf der Expedition So 224 sowie eine erste Bewertung der Daten sind im Fahrtbericht nachzulesen (Uenzelmann-Neben, 2012b).

Unmittelbar nach Rückkehr von der Expedition wurde mit der Bearbeitung der seismischen Daten begonnen. Die Bearbeitung konnte dem Zeitplan entsprechend durchgeführt werden. Es schloss sich eine Modellierung der refraktionsseismischen Daten und die Interpretation der reflexionsseismischen Daten an. Weiterhin wurden die Daten unter dem Gesichtspunkt der Auswahl dreier möglicher Lokationen für ein IODP Bohrproposal interpretiert. Erste Publikationen sind im Review bzw werden in Kürze eingereicht. Weitere Manuskripte sind in Planung.

I.4 Wissenschaftlich-technischer Stand

Der wissenschaftlich-technische Stand wurde im Antrag zu dem Forschungsvorhaben So 224 Manihiki II – Leg 1 ausführlich beschrieben.

I.5 Zusammenarbeit mit anderen Stellen

Während der Laufzeit des Projektes wurde mit verschiedenen nationalen und internationalen Instituten intensiv zusammengearbeitet. Diese Kooperationen, die im Rahmen anderer Projekte fortgesetzt werden soll, hat zahlreiche Ergebnisse erbracht, die unter Punkt II.1

näher dargestellt sind. Unsere wichtigsten Kooperationspartner sind (Reihenfolge alphabetisch und national-international):

Alfred-Wegener-Institut
Prof. Dr. R. Tiedemann (paläozeanographische Untersuchungen am Manihiki Plateau, So 225)

GEOMAR
Dr. R. Werner, Dr. F. Hauff, R. Golowin, Prof. Dr. K. Hoernle (vulkanologisch-geochronologisch-petrologische Untersuchung des Manihiki Plateaus, So 225; Erarbeitung eines IODP Bohrvorschlags)
PD Dr D. Nürnberg (paläozeanographische Untersuchungen am Manihiki Plateau, So 225)

Institute for Marine and Antarctic Studies, Hobart, Tasmanien, Australien
Prof. Dr. M. Coffin (Tektonik und Struktur des Plateaus)

Universität Mailand, Italien
Prof. Dr. E. Erba (Erarbeitung eines IODP Bohrvorschlags, Ab- und Umlagerungsprozesse)

Oregon State University, USA
Prof. R.A. Duncan (Erarbeitung eines IODP Bohrvorschlags, Evolution des Plateaus)

Caltech, USA
Prof. Dr. J. Stock (zusätzliche seismische Einkanaldaten der KIWI Expedition)

II.1 Darstellung der erzielten Ergebnisse

II.1.1 Publizierte oder in Manuskripten vorliegende Ergebnisse

Ein Teil der wissenschaftlich-technischen Ergebnisse von So 224 Manihiki II – Leg 1 ist bereits in zwei Manuskripte eingeflossen, die zur Publikation eingereicht wurden. Diese Ergebnisse werden hier nicht detaillierter erläutert. Stattdessen sind die folgenden Manuskripte dem Schlussbericht im Anhang beigelegt:

(a) Hochmuth, K., Gohl, K., Uenzelmann-Neben, G., Werner, R. (in revision): The diverse crustal structure and magmatic evolution of the Manihiki Plateau, Central Pacific. Journal of Geophysical Research, Solid Earth.

In dieser Arbeit werden die refraktions/weitwinkelseismischen Daten vorgestellt, die sich über die Western Plateaus und das High Plateau erstrecken. Eine Analyse und Modellierung führt zu neuen Informationen über die Krustenstruktur beider Sub-Plateaus und ihre magmatische Entwicklung. Während das High Plateau über Krustenmächtigkeit und P-Wellengeschwindigkeiten als eine Large Igneous Province identifiziert werden kann, die einen starken sekundären Vulkanismus erfahren hat, zeigt es sich, dass die Western Plateaus keinem sekundären Vulkanismus ausgesetzt waren. Für beide Sub-Plateaus wird eine unterschiedliche Entwicklungsgeschichte abgeleitet.

(b) Hochmuth, K., Gohl, K., Uenzelmann-Neben, G. (submitted): Playing jigsaw with Large Igneous Provinces – a plate tectonic reconstruction of Ontong Java Nui. Solid Earth.

In dieser Arbeit wird genauer untersucht, wie die pazifische Super-LIP in die einzelnen Teile zerfallen sein kann. Welche Rolle hat das Manihiki Plateau als zentraler Teil gespielt, wie interagierte der aufsteigende Plume mit dem Spreizungszentrum, wie wurde das Plateau dabei tektonisch modifiziert, welche Auswirkungen hatte diese tektonische Aktivität auf die Plattenkinematik des pazifischen Ozeans.

(c) Pietsch, R., Uenzelmann-Neben, G. (in revision): The Manihiki Plateau – a multistage volcanic emplacement history. Geochemistry, Geophysics, Geosystems.

In dieser Arbeit werden die reflexionsseismischen Daten, welche im Gebiet des High Plateaus aufgenommen wurden, vorgestellt. Die Daten werden über synthetische Seismogramme an die Ergebnisse des DSDP Leg 33 Site 317 angeschlossen. Die bereits definierte Seismostratigraphie konnte um drei Horizonte erweitert werden, welche a) den Beginn der Ablagerung sedimentärer Sequenzen, b) den Beginn einer späteren magmatischen Phase, und c) das Top der initialen magmatischen Phase interpretiert werden. Über die Interpretation der reflexionsseismischen Daten werden drei magmatische Phasen unterschieden, welche das High Plateau von einem zentralen, großen Schlot zu einer Vielzahl an kleineren Extrusionszentren aufgebaut haben.

(d) Pietsch, R., Uenzelmann-Neben, G. (submitted): Hotspot tracks discovered at the Manihiki Plateau, a Large Igneous Province in the central Pacific, Erath and Planetary Science Letters.

Die reflexionsseismischen Daten zeigen zusätzlich zum kretazischen und früh-tertiären Vulkanismus noch eine spätere magmatische Phase, welche die jungen Sedimentsequenzen deformiert. Als Ursache für diese späte Reaktivierung werden verschiedene Hotspot Spuren diskutiert. Weiterhin zeigte sich, dass der Suvarov Trog erst deutlich nach Bildung der LIP und Trennung des Manihiki vom Hikurangi Plateau gebildet wurde. Auch diese tektonische Reaktivierung wird in der Arbeit untersucht.

II.1.2 Ausbildungs- und Qualifizierungsarbeiten

K. Hochmuth (voraussichtlich 2015), From Crustal Structure to Platekinematics - the Role of Large Igneous Provinces in the Pacific Ocean. Dissertation, Universität Bremen.

Frau Hochmuth wird im Rahmen des Projektes Manihiki II – Leg 1 eine Dissertation (Betreuer: K. Gohl) anfertigen, welche sich mit der Krustenstruktur des Manihiki Plateaus und seiner Rolle im Verbund der diskutierten ‚Super-LIP' Ontong Java Nui auseinandersetzt. Diese Überlegung wird sie in generelle Betrachtungen zur Entwicklung von LIPs erweitern. Als Basis hierzu dienen die refraktions/weitwinkelseismischen Daten, welche während der Expedition So 224 aufgenommen wurden.

R. Pietsch (voraussichtlich 2015), The Manihiki Plateau – A detailed study of the tectonic and volcanic history of a Large Igneous Province and its impact on paleo-currents in the Pacific Ocean. Dissertation, Universität Bremen.

Frau Pietsch wird im Rahmen des Projektes Manihiki II – Leg1 eine Dissertation (Betreuerin: G. Uenzelmann-Neben) anfertigen, welche sich mit der magmatischen Entwicklung des Manihiki Plateaus und der Bedeutung dieses entstehenden Plateaus als Hindernis für die Zirkulation im zentralen Pazifik auseinandersetzt. Als Basis hierfür dienen die reflexionsseismischen Daten, welche während der Expedition So 224 gewonnen wurden, ergänzt durch seismische Einkanaldaten der KIWI Expedition (Clayton, 1998), die Dr. Uenzelmann-Neben über Prof. Dr. J. Stock (Caltech, USA) für das Projekt erhielt, und Daten aus der Bohrung 317 des DSDP Leg 33 (Shipboard Scientific Party, 1976). Die aus dieser Arbeit resultierenden Ergebnisse werden in zwei Manuskripten (Anlage 4 und 5) sowie in Abschnitt II.1.3 dargestellt.

II.1.3 Weitere Ergebnisse

Darüber hinaus wird noch an einigen Teilaspekten der Daten gearbeitet, weitere Publikationen sind in Vorbereitung. Einige zu bearbeitende Fragestellungen des Projektes Manihiki II – Leg 1, die nicht in den beiliegenden Manuskripten publiziert werden, sind im Folgenden kurz zusammengefasst.

Die reflexionsseismischen Daten sollen im Hinblick auf die Absenkungsgeschichte des Manihiki Plateaus analysiert werden, um Hinweise auf den Einfluss des Plateaus auf die Zirkulation im zentralen Pazifik zu ermitteln. Dafür soll die Paläo-Lokation des Plateaus berücksichtigt werden. Es stellt sich die Frage, wie schnell das Plateau thermisch abgesunken ist und welche Konsequenzen dies für Oberflächen- und Tiefenströmungen hatte. Lässt sich eine mögliche Auswirkung des Zerfalls des ursprünglichen Plateaus in Western und High Plateaus in den sedimentären Ablagerungen identifizieren? Welchen Einfluss hatten die zwei späteren vulkanischen Phasen? Wie wirkten sich die Schließung der Tethys, des Indonesian Throughflows und des Isthmus von Panama aus? Diesen Fragen soll in einer detaillierten Analyse und Interpretation der seismischen Daten nachgegangen werden.

II.2 Wichtigste Positionen des zahlenmäßigen Nachweises

Die wichtigsten Positionen waren neben fahrtbezogenen Kosten wie Reisen und Transporten die TVöD Stellen für Frau Katharina Hochmuth und Ricarda Pietsch (E13, 26 Mo 66%) sowie für Frau Sonja Suckro (E13, 6 Mo 66%).

Die wissenschaftlichen Mitarbeiter haben wesentlich zum Erfolg des Vorhabens beigetragen. Frau Suckro war während der Expedition So 224 wesentlich an der Vorbereitung der Ozeanbodenseismometer für das refraktionsseismische Programm beteiligt und hat weiterhin die Vorbereitung und Durchführung der reflexionsseismischen Arbeiten unterstützt. Frau Hochmuth war während der Expedition an der Durchführung der refraktionsseismischen Arbeiten beteiligt, hat anschließend die OBS Daten prozessiert und modelliert und arbeitet an der Interpretation der Daten. Sie hat die Ergebnisse ihrer Arbeiten bereits in zwei Manuskripte einfließen lassen und bereitet ein drittes vor. Frau Pietsch war an der Vorbereitung und Durchführung der reflexionsseismischen Arbeiten während der Expedition So224 beteiligt und nach der Rückkehr von der Reise die Bearbeitung der Daten durchgeführt. Ihre Interpretation der reflexionsseismischen Daten hat ebenfalls bereits Eingang in zwei Manuskripte gefunden, und sie bereitet ein drittes Manuskript vor.

Auf Antrag konnten Restmittel aus den unmittelbaren Vorhabenkosten, die durch sorgfältiges Wirtschaften entstanden sind, in Personalmittel für einen Post-Doc (E13, 5 Mo) umgewandelt werden. Dr. Jens Grützner hat während dieser Zeit daran gearbeitet, die für einen Bohrvorschlag erforderlichen Datensätze zusammenzustellen.

II.3 Notwendigkeit und Angemessenheit der geleisteten Arbeit

Die wichtigsten Arbeitsschritte (neben Nachwuchsförderung, etc) waren 1) die Literaturarbeiten, 2) Bearbeitung beider seismischer Datensätze, 3) Modellierung der refraktions/weitwinkelseismischen Daten, 4) Interpretation beider Datensätze und 5) Präsentation der Ergebnisse auf Tagungen, in Berichten und wissenschaftlichen Publikationen. Diese Arbeiten wurden von den Antragstellern und den im Projekt angestellten Mitarbeitern (Hochmuth, Pietsch, Grützner) und in Zusammenarbeit mit unseren Kooperationspartnern durchgeführt. Die Arbeiten waren für den Erfolg des Projektes Manihiki II – Leg 1 absolut notwendig und somit angemessen.

II.4 Voraussichtlicher Nutzen, Verwertbarkeit

Durch den erfolgreichen Verlauf des Projektes Manihiki II – Leg 1 haben wir u.a. neue Erkenntnisse über die Entwicklung und den Aufbau des Manihiki Plateaus erhalten. Damit wird u. a. zu einem besseren Verständnis der Entwicklung von Large Igneous Provinces beigetragen. Unsere Ergebnisse geben deutliche Hinweise auf die Prozesse, die während der Bildung einer Groß-LIP im zentralen Pazifik bestehend aus Ontong Java, Manihiki und

Hikurangi Plateaus sowie des anschließenden Zerfalls in die drei einzelnen Plateaus aktiv waren. Darüber hinaus sollen neue Erkenntnisse über die Entwicklung des Strömungssystems im zentralen Pazifik während des Tertiärs gewonnen werden. Diese Erkenntnisse werden u.a. in die Modellierung der weltweiten Strömungen einfließen. Die Ergebnisse des Projektes wurden auf zahlreichen Tagungen vorgestellt und werden in einer Reihe von Publikationen präsentiert, die bereits eingereicht oder in Vorbereitung sind.

Weiterhin bilden die während des Projektes Manihiki II – Leg 1 gewonnenen Daten die Basis für einen IODP Bohrvorschlag, der in Planung ist.

Die Daten, die während Manihiki II – Leg 1 gewonnen wurden, wurden der Öffentlichkeit über die Datenbank PANGAEA und das BSH zur Verügung gestellt. Weiterhin sind die bathymetrischen, sedimentechographischen und seismischen Daten dem Ministry of Marine Resources der Cook Islands übergeben worden. Dort fließen die Daten in die Vorbereitung des Claims der Cook Island im Rahmen des ‚Law of the Sea' (UNCLOS) ein.

II.5 Fortschritt bei anderen Stellen

Das von Dr. R. Werner, Dr. Hauff, Prof. Dr. Hoernle, Dr. Nürnberg und Prof. Dr. Tiedemann geführte Projekt SO 225 Manihiki II – Leg 2 erbrachte wichtige Hinweise auf die magmatisch-geochronologisch-petrologische sowie paläozeanographische Entwicklung des Manihiki Plateaus. Dredgeproben, die während So 225 genommen wurden, geben Informationen über den Ursprung des magmatischen Materials und werden weiterhin auf ihr Alter analysiert. Diese Ergebnisse fließen in unsere Untersuchungen ein. Im Hinblick auf die jüngere paläozeanographische Entwicklung des Plateaus arbeiten wir mit Dr. Nürnberg und Prof. Dr. Tiedemann zusammen.

II.6 Erfolgte und geplante Publikationen der Ergebnisse

Artikel (peer-reviewed)

Hochmuth, K., Gohl, K., Uenzelmann-Neben, G., Werner, R. (in revision): The diverse crustal structure and magmatic evolution of the Manihiki Plateau, Central Pacific. Journal of Geophysical Research, Solid Earth.

Hochmuth, K., Gohl, K., Uenzelmann-Neben, G. (submitted): Playing jigsaw with Large Igneous Provinces – a plate tectonic reconstruction of Ontong Java Nui. Solid Earth.

Pietsch, R., Uenzelmann-Neben, G. (in revision): The Manihiki Plateau – a multistage volcanic emplacement history. Geochemistry, Geophysics, Geosystems.

Pietsch, R., Uenzelmann-Neben, G. (submitted): Hotspot tracks discovered at the Manihiki Plateau, a Large Igneous Province in the central Pacific, Earth and Planetary Science Letters.

Vorträge und Poster auf Tagungen

Pietsch, R. and Uenzelmann-Neben, G. (2013): Seismic stratigraphy of the High Plateau, Mahiniki Plateau, as seen in seismic reflection data measured during cruise SO224 , PhD Days, 3 June 2013 - 6 June 2013.

Uenzelmann-Neben, G., Erba, E., Duncan, R., Hoernle, K., Werner, R., Gohl, K., Hauff, F., Nürnberg, D. and Geldmacher, J. (2013): Manihiki Plateau, southwestern Pacific: Archive of multi-phase magmatism, anoxic environment and modifications in ocean circulation and climate , IODP Cretaceous Climate workshop, London, UK, 15 April 2013 - 17 April 2013.

Werner, R., Uenzelmann-Neben, G., Nürnberg, D., Hoernle, K., Gohl, K., Tiedemann, R., Hauff, F., Partnyagin, M., So224 Fahrtteilnehmer, and So225 Fahrtteilnehmer, (2013) Hintergrund und erste Ergebnisse von SO-224 und SO-225 (MANIHIKI II): Das Manihiki-Plateau - Entstehung, Aufbau und Auswirkungen ozeanischer Plateaus und pleistozäne Dynamik des westpazifischen Warmwasserpools, Sonne Statusseminar, Kiel, Germany, 13 February 2013 - 15 February 2013 .

Hochmuth, K., Gohl, K. , Uenzelmann-Neben, G. and Werner, R. (2014): The fragmented Manihiki Plateau – Key region for understanding the break-up of the "Super" Large Igneous Province Ontong Java Nui, AGU Fall Meeting, San Francisco, USA, 15 December 2014 - 19 November 2014 .

Hochmuth, K., Gohl, K. and Uenzelmann-Neben, G. (2014): Insights into the crustal structure and magmatic evolution of the High and Western Plateau of the Manihiki Plateau, Central Pacific , General Assembly of the European Geosciences Union, Wien, 27 April 2014 - 3 May 2014.

Hochmuth, K., Gohl, K. and Uenzelmann-Neben, G. (2014): The tectonic and magmatic evolution of the Manihiki Plateau, Central Pacific, DGG 2014, Karlsruhe, 10 March 2014 - 13 March 2014 .

Hochmuth, K., Gohl, K. and Uenzelmann-Neben, G. (2014): The Crustal Structure of the Manihiki Plateau (Equatorial Pacific) - First results of refraction/wide-angle reflection seismic experiments -, Geophysik Seminar, Universität Bremen.

Pietsch, R. and Uenzelmann-Neben, G. (2014): A multistage volcanic and tectonic evolution history on the High Plateau, part of the Manihiki Plateau, Central Pacific, AGU Fall meeting, San Francisco, USA, 15 December 2014 - 19 December 2014.

Pietsch, R. (2014): Volcanic history of the Manihiki Plateau seen in seismic reflection data, Doktorandenseminar, Bremen, Germany, 14 October 2014 - 15 February 2015 .

Pietsch, R. and Uenzelmann-Neben, G. (2014): The Manihiki Plateau - New insights into the magmatic evolution history, DGG, Karlsruhe, 10 March – 13 March 2014.

Hochmuth, K., Gohl, K. , Uenzelmann-Neben, G. and Werner, R. (2015): Untersuchung der Krustenstruktur des Manihiki Plateaus im Rahmen der Expedition SO-224 , Statusseminar 'Meeresforschung mit FS Sonne', Universität Bremen, Germany, 12 February 2015 - 13 February 2015.

Hochmuth, K., Gohl, K., Uenzelmann-Neben, G. and Werner, R. (2015): The break-up of Ontong Java Nui – the dispersal of a super large igneous province over the central Pacific, Deutsche Geophysikalische Gesellschaft, Hannover, 23 March 2015 - 27 March 2015.

Hochmuth, K., Gohl, K., Uenzelmann-Neben, G. and Werner, R. (2015): Playing jigsaw with large igneous provinces – a plate-tectonic reconstruction of Ontong Java Nui, General Assembly of the European Geosciences Union, Wien, 12 April 2014 - 17 April 2014.

Pietsch, R., Uenzelmann-Neben, G. and Werner, R. (2015): Untersuchung der mehrphasigen vulkanischen und tektonischen Aktivität im Gebiet des Manihiki Plateaus mittels Reflexionsseismik (Expedition SO224) , Statusseminar 'Meeresforschung mit FS Sonne', Universität Bremen, Germany, 12 February 2015 - 13 February 2015 .

Pietsch, R. and Uenzelmann-Neben, G. (2015): Hotspot Vulkanismus als Ursache für späte tektonische und vulkanische Reaktivierung des Manihiki Plateaus im zentralen Pazifik?, Deutsche Geophysikalische Gesellschaft, Hannover, 23 March 2015 - 27 March 2015.

Pietsch, R. and Uenzelmann-Neben, G. (2015): Can hotspot volcanism explain the Tertiary tectonic and volcanic reactivation of the Manihiki Plateau, in the central Pacific?, General Assembly of the European Geosciences Union, Wien, 12 April 2014 - 17 April 2014.

Berichte, Öffentlichkeitsarbeit

Uenzelmann-Neben, G. (2012): The expedition of the research vessel "Sonne" to the Manihiki Plateau in 2012 (So 224), Berichte zur Polar- und Meeresforschung = Reports on polar and marine research, Bremerhaven, Alfred Wegener Institute for Polar and Marine Research, 656, 70 p.

Uenzelmann-Neben, G., Gohl, K., Hochmuth, K., Pietsch, R. (2014): Ausstellung während der Vorstellungsreise des neuen Tiefseeforschungsschiff Sonne, Wilhelmshaven-Kiel, November 17th – 29th.

Danksagung

Wir danken dem BMBF für die gewährte Unterstützung und dem PTJ-Warnemünde für die unbürokratische administrative Abwicklung. Unser besonderer Dank gilt Kapitän Mallon und der Besatzung des FS *Sonne* für ihre hervorragende Kooperation während der Expedition. Ihre harte Arbeit, große Erfahrung, stete Hilfsbereitschaft und die sehr angenehme Arbeitsatmosphäre an Bord haben wesentlich zum Erfolg von So 224 Manihiki II – Leg 1 beigetragen.

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ERFOLGSKONTROLLBERICHT – ANLAGE 1

1. Beitrag der Ergebnisse zu den förderpolitischen Zielen des Förderprogramms

Das Forschungsvorhaben So 224 Manihiki II – Leg 1 hat neue Erkenntnisse über die frühe strukturelle Bildung des Manihiki Plateaus sowie die Verbindung dieses Plateaus mit dem Ontong Java und dem Hikurangi Plateau erbracht. Eine bessere Kenntnis dieser Prozesse ist nicht nur grundlegend für die Rekonstruktion der Auswirkungen von massivem Vulkanismus, sondern vor allem für ein besseres Verständnis des Systems Erde. Laut des Rahmenprogramms ‚Forschung für nachhaltige Entwicklungen' (FONA) des BMBF sind ‚Erkenntnisse aus der Erforschung und Beobachtung der Erd- und Umweltprozesse für das Verständnis der Erde als System fundamental und können schließlich in ein effektives Erdsystem-Management überführt werden'. Hier trägt das Projekt So 224 Manihiki II – Leg 1 direkt zu FONA bei, indem es sich durch die Untersuchung von magmatisch-tektonischen Prozessen und ihrer Auswirkung auf das ozeanische Zirkulationssystem im Erdsystem in Schwerpunkte des Aktionsfeldes ‚Erdsystem und Geotechnologien' einfügt.

Unsere Arbeitsansätze tragen somit hinsichtlich der meereswissenschaftlichen Grundlagenforschung, aber auch hinsichtlich der angewandten Forschung zu den förderpolitischen Zielen des BMBF bzw. der Bundesregierung bei. Die Veröffentlichung der Ergebnisse in der Fachliteratur hat zudem auch einen Werbeeffekt für die deutsche Meeresforschung, aus dem Nachfrage nach Expertise, Verfahren und Instrumenten erwachsen mag.

2. Wissenschaftlich-technische Ergebnisse des Vorhabens

Die wissenschaftlichen Erfolge des Projektes AISTEK-I sind im Abschnitt II.1 des Schlußberichtes und in den dem Schlußbericht beigelegten vier Manuskripten sowie im Fahrtbericht ausführlich beschrieben. Auf zahlreichen Tagungen wurde über das Projekt berichtet und weitere Arbeiten, die auf den vorliegenden Ergebnissen aufbauen, sind in Planung. Einige der wichtigsten Ergebnisse sind:

- Es wurden die ersten Refraktions/weitwinkelreflexionsseismischen Modelle über die zwei Hauptteile (Western Plateaus und High Plateau) des Manihiki Plateaus vorgelegt. Basierend auf diesen Modellen wurde die magmatisch-tektonische Entwicklung dieser ozeanischen Large Igneous Province (LIP) rekonstruiert.
- Die Western Plateaus sind durch eine Krustenmächtigkeit zwischen 9.3 km (stark gedehnte Kruste im Westen) und 17.2 km im Osten charakterisiert. Die obere Kruste besteht nicht aus Extrusiva, wie eigentlich für eine LIP erwartet. Durch einen Vergleich mit anderen ozeanischen LIPs zeigt sich, dass die Western Plateaus keinem sekundären Vulkanismus ausgesetzt waren.
- Das High Plateau zeigt die für ozeanische LIPs normale Krustenstruktur mit einer Mächtigkeit von 20 km. Für die obere Kruste konnten Extrusionszentren identifiziert werden. Das Fördersystem der LIP ist in der mittleren und unteren Kruste verfolgbar.
- Die Danger Island Troughs bilden eine distinkte Grenze zwischen dem vulkanisch aktiven High Plateau und den Western Plateaus, die keine weitere vulkanische Aktivität erfuhren.
- Die Modelle zeigen eine diverse tektonisch-magmatische Entwicklung nach der Bildung als Teil von Ontong Java Nui. Während das High Plateau sekundärem Vulkanismus und nur begrenzten tektonischen Kräften an den Rändern ausgesetzt war, vergleichbar den Ontong Java und Hikurangi Plateaus, erfuhren die Western Plateaus starke Krustendehnung und keinen weiteren Magmatismus. Somit spielte die

Bildung und Entwicklung der verschiedenen Subplateaus des Manihiki Plateaus eine große Rolle im Aufbruch der ‚Super-LIP‘ Ontong Java Nui.

- Das publizierte seismostratigraphische Modell wurde um drei Horizonte erweitert, die den Beginn der sedimentären Ablagerung und den Beginn der dritten vulkanischen Phase (~ 85 Ma) markieren und einen frühen intrabasement Horizont, welcher der Interpretation nach die erste vulkanische Phase beendet.
- Es konnte die frühe Bildung des Manihiki Plateaus über einen zentralen Schlot im High Plateau identifiziert werden. Bereits die zweite vulkanische Phase ist durch eine Vielzahl an Extrusionszentren geprägt. Während dieser späteren Phase verlagerten sich die Extrusionszentren zu den Rändern des High Plateaus.
- Für den östlichen Rand des High Plateaus verdichten sich die Hinweise, dass dort ein Teil des Manihiki Plateaus vor Bildung des Horizont R7 (~ 85 Ma) abgetrennt wurde.
- Der südwestliche Rand des High Plateaus zeigt alle Charakteristika eines geriften Randes. Der Suvarov Trog erscheint noch nach 65 Ma tektonisch aktiv gewesen zu sein.
- Die Deformation junger Sedimentsequenzen deutet auf rezent tektonisch-magmatische Aktivität.

3. Einhaltung der Zeit- und Aufgabenplanung

Der im Antrag vorgestellte Finanzierungs- und Zeitplan wurde weitestgehend eingehalten. Die einzigen Abweichungen waren:

- Die unmittelbaren Vorhabenkosten 0850 sowie Reisekosten zur Expedition 0838 aus dem Haushaltjahr 2012 wurden nicht vollständig ausgeschöpft, da die Antragsteller gut gewirtschaftet haben, weshalb die Restmittel in die Position 0837 ‚Personalmittel’ umgewidmet wurden. Dr. Jens Grützner wurde fünf Monate aus diesen Mitteln finanziert, um die Bereitstellung der für einen IODP Antrag nötigen Datensätze zu unterstützen.
- Es wurde beantragt, die unter der Position 0830 bewilligten Mittel für eine Reise zur AGU Herbsttagung 2014 für die Teilnahme an einem Workshop in Ghent zu Sedimentdrifts verwenden zu können. Dies wurde bewilligt, allerdings doch nicht in Anspruch genommen. Frau Pietsch hat ihre Arbeiten im Dezember 2014 auf der AGU Herbsttagung vertreten.

4. Verwertbarkeit der Ergebnisse

Durch den erfolgreichen Verlauf des Projektes So 224 Manihiki II – Leg 1 liegt ein umfassender Datensatz vor, der deutlich zum Verständnis der Krustenstruktur und tektono-magmatischen Entwicklung des Manihiki Plateaus und seiner Rolle im Verbund der ‚Super-LIP‘ Ontong Java Nui sowie seiner Bedeutung als Hindernis für die Zirkulation im zentralen Pazifik beiträgt. Die Ergebnisse werden zur Zeit in internationalen Fachzeitschriften publiziert und diskutiert. Weiterhin sind die Daten an nationale und internationale Datenbanken transferiert worden und stellen so die Basis z.B. für die Erweiterung der ökonomischen Zone der Cook Islands.

5. Erfindungen, Schutzrechtsanmeldungen und erteilte Schutzrechte

Erfindungen und Schutzrechtsanmeldungen wurden im Rahmen des Projektes Manihiki II – Leg 1 nicht gemacht, es wurden auch keine Schutzrechte erteilt.

6. Arbeiten, die zu keiner Lösung geführt haben

Die im Antrag formulierten Zielsetzungen des Vorhabens konnten weitestgehend erreicht werden. Aus den gewonnenen Daten und Ergebnissen haben sich aber einige neue

Fragestellungen ergeben, deren Bearbeitung Gegenstand zukünftiger Forschungsvorhaben sein soll.

REFERENZEN

Beiersdorf, H., Bach, W., Duncan, R., Erzinger, J., Weiss, W., 1995. New evidence for the production of EM-type ocean island basalts and large volumes of volcaniclastites during the early history of the Manihiki Plateau. Marine Geology 122, 181-205.

Billen, M.I., Stock, J., 2000. Morphology and origin of the Osbourn Trough. J. Geophys. Res. 105, 13481-13489.

Clayton, R.W., 1998. KIWI 12. CalTech, p. 12.

Coffin, M.F., Eldholm, O., 1993. Scratching the surface: Estimating the dimensions of large igneous provinces. Geology 21, 515-518.

Coffin, M.F., Werner, R., Hauff, F., Hoernle, K., Party, S.S., 2007. Extensional and Transtensional Tectonics of the Manihiki Plateau, Western Equatorial Pacific Ocean, AGU Fall Meeting, San Francisco, USA.

Downey, N.J., Stock, J.M., Clayton, R.W., Cande, S.C., 2007. History of the Cretaceous Osbourn spreading center. J. Geophys. Res. 112, B04102.

Hauff, F., Hoernle, K., Werner, R., van den Bogaard, P., Timm, C., Klügel, K., 2008. Geochemistry and age data from the Manihiki Plateau (R/V SO193): multiple volcanic events from EM1-FOZO-HIMU mantle sources within a single oceanic LIP, AGU Fall Meeting, San Francisco, California, USA.

Hoernle, K., Hauff, F., Werner, R., Mortimer, N., 2004. New Insights into the Origin and Evolution of the Hikurangi Oceanic Plateau (Southwest Pacific) from Multi-beam Mapping and Sampling. EOS Transactions 41, 401-408.

Hoernle, K., Hauff, F., Werner, R., van den Bogaard, P., Timm, C., Coffin, M.F., Mortimer, N., Davy, B., 2008. A similar multi-stage geochemical evolution for the Manihiki, Hikurangi and Ontong Java Plateaus?, AGU Fall Meeting, San Francisco, California, USA.

Ingle, S., Coffin, M.F., 2004. Impact origin for the greater Ontong Java Plateau? Earth Planet Sci Lett 218, 123-134.

Ingle, S., Mahoney, J.J., Sato, H., Coffin, M.F., Kimura, J.-I., Hirano, N., Nakanishi, M., 2007. Depleted mantle wedge and sediment fingerprint in unusual basalts from the Manihiki Plateau, central Pacific Ocean. Geology 35, 595-598.

Shipboard_Scientific_Party, 1976. Site 317, in: Schlanger, S.O., Winterer, E.L. (Eds.), Initial Reports. Deep Sea Drilling Project, Washington, D.C., pp. 161-300.

Taylor, B., 2006. The single largest oceanic plateau: Ontong Java–Manihiki–Hikurangi. Earth and Planetary Science Letters 241, 372-380.

Uenzelmann-Neben, G., 2012. The expedition of the research vessel "Sonne" to the Manihiki Plateau in 2012 (So 224). Berichte zur Polar- und Meeresforschung = Reports on polar and marine research 656, 70.

Werner, R., Hauff, F., 2007. FS Sonne Cruise Report SO 193: MANIHIKI Temporal, Spatial, and Tectonic Evolution of Oceanic Plateaus, IFM-GEOMAR Report. IFM-Geomar, Kiel, p. 201.

Werner, R., Hauff, F., Hoernle, K., Coffin, M.F., Party, S.S., 2007. R/V Sonne Cruise SO193: New Insights Into the Geodynamic History of the Manihiki Plateau, SW-Pacific, AGU Fall Meeting, San Francisco.

Winterer, E., Lonsdale, P.L., Matthews, J.L., Rosendahl, B.R., 1974. Structure and acoustic stratigraphy of the Manihiki Plateau. Deep-Sea research I 21, 793-814.

HOCHMUTH, K., GOHL, K., UENZELMANN-NEBEN, G., WERNER, R. (IN REVISION): THE DIVERSE CRUSTAL STRUCTURE AND MAGMATIC EVOLUTION OF THE MANIHIKI PLATEAU, CENTRAL PACIFIC. JOURNAL OF GEOPHYSICAL RESEARCH - ANLAGE 2

The diverse crustal structure and magmatic evolution of the Manihiki Plateau, Central Pacific

K. Hochmuth[1], K. Gohl[1], G. Uenzelmann-Neben[1], R. Werner[2]

[1] Alfred-Wegener-Institut Helmholtz-Zentrum für Polar- und Meeresforschung,
Am Alten Hafen 26, 27568 Bremerhaven, Germany
[2]GEOMAR Helmholtz-Zentrum für Ozeanforschung,
Wischhofstr. 1-3, 24148 Kiel, Germany

corresponding author: K. Hochmuth
E-mail: Katharina.Hochmuth@awi.de

Abstract:
The Manihiki Plateau is a Large Igneous Province (LIP) in the central Pacific. It is proposed to have been emplaced during the “Greater-Ontong-Java Event” in the early Cretaceous and deformed internally into three sub-provinces, possibly during the plate tectonic reorganization of the Pacific Ocean during the Cretaceous Normal Superchron. The Manihiki Plateau lays in the center of the Pacific LIPs and its investigation can therefore decipher tectonic mechanisms as well as the evolution of the plateau after its initial emplacement. By analyzing two seismic refraction/wide-angle reflection profiles crossing the two largest sub-provinces of the Manihiki Plateau, the High Plateau and the Western Plateaus, we give new insights into their crustal structure and magmatic evolution. The High Plateau shows a crustal structure of 20 km thickness and a seismic P-wave velocity distribution, which is comparable to other LIPs. The High Plateau experienced a strong secondary magmatism, which can be seen in relicts of seamount chain volcanism, basaltic flow units, and major magmatic pathways. The Western Plateaus on the other hand show only low volume secondary seamount magmatism and are mainly structured by fault systems and sedimentary basins. A constant decrease in Moho depth (9 km – 17 km) is a further indicator of crustal stretching on the Western Plateaus. Those findings lead to the conclusion, that the two sub-provinces of the Manihiki Plateau experienced a different magmatic and tectonic history. Whereas the High Plateau experienced massive secondary magmatism, the Western Plateaus underwent crustal stretching during and after its initial emplacement. This indicates, that the sub-provinces of the Manihiki Plateau play an individual part in the plate tectonic framework of the central Pacific.

1. Introduction:

Large Igneous Provinces (LIP) are large (>0.1 x 10^6 km^2) marine and terrestrial areas overprinted by massive volcanic activity (Bryan and Ernst, 2008; Coffin and Eldholm, 1994) LIPs have a great impact on the environment during and after their emplacement, for instance by the release of greenhouse gases during massive volcanism (Wignall, 2001) or their role in plate motion by thickening of oceanic crust (Bryan and Ernst, 2008; Miura et al., 2004). The emplacement of LIPs can for instance be linked to multiple global mass extinction events (Bryan and Ernst, 2008; Coffin and Eldholm, 1994; Courtillot et al., 1999; Larson and Erba, 1999; Tarduno, 1998; Wignall, 2001). LIPs are the result of massive magmatism occurring during a relatively short time period (1-5 M.y.) resulting in an anomalously thick mafic crust (Bryan and Ernst, 2008; Coffin and Eldholm, 1993; 1994; Wignall, 2001).This first magmatic stage of a LIP event is characterized by the emplacement of approximately 75% of its igneous volume (Bryan and Ernst, 2008; Bryan and Ferrari, 2013; Karlstrom and Richards, 2011; Miura et al., 2004). Later magmatic stages can be summarized as secondary magmatism and show longer emplacement periods as well as smaller emplacement rates. The crustal structure of oceanic LIPs commonly consists of a lower crustal layer with high seismic P-wave velocities (> 7.1 km/s), mafic intrusions in the middle crust, and volcanic flow units in the uppermost crust (Coffin and Eldholm, 1994; Ridley and Richards, 2010). In previous publications, the formation of LIPs on continents and oceans was explained by the impact of a mushroom-shaped plume head at the base of the lithosphere, leading to widespread magmatism (Morgan, 1971; White and McKenzie, 1995). But this scenario has been debated, since this plume head model cannot explain all characteristics of oceanic LIPs (Bryan and Ernst, 2008; Coffin et al., 2006; 2002; Courtillot et al., 1999; Korenaga, 2005; Larson and Erba, 1999; McNutt, 2006; Tarduno, 1998). The Manihiki Plateau, located in the western central Pacific, is such a LIP with different characteristics within its sub-provinces such as a laterally heterogeneous crustal character.

In this paper, we try to unravel the relationship between the two largest sub-provinces of the Manihiki Plateau, the Western Plateaus and the High Plateau, which are separated by the Danger Islands Troughs. Did both sub-provinces experience the same magmatic history? The fragmentation of the Manihiki Plateau poses the question, whether distinct phases of magmatic or tectonic processes led to the deformation of the Manihiki Plateau and which role the Danger Islands Troughs played in this scenario. By processing, modeling and interpreting recently acquired seismic refraction/wide-angle reflection data the deep crustal structure of both sub-provinces is revealed and interpreted for the first time.

2. Geological background

The formation of the Manihiki Plateau took place during the early Cretaceous (124-120 Ma) (Clague, 1976; Ingle et al., 2007; Timm et al., 2011) as centerpiece of the “Super-LIP” Ontong Java Nui according to Taylor (2006) and Chandler et al. (2012, 2013). This giant LIP covered more than 1% of Earth’s surface (Bryan and Ferrari, 2013; Coffin and Eldholm, 1993; Ingle et al., 2007; Timm et al., 2011; Wignall, 2001) and consisted of the Ontong Java Plateau, the Hikurangi Plateau and the Manihiki Plateau (Fig.1 (inlet)) as well as presumed multiple smaller fragments to the northeast and east of the Manihiki Plateau (“Greater Manihiki Plateau”), which have possibly been subducted (Larson et al., 2002; Viso et al., 2005). The plate tectonic reconstructions of Ontong Java Nui are still under debate, especially the fit between the Ontong Java Plateau and the Manihiki Plateau, as the spreading mechanisms of the Ellice Basin between the plateaus remain speculative (Chandler et al., 2012; Taylor, 2006). The coupled emplacement of the Hikurangi Plateau and the Manihiki Plateau appears much more certain, because the Osbourn Trough can be identified as the former spreading center separating the two plateaus (Billen and Stock, 2000; Downey et al., 2007; Worthington et al., 2006). The break-up of Ontong Java Nui took place shortly after its emplacement, since secondary magmatism, which can be related to later magmatic phases with small eruption rates and mainly seamount volcanism, shows different petrological and geochemical characteristics on the three remaining plateaus (Hoernle et al., 2010; Timm et al., 2011). This led to the conclusion, that after the coupled emplacement of the plateaus, the multiple stages of secondary magmatism were supplied by different mantle sources. Seafloor spreading was established possibly around 118 Ma in Ellice Basin (Ontong Java – Manihiki) (Chandler et al., 2012) and at the Osbourn Trough (Hikurangi – Manihiki) (Worthington et al., 2006). The onset and the duration of seafloor-spreading at both locations is difficult to date, since the complete activity of those spreading-centers has occurred within the Cretaceous Normal Superchron. Therefore, no magnetic seafloor-spreading anomalies can be traced, and smaller changes in the magnetic field cannot be observed due to the severe magmatic overprinting (Billen and Stock, 2000).

Figure 1

The Manihiki Plateau itself experienced deformation into three sub-provinces, the High Plateau, the Western Plateaus and the North Plateau (Winterer et al., 1974) (Fig.1). Further significant features of the Manihiki Plateau are the Danger Islands Troughs, which are created by rifting and/or transform forces separating the High Plateau from the Western Plateaus (Larson, 1997; Mahoney and Spencer, 1991; Winterer et al., 1974), the Suvarov Through, and the Manihiki Scarp, a former shearing zone at the eastern part of the plateau (Larson et al., 2002).

Hussong et al. (1979) published first estimates on the crustal structure of the Manihiki Plateau. However, their experiment, using seismic sonobuoy data, did not provide data of the lower crustal layers or the upper mantle, but it was inferred that the crust of the Manihiki Plateau was 3.1 times thicker than average oceanic crust and that it showed similar features as described for the crust of the Ontong Java Plateau (Hussong et al., 1979).

The upper basement and the sedimentary cover of the Manihiki Plateau has been studied from samples collected by multiple dredges (Beiersdorf et al., 1995a; Clague, 1976; Hoernle et al., 2010; Ingle et al., 2007; Schlanger et al., 1976; Timm et al., 2011; Werner et al., 2013) and short sediment cores. Drilling at Deep Sea Drilling Project (DSDP) Leg 33 Site 317 (Fig.1) reached a basaltic layer in a depth of 910 m below seafloor (Schlanger et al., 1976). $^{40}Ar/^{39}Ar$ dated tholeiitic basement basalts of the Manihiki Plateau reveal a mean age of 124.6 ± 1.6 Ma (Timm et al., 2011). Lava drilled at DSDP Site 317 of the High Plateau exhibit an age of 117 Ma (Hoernle et al., 2010). In the Danger Islands Troughs, tholeiitic basalts ($^{40}Ar/^{39}Ar$ age of 117.9 ± 3.5 Ma) with an unusual composition are present, as well as alkalic basalts possibly related to later volcanic stages ($^{40}Ar/^{39}Ar$ age of 99.5 ± 0.7 Ma) (Ingle et al., 2007).

Later stages of episodic volcanism on the Manihiki Plateau are also manifested by multiple seamounts and basaltic flow units on the High Plateau (Beiersdorf et al., 1995b; Coulbourn and Hill, 1991) and seamounts on the margins of the Western Plateau (Hoernle et al., 2009; Sandwell and Smith, 1997; Werner and Hauff, 2007). Seismic reflection data of the High Plateau image several volcaniclastic layers in the lower sedimentary column (Ai et al., 2008; Schlanger et al., 1976; Winterer et al., 1974), pointing to a shallow or subaerial environment during the secondary phases of magmatism (Ai et al., 2008). Pietsch and Uenzelmann-Neben (in prep.) additionally identify magmatic events occurring within younger strata (< 45 Ma), which leads to the conclusion, that the Manihiki Plateau experienced a period of magmatic reactivation, possible due to the overriding of multiple hotspots.

It is important to note, that most of the publications have focused on the High Plateau along with the Danger Islands Troughs and the Manihiki Scarp. Other areas of the Manihiki Plateau, such as the basement of the Western Plateaus, are poorly sampled and therefore the evolution of the different sub-provinces and their magmatic and tectonic relationship to the other parts of the plateau is still poorly understood. Although satellite-derived gravity anomaly maps (Sandwell and Smith, 1997) indicate different crustal structures of the sub-provinces, plate tectonic reconstructions of the Cretaceous western Pacific (e.g. Chandler et al., 2012; 2013; Davy et al., 2008; Musgrave, 2013; Taylor, 2006) treat the Manihiki Plateau as a single tectonic block and disregard its different sub-provinces.

3. Data acquisition, processing, and modeling parameterization

3.1 Data acquisition

During RV *Sonne* cruise SO-224 in 2012 (Uenzelmann-Neben, 2012), the Alfred Wegener Institute (AWI) collected two deep crustal seismic refraction/wide-angle reflection lines (Fig. 1), crossing the two main sub-provinces of the Manihiki Plateau: the Western Plateaus (AWI-20120100) and the High Plateau (AWI-20120200). Both seismic lines consisted of 28 ocean-bottom seismometer (OBS) and 5 ocean-bottom hydrophone (OBH) stations spread out over 500 km long profiles each. The OBS stations were equipped with a 3-component (4.5 s natural period) seismometer and a hydrophone. OBH stations are only equipped with a hydrophone. Station spacing varied between a constant 14.1 km on AWI-

20120100 and 10.9 km (at the margins) to 24.1 km (on the High Plateau) on AWI-20120200 to ensure better ray-coverage at the plateau margins.

An array of 8 G-Guns™ (type 520) with a total volume of 68 liters (4160 in^3) was used as a seismic source. The G-Guns™ were towed in four clusters of two guns each at 10 m water-depth and fired at nominally 200 bar operating pressure every full minute. During the acquisition, multichannel seismic reflection data was also recorded with a 3000 m long digital solid streamer (Sercel Sentinel™) of 240 channels. In addition to the seismic experiments, multibeam bathymetry data were collected throughout the cruise with a *Kongsberg Simrad EM120,* which is permanently installed on RV *Sonne.*

3.2 Processing and modeling of seismic refraction/wide-angle reflection data

The OBS/OBH field data were merged with navigation data and converted to SEGY-format. After relocalization of the OBS positions using the direct-wave arrivals, the application of a bandpass filter (6-15 Hz), all refracted and reflected arrival phases were picked with the software *ZP* (Zelt, 2004). We assigned picking individual uncertainties to the individual picks taking the signal-to-noise-ratio into account (Zelt, 2004). The picking uncertainties range between 0.075 and 0.25s with a median picking uncertainty of 0.1s.

P-wave and S-wave velocity-depth modeling was carried out using the forward modeling software *Rayinvr* (Zelt and Smith, 1992) and its graphical interface *P-Ray* (Fromm, 2012). We used bathymetric and seismic reflection records obtained during the cruise (Pietsch and Uenzelmann-Neben, submitted; Uenzelmann-Neben, 2012) , in order to constrain modeling parameters for the seafloor and the thickness of the sedimentary cover. The conversion from two-way travel times of the reflection seismic records to the actual depths below seafloor was carried out with the velocity field established by velocity analysis (Pietsch and Uenzelmann-Neben, submitted). The layer boundaries and model coordinates of the velocity nodes established by the P-wave model were used for S-wave modeling to ensure comparable models and allow the calculation of the Poisson's ratio. The set-up of the initial model included the following constraints: the topography of the seafloor, the thickness of the sedimentary column and the crustal structure previously published by Hussong et al. (1979) and that derived from 1D-Profiles at our OBS/OBH-stations.

Refracted and reflected phases in the P- and S-wave models are named respectively to their corresponding layer P_{layer} and $P_{layer}P$/ S_{layer} and $S_{layer}S$. Reflected phases always represent the reflection at the base of the layer. Mantle phases are called P_n/S_n for the refracted phase and P_mP/S_mS for the reflection at the crust-mantle boundary (Moho). We observed three distinct groups of crustal phases (P_{uc}/S_{uc} (upper crust), P_{mc}/S_{mc} (middle crust), P_{lc}/S_{lc} (lower crust)) on the Western Plateaus (Fig. 2, Fig. 3,) and four distinct groups of crustal phases (P_{uc}/S_{uc} (upper crust), P_{umc}/S_{umc} (upper-middle crust), P_{lmc}/S_{lmc} (lower-middle crust), P_{lc}/S_{lc} (lower crust)) on the High Plateau (Fig. 4, Fig. 5). The resolution of the S-wave velocity models is generally lower than the resolution of the presented P-wave velocity models. Unfortunately, only little to no information on the S-wave velocity distribution was returned from the lower crust and the mantle. We assessed the uncertainty of layer boundaries of the intercrustal reflections (Tab. 1). The depth uncertainties of the individual layers range between 0.4 and 0.8 km, with the largest uncertainties for the Moho depths. This is well in the uncertainty range given by the ray-tracing method.

The Poisson's ratio can contribute further parameters on the composition of the crust (Christensen, 1996). For example, alteration processes such as serpentinization or the distribution of predominantly mafic and felsic rocks can be constrained. We calculated the Poisson's ratio for every velocity node of the P-and S-wave models and gridded their distribution. Since the S-wave model has a sparser resolution than the P-wave model, it limits the resolution of the Poisson's ratio model. For AWI-20120100, the calculated Poisson's ratio is not presented due to the poor resolution of this profile.

3.3 Modeling of gravity data:

As shipborne gravity data were not collected, we relied on free-air gravity anomaly records extracted from the global satellite-derived gravity anomaly grid by Sandwell and Smith (1997) with values extracted along our seismic profiles. In a starting model using the forward modeling software IGMAS (Götze et al., 2002), we incorporated the layer boundaries of the P-wave velocity-depth model and converted the P-wave velocities to rock densities using the relationships by Hamilton (1978) and Barton (1986). We assigned rock densities to P-wave velocity clusters representing the different crustal layers of the P-wave velocity models (Tab. 2). As this 2-D approach turned out to be affected by large-amplitude anomaly features offline, a perfect fit could not be achieved by retaining realistic model parameters such as the seafloor topography. In several iterations, the model parameters were altered to generate a best fit to the observed the gravity anomalies and the P- and S-wave models.

Table 1
Table 2

4. Results of data analysis and modeling

4.1. Bathymetric and sedimentary features

The two seismic refraction/wide-angle reflection profiles allow a comparison of the two main provinces of the Manihiki Plateau. Profile AWI-20120100 crosses the Western Plateaus from the northwest to the southeast (Fig.1). The water depth ranges from 4800 m in the Tokelau Basin to 3800 m on the Western Plateaus. The deepest areas of the profile are the Danger Islands Troughs with a maximum depth of 4900 m. A small part of the westernmost High Plateau is also covered at the eastern end of this profile. Bathymetric and seismic reflection data reveal a rough seafloor with multiple faults and graben systems (Fig.6). Sedimentary cover is mainly restricted to the basins between the numerous basement highs (Uenzelmann-Neben, 2012; Winterer et al., 1974).

Profile AWI-20120200 images the High Plateau of the Manihiki Plateau in a west to east section (Fig.1). The water depth ranges from 5100 m in the Samoan Basin and Penrhyn Basin to a mean depth of the High Plateau of 2500 m with a relatively smooth seafloor. The Manihiki Scarp in the east and the margin towards the Samoan Basin in the west form sharp flanks bordering the sub-province. Seismic reflection data reveal an approximately 800 m thick sedimentary cover and several basement highs on the central plateau. Sedimentation appears to be restricted to pockets and small basins between basement highs on the margins of the High Plateau (Ai et al., 2008; Beiersdorf et al., 1995b; Pietsch and Uenzelmann-Neben, submitted; Winterer et al., 1974).

4.2. Upper crustal layers

The upper crustal layer of the Western Plateaus and the High Plateau include the acoustic basement of the seismic reflection data and are represented by the refracted phases (P_{uc}/S_{uc}) and constrained by the wide-angle reflections at the upper layer boundary P_sP and $P_{uc}P$ at the base of the layer (Figs. 2, 3, 4 and 5).
Figure 2
Figure 3
Figure 4
Figure 5
Figure 6
On the Western Plateaus, the upper crust is characterized by a rough basement topography and a widely variable velocity field. The western 50 km of the profile consist of an upper crustal layer with P-wave velocities between 5.2 and 5.5 km/s (Fig. 7). Between 50 km and 300 km profile distance, P-wave velocities decrease to 4.5 km/s with small patches of lower velocities (3.8 km/s). It is important to note that due to the malfunction of two neighboring stations (st19, st20) the ray coverage is not ideal in this area. Between 300 and 370 km profile

distance, unusual low velocities ranging from 3.0 to 4.3 km/s are present. S-wave velocities show a block-like structure (Fig. 8) rather than a continuous change. From 0 to 100 km, S-wave velocities range from 2.9 to 3.2 km/s. Velocities decrease to 2.5 to 2.8 km/s between st8 and st12. From 180 km to 250 km profile distance, S-wave velocities from 2.9 to 3.3 km/s are common. This block is followed by an area of unusual high S-wave velocities, which corresponds to higher velocities in the middle crust (3.7 to 4.3 km/s). The Danger Islands Troughs area lacks a visible upper crust (Figs. 2b,and 7).

Figure 7

Figure 8

The High Plateau shows higher and more homogenous P-wave velocities between 4.7 and 5.6 km/s (Figs. 4 and 9) and S-wave velocities between 2.7 and 3.2 km/s (Figs. 5 and 10). Several extrusive and intrusive magmatic features can be identified in the seismic reflection data (Pietsch and Uenzelmann-Neben, submitted). Those magmatic centers can be connected to areas of higher P-wave velocities (e.g. at st21) in the P-wave velocity model.

Figure 9

Figure 10

The upper crust of the surrounding ocean basins close to the margins of the Manihiki Plateau, the Penrhyn Basin and the Samoan Basin show P-wave velocities between 5.2 and 5.8 km/s (Fig. 9). S-wave velocities range from 3.3 km/s in the Samoan Basin to 2.9 km/s in the Penrhyn Basin (Fig. 10). Whereas the Samoan Basin shows Poisson's ratio values of around 0.25, which can be related to basalt (Christensen, 1996), the Penrhyn Basin on the other hand shows high Poisson's ratio values of over 0.30 (Fig. 11), which is typical for basalts and partly serpentinized basalts in the uppermost crust.

Figure 11

4.3. Middle crustal layers

The middle crustal layers of the Manihiki Plateau consist of two layers (P_{umc}/ S_{umc} from upper-middle crust and P_{lmc}/ S_{lmc} from lower-middle crust) on the High Plateau (Figs. 4 and 5) and one layer on the Western Plateaus (P_{mc}/S_{mc}) (Figs. 4 and 5). On the Western Plateaus, the P-wave velocity structure of the middle crust is homogeneous and ranges between 5.8 and 6.8 km/s (Fig. 7). In between the ridge-like structures at 230 and 270 km profile distance, the upper crustal layer is absent and the top of the middle crustal layer represents the acoustic basement. This feature is also present in the gravity anomaly model (Fig. 12). The S-wave velocity model shows more variation in the middle crust, with values ranging from 3.6 to 4.0 km/s in the west to very high S-wave velocities of up to 4.3 km/s at st22 (Fig. 8). S-wave velocities decrease to 3.8 to 4.0 km/s towards the Danger Islands Troughs.

The middle crust is present at the Danger Islands Troughs. East of the troughs, the middle crust is divided into two separate crustal layers, the upper-middle crust and the lower-middle crust (Figs. 2b and 7). The upper-middle crust shows P-wave velocities between 5.4 and 6.4 km/s (Fig. 7) and S-wave velocities between 3.0 and 3.6 km/s (Fig. 8). This crustal layer is thicker in the western part of the High Plateau. The lower-middle crust presents relatively high P-wave velocities (6.7 to 6.9 km/s) and S-wave velocities between 4.0 and 4.3 km/s (Figs. 9 and 10). The lower-middle crustal layer crops out at the seafloor at the Manihiki Scarp. Here, S-wave velocities up to 4.65 km/s are present. The Poisson's ratio shows three distinct areas of higher values (> 0.26) below the central plateau (Fig. 11). Those can be connected to areas of higher P-wave velocities and therefore the magmatic structures within the upper crust. Middle crustal layers are not present in the adjacent oceanic basins.

Figure 12

4.4. Lower crustal layers

The lower crust of most LIPs consists of a zone of high P-wave velocities (>7.1 km/s), a so called high-velocity zone (HVZ) (Coffin and Eldholm, 1994; Ridley and Richards, 2010).

On the Manihiki Plateau, we observe this HVZ on both sub-provinces within the lower crust (P_{lc}/S_{lc}). The HVZ is also present in the lower crust of the Tokelau Basin and extends into the Samoan Basin. At the Manihiki Scarp, the HVZ terminates abruptly. P-wave velocities range between 6.8 and 7.8 km/s on the Western Plateaus (Fig. 7) and 7.0 to 7.8 km/s on the High Plateau (Fig. 9). The ray-coverage of S-waves is poor on the High Plateau and only a few interpretable signals were returned for the lower crust of the Western Plateaus. S-wave velocities range between 4.1 and 4.3 km/s on the Western Plateaus (Fig. 8). At the western margin of the High Plateau, S-wave velocities reach 4.4 km/s and at the Manihiki Scarp values of 4.7 km/s are present (Fig. 10).

4.5. Crust-mantle boundary and mantle

The boundary between the crystalline crust and the mantle (Moho) can be constrained by refracted phases from the uppermost mantle (P_n) as well as reflections at the Moho itself (P_mP) (Figs. 2 and 4). On the High Plateau, the Moho is visible in the data throughout the central plateau at a depth of 20 km below the seafloor (Figs. 9 and 13). In the Samoan Basin and the Penrhyn Basin the crustal thickness, the combined thickness of the sedimentary column and the igneous crust, changes rather abruptly to only 5 km. Those depths for the crust-mantle boundary are also consistent with gravity anomaly modeling (Fig. 13). The nature of the upper mantle can be inferred from P_n refraction phases. The uppermost mantle shows normal mantle P-wave velocities of 8.1 km/s. The P-wave velocities of the mantle below the Western Plateaus are slightly higher with 8.2 km/s. The crustal thickness of the Western Plateaus ranges between 9.3 km in the Tokelau Basin to 17.2 km east of the Danger Islands Troughs (Figs. 7 and 12).

Figure 13

5. Two different magmatic and tectonic regimes on the Manihiki Plateau

In general, the crust of the Manihiki Plateau is severely faulted at its margins and by the troughs intersecting the plateau (Ai et al., 2008; Pietsch and Uenzelmann-Neben, submitted; Winterer et al., 1974), likely due to plate-tectonic reorganization of the Pacific Ocean during the Cretaceous Normal Superchron. In previous works, the Western Plateaus were assumed to be of a similar structure as the High Plateau (Hussong et al., 1979; Viso et al., 2005) due to the lack of data from deep crustal layers. By interpreting our P- and S-wave velocity and gravity models along with the Poisson's ratio model, the crustal structure of the different sub-provinces as well as their individual margins and adjacent ocean basins can be further constrained.

Both sub-provinces exhibit a HVZ in the lower crust. These HVZs with velocities higher than 7.1 km/s are also common at other oceanic LIPs (Coffin and Eldholm, 1994; Coffin et al., 2006; Richards et al., 2013; Ridley and Richards, 2010) and can be related to the presence of gabbros as well as the olivine and pyroxene cumulates (Karlstrom and Richards, 2011; Ridley and Richards, 2010) (Figs. 14 and 15). The crustal thicknesses of the Western Plateaus and the High Plateau differ. Whereas the High Plateau shows a constant crustal thickness of 20 km, which is comparable to other oceanic LIPs such as the Agulhas Plateau and southern Mozambique Ridge (Gohl and Uenzelmann-Neben, 2001; Gohl et al., 2011; Uenzelmann-Neben et al., 1999) with steep boundaries towards the adjacent oceanic basins, the Western Plateaus show a gradual decrease in the crustal thickness from 17.2 km in the east to 9 km in the west. A clear boundary between normal oceanic crust in the Tokelau Basin and the oceanic LIP cannot be identified because the HVZ is still present below the Tokelau Basin.

The middle crust of both sub-provinces shows similarities in the P-and S- wave velocity structure. The middle crust of the Western Plateaus has a velocity field, which is comparable to the lower-middle crust of the High Plateau. This velocity and density distribution can be associated with gabbroic intrusions (Christensen, 1996; Coffin and Eldholm, 1994; Richards et al., 2013; Ridley and Richards, 2010). On the High Plateau, the upper-middle crust shows a slightly different velocity field and a small decrease in density (Tab. 1). This transitional layer is not present in the Western Plateaus. Additionally, the middle crust underlying the central High Plateau exhibits three large areas of higher Poisson's ratio, which are correlated to former magmatic centers observed in the seismic reflection data (Fig. 11) (Pietsch and Uenzelmann-Neben, submitted). Those "chimneys" mark former magmatic pathways towards the seafloor of the High Plateau, possibly during a later volcanic stage (Fig. 15) (Pietsch and Uenzelmann-Neben, submitted). Karlstrom and Richards (2011) published a model of LIP crustal magma transport, which indicates the formation of individual upward migrating sills during later stages of magmatic activity. Similar structures as e.g. magmatic centers and massive basaltic flow units cannot be observed on the Western Plateaus. The secondary (or late stage) magmatic activity on the Western Plateau is restricted to low volume seamount volcanism.

The upper crust of the two sub-provinces differs significantly. The High Plateau consists of countless volcanic centers, which are partly eroded and covered by volcaniclastic and pelagic sedimentary rocks (Ai et al., 2008; Beiersdorf et al., 1995b; Pietsch and Uenzelmann-Neben, submitted). Thus, the upper crust consists of basaltic flow units as well as subvolcanic intrusions (Karlstrom and Richards, 2011), possibly interlayering each other as a result of the multiple stages of secondary magmatism. Massive fault systems are only present at the margins of the High Plateau (Fig. 15) (Ai et al., 2008; Pietsch and Uenzelmann-Neben, submitted; Viso et al., 2005; Winterer et al., 1974). At the Manihiki Scarp, deeper crustal layers, most likely consistent with the lower-middle crust, crop out at the seafloor in multiple ridges separated by deep reaching faults. Based on gravity data, Viso et al. (2005) postulated that the upper crust was serpentinized in this area. We can further support this assumption by high values in the Possion's ratio (Fig. 11) and lower densities (Fig. 13) at the Manihiki Scarp and the Penrhyn Basin and interpret mafic material which has been partly serpentinized in this area. The margins towards the Samoan Basin and the Samoan Basin themselves show basaltic flow units and have no indications for large-scale serpentinization in this area.

Whereas the High Plateau shows clear indications for multiple magmatic phases within its upper crust, the data from the Western Plateaus present possible basaltic flow units interlayered with mafic intrusions comparable to the upper crust of the High Plateau only in the Tokelau Basin. The upper crust of the Western Plateaus is heavily structured by fault systems and horst and graben features (Fig. 14) and seamounts indicate only local secondary magmatic activity. P- and S-wave modeling as well as gravity anomaly modeling reveal the presence of diverse rock types ranging between igneous rocks and sedimentary rocks. It can be debated whether the upper crust consists of volcaniclastic deposits or even massive carbonate banks (Grevemeyer et al., 2001). The majority of the faults are normal faults, which is consistent with crustal stretching processes (Fig. 6). The thinning of the crust of the Western Plateaus could also be due to its greater distance from the center of emplacement of the plateau. On the Ontong Java Plateau, this mode of crustal thinning can be observed at its northern boundaries. The local bathymetry indicates a relative smooth and constant lowering towards the oceanic basin, consistent with basaltic flow patterns (Mochizuki et al., 2005). On the Western Plateaus the decrease of crustal thickness is stepwise, dividing the Western Plateaus in terraces. This could not have been achieved by the lack of magmatic material. Another indicator for crustal extension is the constant decrease in Moho depth towards the Tokelau Basin. Therefore, tectonic deformation seems to have modified the plateau's structure enormously. Volcanic extrusions, e.g. the occurrence of basaltic flows, can only be interpreted for the Tokelau Basin and are only locally present on the Western Plateaus (Fig. 14). The southern Western Plateaus are fringed by seamounts which can be related to a phase of magmatic reactivation (< 45 Myr), similar to the islands and prominent seamounts on the High Plateau (Pietsch and Uenzelmann-Neben, submitted) Sedimentary basins as well as horst and graben structures form the dominant features of the central Western Plateaus.

The Danger Islands Troughs dissect the Manihiki Plateau into the two main sub-provinces. The troughs are characterized by the absence of typical upper crustal material (Fig. 14). The lower crustal layers as well as the crust-mantle boundary show only few relicts of former tectonic activity such as an updoming of the mantle at this position or a thinning of the lower crust. In the middle crust, the Danger Islands Troughs mark the distinct change over between a single layer middle crust on the Western Plateaus and the upper-middle crust and lower-middle crust of the High Plateau (Figs. 2b and 14). The absence of the upper crustal material and the lack of an updoming of the mantle suggest, that the Danger Islands Troughs are formed as a series of pull-apart basins. The previously suggested formation as an initial rift system (Taylor, 2006; Winterer et al., 1974) seems unlikely since upper crustal material is absent and fault systems indicate a shearing component. Additionally, the Danger Islands Troughs separates two structural regimes, the mainly tectonically deformed Western Plateaus and the magmatically altered High Plateau.

Figure 14
Figure 15

6. Discussion:
Western Plateaus vs. High Plateau – two different crustal environments of the Manihiki Plateau

This paper presents the first comprehensive insight into the crustal structure of the two major sub-provinces of the Manihiki Plateau. The results of our analyses revise earlier models. The Western Plateaus and the High Plateau of the Manihiki Plateau exhibit similar features, such as a continuous HVZ within the lower crust, but also show large differences, especially in the upper crustal layers and the crustal thicknesses.

6.1 Comparison with other oceanic LIPs

By comparing the Western Plateaus and the High Plateau with other oceanic LIPs, the crustal thickness and velocity-depth distribution of the High Plateau resembles those of the Kerguelen Plateau (Charvis and Operto, 1999), Broken Ridge (Francis and Raitt, 1967), Agulhas Plateau (Gohl and Uenzelmann-Neben, 2001; Parsiegla et al., 2008) and Mozambique Ridge (Gohl et al., 2011) (Fig.16). The proportional velocity-depth distribution is also similar to that of the southern Ontong Java Plateau (Miura et al., 2004), but the Ontong Java Plateau exhibits up to twice the crustal thickness.

A dominant feature of all oceanic LIPs is the HVZ of the lower crust with P-wave velocities ranging between 7.1 and 7.7 km/s (Fig.16) (Coffin and Eldholm, 1994; Ridley and Richards, 2010). Both the Western Plateaus and the High Plateau show such high P-wave velocities in the lower crust. The HVZ is also continuous between the two sub-provinces in the Danger Island Troughs area. High velocities are also present in the lower middle crust (P-wave velocities between 6.3 and 7.0 km/s). The continuous presence of the HVZ suggests a process, in which the Manihiki Plateau was emplaced as a single continuous and larger LIP ("Greater Manihiki Plateau") during a first volcanic phase.

The High Plateau along with other oceanic LIPs has an upper middle crustal layer of P-wave velocities 5.0 to 5.8 km/s (yellow to orange colors in Fig.16). This layer is present on all previously presented LIPs as part of the middle crust but is not always resolved as an individual crustal unit constrained by intracrustal reflections. Surprisingly, this transitional layer with this specific velocity range is not present on the Western Plateaus (Fig.16) leaving a large gap in the normally relative continuous P-wave velocity distribution within the igneous crust of a LIP. Therefore the Western Plateaus are missing a crustal layer present in all other oceanic LIPs, a layer associated with mafic intrusions formed during later stages of magmatic activity (Karlstrom and Richards, 2011).

The uppermost crust of oceanic LIPs mainly consists of basaltic flow units interlayered with pillow lavas and possibly sedimentary layers formed by regional secondary volcanic phases (Hoernle et al., 2010; Inoue et al., 2008; Timm et al., 2011). The High Plateau shows local volcanic centers, which have been the outlets of extrusive volcanism during secondary magmatic phases, similar as observed on the Ontong Java Plateau (Inoue et al., 2008) or the

Hikurangi Plateau (Davy and Wood, 1994; Davy et al., 2008; Hoernle et al., 2004). Extrusive and intrusive magmatic activity is also visible in seismic reflection data throughout the High Plateau (Pietsch and Uenzelmann-Neben, submitted). Buried seamount chains can also be located in gravity anomaly grids on the High Plateau of the Manihiki Plateau (Fig. 13).

On the contrary, the upper crust of the Western Plateaus consist of several areas of very low velocities and low rock densities that are associated with volcaniclastic sedimentation and carbonate banks rather than with igneous rocks (Grevemeyer et al., 2001). Seamounts and seaknolls are mainly located at the margins and result from a later magmatic reactivation phase (< 45 Ma) possibly due to riding over the Tahiti and Society Islands hotspots. Neither the seismic refraction/wide-angle reflection data nor the multichannel seismic reflection data acquired on the central Western Plateaus indicate extensive secondary magmatism such as the emplacement of basaltic flow units. Short wavelength gravity anomalies are mainly attributed to fault complexes, ridge systems and seamounts (Figs. 6 and 12). Basaltic flow units cropping out at the seafloor are only present in the Tokelau Basin. On the central Western Plateaus, deep basins of lower velocities are imbedded into mafic material.

Therefore, the Western Plateaus of the Manihiki Plateau do not have the typical crustal structure of an oceanic LIP and lack features such as mafic intrusions in the upper-middle crust (Fig. 16). Although, it is important to note that a main characteristic of a LIP, the HVZ, is present in the lower crust of the Western Plateaus. The High Plateau shows key features previously described for oceanic LIPs. The two main sub-provinces of the Manihiki Plateau hence must have undergone a different magmatic and tectonic evolution after their emplacement.

Figure 16

6.2 The role of the HVZ in the lower crust of the Manihiki Plateau

The HVZ in the lower crust is a key component of a LIP (Coffin and Eldholm, 1994; Ridley and Richards, 2010). The HVZ is a result of the upwelling of very hot mantle material and consists of an olivine and pyroxene crystal fraction captured above the Moho (Karlstrom and Richards, 2011; Ridley and Richards, 2010). On the Manihiki Plateau, the HVZ is present within the whole lower crust and is not intersected at the Danger Islands Troughs (Fig. 14). This leads to the conclusion that the Manihiki Plateau was emplaced as a single crustal unit, and that the formation of different sub-provinces is restricted to the upper and middle crustal layers. Furthermore, the HVZ indicates various tectonic forces acting on the margins of the Manihiki Plateau. Our data cover three different margins, the Manihiki Scarp, the margin towards the Samoan Basin and the Tokelau Basin. At the Manihiki Scarp, the HVZ terminates abruptly (Figs. 9 and 15). The Penrhyn Basin does not show high P-wave velocities in its lower crust. Tectonically, the Manihiki Scarp is a strike-slip zone (Larson et al., 2002; Viso et al., 2005), where a fragment of the Manihiki Plateau, which is now subducted, was emplaced. The sudden termination of the HVZ illustrates this razor-sharp boundary between the LIP and the surrounding ocean basin.

This clear distinction between LIP-related crust and surrounding oceanic basins can neither be seen in the Samoan Basin nor in the Tokelau Basin. The Samoan Basin shows a thin HVZ (Figs. 9 and 15). Therefore, the margin of the High Plateau and the adjacent ocean basin are plume/hotspot-influenced and do not only result from the break-up between the Manihiki Plateau and the Hikurangi Plateau. By the reexamination of reflection seismic, gravity and bathymetry data we conclude, that the Samoan margin experienced a volcanic overprint during a secondary magmatic stage and a small-scale stretching component. It is important to note that this crustal extension of the High Plateau is only present at the Samoan margin and possibly at the northern margin. Stretching of the Western Plateaus is notable throughout the plateau and not only at their margins.

On the Western Plateaus, the Tokelau Basin shows a clear HVZ in its lower crust (Figs. 7 and 14). This leads to the conclusion that the crust present in the Tokelau Basin is in fact still part of the Manihiki Plateau and was not produced as a part of the "normal" oceanic seafloor

spreading in the Ellice Basin. The stepwise decrease in Moho depth as well as in the thickness of the HVZ is a further indicator of the massive crustal stretching of the Western Plateaus.
The different margins of the Manihiki Plateau – transform, crustal stretching with volcanic overprint, massive crustal stretching with small volcanic overprint – illustrate how multifaceted the influence of the tectonic mechanisms is on the crustal structure of the Manihiki Plateau. The Manihiki Plateau allows a good insight into the tectonic framework of the Cretaceous Pacific, having been affected by different tectonic mechanisms.
By close evaluation of the measured and modeled P- and S-wave velocities, we can compare our findings to the results of Richards et al. (2013). The authors modeled multiple batch melting processes recreating the high-velocity, ultramafic, intrusive body below hotspot tracks. These intrusive features are - at smaller scale - similar to a HVZ below an oceanic LIP. Applying the calculations by Richards et al. (2013), our P- and S- wave velocities sugguest an emplacement scenario for the HVZ below the High Plateau with a 30% liquid melt fraction of a FeO portion of less than 5% weight. MgO ranges presumably at 15-20% weight. This would result in a potential melting temperature of 1900°K.

6.3 Emplacement scenario of the Manihiki Plateau and its two major sub-provinces

The initial magmatic event of a LIP emplaces approximately 75% of the total igneous volume of the igneous province (Bryan and Ernst, 2008; Karlstrom and Richards, 2011), therefore it is consistent that the lower crustal layers as well as the middle crustal layers show similar attributes on both sub-provinces of the Manihiki Plateau. The lower crust and the lower middle crust is assumed to be emplaced during the “Greater Ontong Java Nui event” creating the LIPs in the western Pacific Ocean (Chandler et al., 2012; 2013; Taylor, 2006). The Manihiki Plateau was emplaced as a “Greater Manihiki Plateau”, including all sub-provinces and eastern and northern fragements. To the south the Hikurangi Plateau was emplaced simultaneously. The High Plateau of the Manihiki Plateau was emplaced at subaerial level or close to sea level (Ai et al., 2008).

Rifting of “Greater Manihiki Plateau”, followed this initial magmatic phase (Chandler et al., 2012; 2013; Hoernle et al., 2004; 2010; Taylor, 2006; Timm et al., 2011; Worthington et al., 2006) (Fig.17). The Danger Islands Troughs, which separate the High and the Western Plateaus, formed during this time (Ingle et al., 2007; Timm et al., 2011) along with several fault systems at the Western Plateaus (Fig. 6) initiating the proposed break-up between the Ontong Java Plateau and the Manihiki Plateau at several locations (Fig. 14). The break-up between the Ontong Java Plateau and the Manihiki Plateau included a rotational component (Chandler et al., 2013), which might have resulted in crustal stretching. Although there is no clear evidence for crustal separation between the Ontong Java Plateau and the Manihiki Plateau, we still support the hypothesis by Taylor (2006) as a general concept. One of the main concerns with the joined emplacement is the major difference in crustal thickness between the Ontong Java Plateau and the Western Plateau. Here it is important to note that no refraction seismic experiments were carried out at the eastern margin of Ontong Java, which resembles the Western Plateaus within gravity anomaly maps. If the reconstructions by Chandler et al. (2012) and Taylor (2006) are incorrect, the Western Plateaus need to be deformed by other forces such as the formation of the Ellice Basin. On the eastern margin of the Manihiki Plateau, a massive north-south trending transform fault creates the Manihiki Scarp (Fig. 1) (Larson et al., 2002; Viso et al., 2005). To the south, the Hikurangi Plateau rifted from the High Plateau of the Manihiki Plateau by the development of the Osbourn Trough ocean spreading center (Billen and Stock, 2000; Davy et al., 2008; Worthington et al., 2006).

With the full establishment of seafloor spreading in the Ellice Basin, the Western Plateaus experienced less tectonic deformation. As faults from the deeper crust to the seafloor can be detected (Figs. 6 and 14), the crustal stretching forces have been present throughout the different stages of the evolution of the Western Plateaus.

Simultaneously, a secondary magmatic stage started resulting in seamount volcanism and the emplacement of basaltic flows (Ai et al., 2008; Pietsch and Uenzelmann-Neben, submitted; Winterer et al., 1974) on the High Plateau (Fig. 17). We interpret relicts of volcanic centers as well as the magmatic pathways within the crust from our data. The secondary magmatism formed the uppermost crust of the Manihiki Plateau by intrusions and extrusive volcanism mainly of alkalic composition (Clague, 1976; Hoernle et al., 2010; Ingle et al., 2007; Timm et al., 2011). At the Danger Islands Troughs, alkaline lavas with a strong enrichment in light rare earth and large-ion lithophile elements were emplaced (Ingle et al., 2007).

At the Western Plateaus, we could not identify evidence for massive secondary magmatism in our data. Localized seamount volcanism is present throughout the Western Plateaus with the majority of seamounts at its margins, but there are no indications for massive basaltic flow units and a leveling and smoothing of the basement as visible on the High Plateau (Pietsch and Uenzelmann-Neben, submitted). Of course, upper crustal layers created during second magmatic stages could have been eroded as they were close to the sea surface. In this case, erosion should have been very effective, and magmatic pathways and late stage mafic intrusions should be visible in the middle crustal layers of the presented models. Also, the unusual low velocities of parts of the acoustic basement infer that the Western Plateaus experienced volcaniclastic and carbonate sedimentation (Fig. 14). As no rock samples from the central Western Plateaus are available for ground-truthing, we rely only on the geophysical parameters. The source of the volcaniclastic sedimentation is located most likely on the High Plateau, since high volume magmatic activity along with erosional processes can be reported from the High Plateau. Carbonate sedimentation, e.g. the build-up of carbonate platforms, is also likely since the Manihiki Plateau did not reside below the Carbonate Compensation Depth (Van Andel, 1975) during its entire lifespan. It is common, that carbonate platforms form on fault blocks created by rifting processes and are later on buried by sedimentation (Bosence, 2005). Since the upper crust, consisting of sedimentary rocks, thins towards the west, and magmatic rocks become present again, it is likely that this portion of the Western Plateaus had already subsided deeper and was farther away from the source region of volcaniclastic sedimentation. Therefore, the western Western Plateaus, towards the Tokelau Basin experienced less sedimentation, and magmatic rocks are exposed in the upper crust.

During a later stage of secondary magmatism, its activity on the High Plateau weakened and moved its activity center to the east and towards the margins, as eastern volcanic structures are better preserved and early deformed sedimentary layers are visible in seismic reflection data (Ai et al., 2008; Pietsch and Uenzelmann-Neben, submitted) (Fig.17). Older volcanic edifices are eroded, leading to an almost smooth acoustic basement in the seismic reflection data (Pietsch and Uenzelmann-Neben, submitted). The eroded material is deposited on the High Plateau, but also becomes reworked and integrated into the magma surfacing at the Danger Islands Troughs (Ingle et al., 2007). Here, tholeiitic basalts with a U-shaped incompatible element pattern and unusually low abundance of several elements are formed. These basalts have not been reported so far anywhere else on the Manihiki Plateau or the other parts of the “Greater-Ontong-Java event” related LIPs. Ingle et al. (2007) propose a magma formation characterized by extensive melting of depleted mantle wedge material and mixed with volcaniclastic sediment, possibly originated from the High Plateau. How this mixing of mantle material can be achieved is still under discussion. Ingle et al. (2007) propose a setting similar to a subduction zone, which cannot be supported by our interpretation of the Danger Islands Troughs. Golowin et al. (2014) however, postulate formation of the lavas showing U-shaped incompatible element patterns in an intraplate setting by second stage melting of FOZO type mantle at high temperatures and mixing of smaller amounts of an enriched HIMU component. On the Western Plateaus, faults can be observed up to the acoustic basement (Fig. 6), which leads to the conclusion that crustal stretching was still present during late secondary volcanism along with further sedimentation.

After the last main magmatic pulse ceased, further pelagic sedimentation covered most of the volcanic relicts and the faults and ridges generated by tectonic motion. The Manihiki Plateau subsided to the current water-depth of 2600 m for the High Plateau and 4000-3600 m for the Western Plateaus (Fig.17). The atolls and islands on the High Plateau and the Western Plateaus, which are currently above sea level result from smaller magmatic pulses younger than 45 Ma (Pietsch and Uenzelmann-Neben, submitted).

In summary, the Manihiki Plateau experienced multiple stages of magmatic emplacement and tectonic deformation. The Western Plateaus was mainly a subject to the tectonic forces related to the development of the Ellice Basin and the resulting stretching of the LIP's crust. The Western Plateaus had limited magma supply initiating the secondary magmatic stages, which are present on the High Plateau, but also on the proposed conjugate Ontong Java Plateau and Hikurangi Plateau (Hoernle et al., 2004; 2010; Mahoney and Spencer, 1991; Mahoney et al., 1993). The Danger Islands Troughs form the prominent dissection between those two sub-provinces and their different magmatic and tectonic evolution. Our data indicate the lack of upper crust, and lower crustal material crops out at the base of the highly sedimented trough..

The interplay between tectonic and magmatic overprint proves to be a crucial key aspect in understanding the crustal nature of the Manihiki Plateau and its role in the plate tectonic framework of the central Pacific.

Figure 17

7. Conclusions

We present the first seismic refraction/wide-angle reflection models of the two main sub-provinces – the Western Plateaus and the High Plateau – of the Manihiki Plateau. From this newly gained information on the crustal structure of the sub-provinces, the tectonic and magmatic evolution of this oceanic LIP can be further illuminated.

The Western Plateaus have a crustal thickness between 9.3 km in highly stretched LIP crust in the west and 17.2 km in the east. The upper crust does not only consist of mafic extrusives as expected for an oceanic LIP. The upper crust consists of highly compacted sediments in basins created by deep reaching faults. By comparison with other oceanic LIPs, it becomes obvious that the Western Plateaus did not experience extensive secondary magmatic stages.

The High Plateau shows the typical crustal structure of an oceanic LIP with a crustal thickness of up to 20 km. In the upper crust former eruptive centers of a secondary magmatic stage are visible. The major magmatic pathways of the LIP are traceable in the middle and lower crust.

The Danger Islands Troughs is the distinct border between the magmatically active High Plateau and the Western Plateaus, which did not experience major secondary magmatic events.

Our models indicate that the Manihiki Plateau experienced a different tectonic and magmatic evolution after its emplacement. Whereas the High Plateau underwent high volume secondary magmatism, comparable to the Ontong Java Plateau and the Hikurangi Plateau, and limited tectonic forces at its margins, the Western Plateaus experienced massive crustal stretching and only low volume secondary magmatism. These new insights in the formation and evolution of the Manihiki Plateau infer that the different sub-provinces of the Manihiki Plateau might allow examining the "Taylor-Hypothesis" (Taylor, 2006) of the "Super-LIP" Ontong Java Nui further.

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References

Ai, H.-A., Stock, J. M., Clayton, R. and Luyendyk, B.: Vertical tectonics of the High Plateau region, Manihiki Plateau, Western Pacific, from seismic stratigraphy, Marine Geophysical Research, 29(1), 13–26, doi:10.1007/s11001-008-9042-0, 2008.

Barton, P. J.: The relationship between seismic velocity and density in the continental crust — a useful constraint? Geophysical Journal International, 1986.

Beiersdorf, H., Bach, W., Duncan, R. A., Erzinger, J. and Weiss, W.: New evidence for the production of EM-type ocean island basalts and large volumes of volcaniclastites during the early history of the Manihiki Plateau, Marine Geology, 122(3), 181–205, doi:10.1016/0025-3227(94)00107-V, 1995a.

Beiersdorf, H., Bickert, T., Cepek, P., Fenner, J., Petersen, N., Schönfeld, J., Weiss, W. and Won, M. Z.: High-resolution stratigraphy and the response of biota to Late Cenozoic environmental changes in the central equatorial Pacific Ocean (Manihiki Plateau), Marine Geology, 125(1), 29–59, doi:10.1016/0025-3227(95)00021-P, 1995b.

Billen, M. I. and Stock, J.: Morphology and origin of the Osbourn Trough, Journal of Geophysical Research, 105(B6), 13481–13489, doi:10.1029/2000JB900035, 2000.

Bosence, D.: A genetic classification of carbonate platforms based on their basinal and tectonic settings in the Cenozoic, Sedimentary Geology, 175(1-4), 49–72, doi:10.1016/j.sedgeo.2004.12.030, 2005.

Bryan, S. E. and Ernst, R. E.: Revised definition of large igneous provinces (LIPs), Earth-Science Reviews, 86(1-4), 175–202, doi:10.1016/j.earscirev.2007.08.008, 2008.

Bryan, S. E. and Ferrari, L.: Large igneous provinces and silicic large igneous provinces: Progress in our understanding over the last 25 years, Geological Society of America Bulletin, 125(7-8), 1053–1078, doi:10.1130/B30820.1, 2013.

Chandler, M. T., Wessel, P. and Sager, W. W.: Analysis of Ontong Java Plateau palaeolatitudes: evidence for large-scale rotation since 123 Ma? Geophysical Journal International, 194(1), 18–29, doi:10.1093/gji/ggt075, 2013.

Chandler, M. T., Wessel, P., Taylor, B., Seton, M., Kim, S.-S. and Hyeong, K.: Reconstructing Ontong Java Nui: Implications for Pacific absolute plate motion, hotspot drift and true polar wander, Earth and Planetary Science Letters, 331-332, 140–151, doi:10.1016/j.epsl.2012.03.017, 2012.

Charvis, P. and Operto, S.: Structure of the Cretaceous Kerguelen Volcanic Province (southern Indian Ocean) from wide-angle seismic data, Journal of Geodynamics, 28(1), 51–71, doi:10.1016/S0264-3707(98)00029-5, 1999.

Christensen, N. I.: Poisson's ratio and crustal seismology, Journal of Geophysical Research, 101(B2), 3139–3156, doi:10.1029/95JB03446, 1996.

Clague, D. A.: Petrology of basaltic and gabbroic rocks dredged from the Danger Island Troughs, Manihiki Plateau, Initial Reports of the Deep Sea Drilling Project, 33, 891–911, 1976.

Coffin, M. F. and Eldholm, O.: Scratching the surface: Estimating dimensions of large igneous provinces, Geology, 21(6), 515–518, doi:10.1130/0091-7613(1993)021<0515:STSEDO>2.3.CO;2, 1993.

Coffin, M. F. and Eldholm, O.: Large igneous provinces: crustal structure, dimensions, and external consequences, Reviews of Geophysics, 32, 1–36, doi:10.1029/93RG02508, 1994.

Coffin, M. F., Duncan, R. A., Eldholm, O., Fitton, J. G., Frey, F. A., Larsen, H. C., Mahoney, J. J., Saunders, A. D., Schlich, R. and Wallace, P. J.: Large igneous provinces and scientific ocean drilling: Status quo and a look ahead, Oceanography, 19(4), 150–160, doi:10.5670/oceanog.2006.13, 2006.

Coffin, M. F., Pringle, M. S., Duncan, R. A., Gladczenko, T. P., Storey, M., Müller, R. D. and Gahagan, L. M.: Kerguelen hotspot magma output since 130 Ma, Journal of Petrology, 43(7), 1121–1137, doi:10.1093/petrology/43.7.1121, 2002.

Coulbourn, W. T. and Hill, P. J.: A field of volcanoes on the Manihiki Plateau: Mud or lava? Marine Geology, 98(2-4), 367–388, doi:10.1016/0025-3227(91)90111-G, 1991.

Courtillot, V., Jaupart, C., Manighetti, I., Tapponnier, P. and Besse, J.: On causal links between flood basalts and continental breakup, Earth and Planetary Science Letters, 166(3), 177–195, doi:10.1016/S0012-821X(98)00282-9, 1999.

Davy, B. and Wood, R.: Gravity and magnetic modelling of the Hikurangi Plateau, Marine Geology, 118(1-2), 139–151, doi:10.1016/0025-3227(94)90117-1, 1994.
Davy, B., Hoernle, K. and Reinhard, W.: Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history, Geochemistry Geophysics Geosystems, 9, –, doi:10.1029/2007GC001855, 2008.
Downey, N. J., Stock, J. M., Clayton, R. and Cande, S. C.: History of the Cretaceous Osbourn spreading center, Journal of Geophysical Research, 112, B04102, 2007.
Francis, T. J. G. and Raitt, R. W.: Seismic refraction measurements in the southern Indian Ocean, Journal of Geophysical Research, 72(12), 3015–3041, doi:10.1029/JZ072i012p03015, 1967.
Fromm, T.: AWIForge > Projects > PRay > Home, aforge.awi.de [online] Available from: http://aforge.awi.de/gf/project/pray (Accessed 22 January 2014), 2012.
Gohl, K. and Uenzelmann-Neben, G.: The crustal role of the Agulhas Plateau, southwest Indian Ocean: evidence from seismic profiling, Geophysical Journal International, 144(3), 632–646, doi:10.1046/j.1365-246X.2001.01368.x, 2001.
Gohl, K., Uenzelmann-Neben, G. and Grobys, N.: Groth and Dispersal of a southeast African Large Igneous Province, South African Journal of Geology, 114(3-4), 379–386, doi:10.2113/gssajg.114.3-4.379, 2011.
Golowin, R., Hoernle, K., Portnyagin, M., Hauff, F., Gurenko, A., Garbe-Schönberg, D. and Werner, R.: Mantle in the Manihiki Plateau source with ultra-depleted incompatible element abundances but FOZO-like isotopic signature. 2014.
Götze, H. J., Schmidt, S., Lahmeyer, B., Goltz, G., Alvers, M., Klesper, C., Schneider, M. and Schulte, J.: IGMAS - Interactive Gravity and Magnetic Application System, geophysik.uni-kiel.de [online] Available from: http://www.geophysik.uni-kiel.de/~sabine/Sabine_IGMAS.html (Accessed 29 September 2014), 2002.
Grevemeyer, I., Weigel, W., Schüssler, S. and Avedik, F.: Crustal and upper mantle seismic structure and lithospheric flexure along the Society Island hotspot chain, Geophysical Journal International, 147(1), 123–140, doi:10.1046/j.0956-540x.2001.01521.x, 2001.
Hamilton, E. L.: Sound velocity–density relations in sea-floor sediments and rocks, The journal of the Acoustical Society of America, 63(2), 366–377, doi:10.1121/1.381747, 1978.
Hoernle, K., Hauff, F., Reinhard, W. and Mortimer, N.: New insights into the origin and evolution of the Hikurangi oceanic plateau, Eos Trans. AGU, 85(41), 401–408, doi:10.1029/2004EO410001, 2004.
Hoernle, K., Hauff, F., van den Bogaard, P., Reinhard, W., Mortimer, N., Geldmacher, J., Garbe-Schönberg, D. and Davy, B.: Age and geochemistry of volcanic rocks from the Hikurangi and Manihiki oceanic Plateaus, Geochimica et Cosmochimica Acta, 74, 1–24, doi:10.1016/j.gca.2010.09.030, 2010.
Hoernle, K., Timm, C., Hauff, F., Rüpke, L., Werner, R., van den Bogaard, P., Michael, P., Coffin, M. F., Mortimer, N. and Davy, B.: New Results for the Multi-stage Geochemical Evolution of the Manihiki and Hikurangi Plateaus, Eos Trans. AGU, 1–24 [online] Available from: http://abstractsearch.agu.org/meetings/2009/FM/sections/V/sessions/V51H/abstracts/V51H-03.html, 2009.
Hussong, D. M., Wipperman, L. K. and Kroenke, L. W.: The Crustal Structure of the Ontong Java and Manihiki Oceanic Plateaus, Journal of Geophysical Research, 84(B11), 6003–6010, doi:10.1029/JB084iB11p06003, 1979.
Ingle, S. P., Mahoney, J. J., Sato, H., Coffin, M. F., Kimura, J.-I., Hirano, N. and Nakanishi, M.: Depleted mantle wedge and sediment fingerprint in unusual basalts from the Manihiki Plateau, central Pacific Ocean, Geology, 35(7), 595, doi:10.1130/G23741A.1, 2007.
Inoue, H., Coffin, M. F., Nakamura, Y., Mochizuki, K. and Kroenke, L. W.: Intrabasement reflections of the Ontong Java Plateau: Implications for plateau construction, Geochemistry Geophysics Geosystems, 9(4), 1–19, doi:10.1029/2007GC001780, 2008.
Karlstrom, L. and Richards, M. A.: On the evolution of large ultramafic magma chambers and timescales for flood basalt eruptions, Journal of Geophysical Research, 116(B8), B08216, doi:10.1029/2010JB008159, 2011.

Korenaga, J.: Why did not the Ontong Java Plateau form subaerially? Earth and Planetary Science Letters, 234(3), 385–399, doi:10.1016/j.epsl.2005.03.011, 2005.
Larson, R. L.: Superplumes and ridge interactions between Ontong Java and Manihiki plateaus and the Nova-Canton trough, Geology, 25(9), 779–782, 1997.
Larson, R. L. and Erba, E.: Onset of the Mid-Cretaceous greenhouse in the Barremian-Aptian: Igneous events and the biological, sedimentary, and geochemical responses, Paleoceanography, 14(6), 663–678, doi:10.1029/1999PA900040, 1999.
Larson, R. L., Pockalny, R. A., Viso, R. F., Erba, E., Abrams, L. J., Luyendyk, B., Stock, J. M. and Clayton, R.: Mid-Cretaceous tectonic evolution of the Tongareva triple junction in the southwestern Pacific Basin, Geology, 30(1), 67–70, doi:10.1130/0091-7613(2002)030<0067:MCTEOT>2.0.CO;2, 2002.
Mahoney, J. J. and Spencer, K. J.: Isotopic Evidence for the Origin of the Manihiki and Ontong Java Oceanic Plateaus, Earth and Planetary Science Letters, 104, 196–210, doi:10.1016/0012-821X(91)90204-U, 1991.
Mahoney, J. J., Storey, M., Duncan, R. A., Spencer, K. J. and Pringle, M.: Geochemistry and geochronology of Leg 130 basement lavas: nature and origin of the Ontong Java Plateau, Proceedings of the Ocean Drilling Programm Scientific Results, 130, 3–22, 1993.
McNutt, M. K.: Another nail in the plume coffin? Science, 313(5792), 1394, doi:10.1126/science.1131298, 2006.
Miura, S., Suyehiro, K., Shinohara, M., Takahashi, N., Araki, E. and Taira, A.: Seismological structure and implications of collision between the Ontong Java Plateau and Solomon Island Arc from ocean bottom seismometer–airgun data, Tectonophysics, 389(3-4), 191–220, doi:10.1016/j.tecto.2003.09.029, 2004.
Mochizuki, K., Coffin, M. F., Eldholm, O. and Taira, A.: Massive Early Cretaceous volcanic activity in the Nauru Basin related to emplacement of the Ontong Java Plateau, Geochemistry Geophysics Geosystems, 6(10), 1–19, doi:10.1029/2004GC000867, 2005.
Morgan, W. J.: Convection Plumes in the Lower Mantle, Nature, 230(5288), 42–43, doi:10.1038/230042a0, 1971.
Musgrave, R. J.: Evidence for Late Eocene emplacement of the Malaita Terrane, Solomon Islands: Implications for an even larger Ontong Java Nui oceanic plateau, J Geophys Res-Sol Ea, 118(6), 2670–2686, 2013.
Parsiegla, N., Gohl, K. and Uenzelmann-Neben, G.: The Agulhas Plateau: Structure and evolution of a large igneous province, Geophysical Journal International, 174(1), 336–350, doi:10.1111/j.1365-246X.2008.03808.x, 2008.
Pietsch, R. and Uenzelmann-Neben, G.: The Manihiki Plateau - a multistage volcanic emplacement history seen in high resolution seismic reflection data, submitted to Earth and Planetary Science Letters
Richards, M. A., Contreras-Reyes, E., Lithgow-Bertelloni, C., Ghiorso, M. and Stixrude, L.: Petrological interpretation of deep crustal intrusive bodies beneath oceanic hotspot provinces, Geochemistry Geophysics Geosystems, 14, 604–619, doi:10.1029/2012GC004448, 2013.
Ridley, V. A. and Richards, M. A.: Deep crustal structure beneath large igneous provinces and the petrologic evolution of flood basalts, Geochemistry Geophysics Geosystems, 11(9), doi:10.1029/2009GC002935, 2010.
Sandwell, D. T. and Smith, W. H.: Marine gravity anomaly from Geosat and ERS 1 satellite altimetry, Journal of Geophysical Research, 102(B5), 10039–10–054, doi:10.1029/96JB03223, 1997.
Schlanger, S. O., Jackson, E., Boyce, R. E., Cook, H., Jenkyns, H. C., Johnson, D., Kaneps, A., Kelts, K., Martini, E., McNulty, C. and Winterer, E. L.: Inital Reports of the Deep Sea Drilling Project, 33, 1–354, 1976.
Tarduno, J. A.: Evidence for Extreme Climatic Warmth from Late Cretaceous Arctic Vertebrates, Science, 282(5397), 2241–2243, doi:10.1126/science.282.5397.2241, 1998.
Taylor, B.: The single largest oceanic plateau: Ontong Java-Manihiki-Hikurangi, Earth and Planetary Science Letters, 241, 372–380, doi:10.1016/j.epsl.2005.11.049, 2006.
Timm, C., Hoernle, K., Reinhard, W., Hauff, F., van den Bogaard, P., Michael, P., Coffin, M. F. and Koppers, A. A. P.: Age and geochemistry of the oceanic Manihiki Plateau, SW Pacific:

New evidence for a plume origin, Earth and Planetary Science Letters, 304, 135–146, doi:10.1016/j.epsl.2011.01.025, 2011.
Uenzelmann-Neben, G.: The expedition of the research vessel“ Sonne” to the Manihiki Plateau in 2012 (So 224), Reports of Polar and Marine Research, 2012.
Uenzelmann-Neben, G., Gohl, K., Ehrhardt, A. and Seargent, M.: Agulhas Plateau, SW Indian Ocean: New evidence for excessive volcanism, Geophysical Research Letters, 26(13), 1941–1944, doi:10.1029/1999GL900391, 1999.
Van Andel, T. H.: Mesozoic/Cenozoic calcite compensation depth and the global distribution of calcareous sediments, Earth and Planetary Science Letters, 26(2), 187–194, doi:10.1016/0012-821X(75)90086-2, 1975.
Viso, R. F., Larson, R. L. and Pockalny, R. A.: Tectonic evolution of the Pacific–Phoenix–Farallon triple junction in the South Pacific Ocean, Earth and Planetary Science Letters, 233(1-2), 179–194, doi:10.1016/j.epsl.2005.02.004, 2005.
Werner, R. and Hauff, F.: FS Sonne, Fahrtbericht/Cruise Report SO193 MANIHIKI: Temporal, Spatial, and Tectonic Evolution of Oceanic Plateaus; Suva/Fiji-Apia/Samoa, 19.05.- …,, 2007.
Werner, R., Nürnberg, D. and Hauff, F.: RV Sonne Fahrtbericht/Cruise Report SO225, GEOMAR Reports, 1–176, 2013.
White, R. S. and McKenzie, D.: Mantle plumes and flood basalts, Journal of Geophysical Research, 100(B9), 17543–17,585, doi:10.1029/95JB01585, 1995.
Wignall, P. B.: Large igneous provinces and mass extinctions, Earth-Science Reviews, 53(1-2), 1–33, doi:10.1016/S0012-8252(00)00037-4, 2001.
Winterer, E. L., Lonsdale, P. L., Matthews, J. L. and Rosendahl, B. R.: Structure and acoustic stratigraphy of the Manihiki Plateau, Deep Sea Research and Oceanographic Abstracts, 21(10), 793–813, doi:10.1029/JB084iB11p06003, 1974.
Worthington, T. J., Hekinian, R., Stoffers, P., Kuhn, T. and Hauff, F.: Osbourn Trough: Structure, geochemistry and implications of a mid-Cretaceous paleospreading ridge in the South Pacific, Earth and Planetary Science Letters, 245(3-4), 685–701, doi:10.1016/j.epsl.2006.03.018, 2006.
Zelt, B.: zp - Software for plotting & picking SEGY seismic refraction data, soest.hawaii.edu [online] Available from: http://www.soest.hawaii.edu/users/bzelt/zp/zp.html (Accessed 22 January 2014), 2004.
Zelt, C. A. and Smith, R. B.: Seismic traveltime inversion for 2-D crustal velocity structure, Geophysical Journal International, 108(1), 16–34, doi:10.1111/j.1365-246X.1992.tb00836.x/abstract, 1992.

Table

<table>
<tr><td>Phase</td><td>AWI-20120100</td><td>AWI-20120200</td></tr>
<tr><td>PuP</td><td>+0.4/-0.5</td><td>+0.5/-0.4</td></tr>
<tr><td>PumP</td><td rowspan="2">+0.7/-0.4</td><td>+0.6/-0.4</td></tr>
<tr><td>PlmP</td><td>+0.6/-0.7</td></tr>
<tr><td>PmP</td><td>+0.7/-0.7</td><td>+0.8/-0.5</td></tr>
</table>

Table 1:

Depth uncertainties within the intracrustal reflectors in km

Layer	AWI-20120100	AWI-20120200
sediments	2050	2050
low velocity upper crust	2400	not present
upper crust	2650	2800
upper middle crust	2800	2900
middle crust	2900	not present
lower middle crust	2950	2950
lower crust	3130	3150
mantle	3300	3300

Table 2:

Rock densities [kg/m^3] (Barton, 1986; Hamilton, 1978) used for specific layers in the gravity anomaly model

Figures:

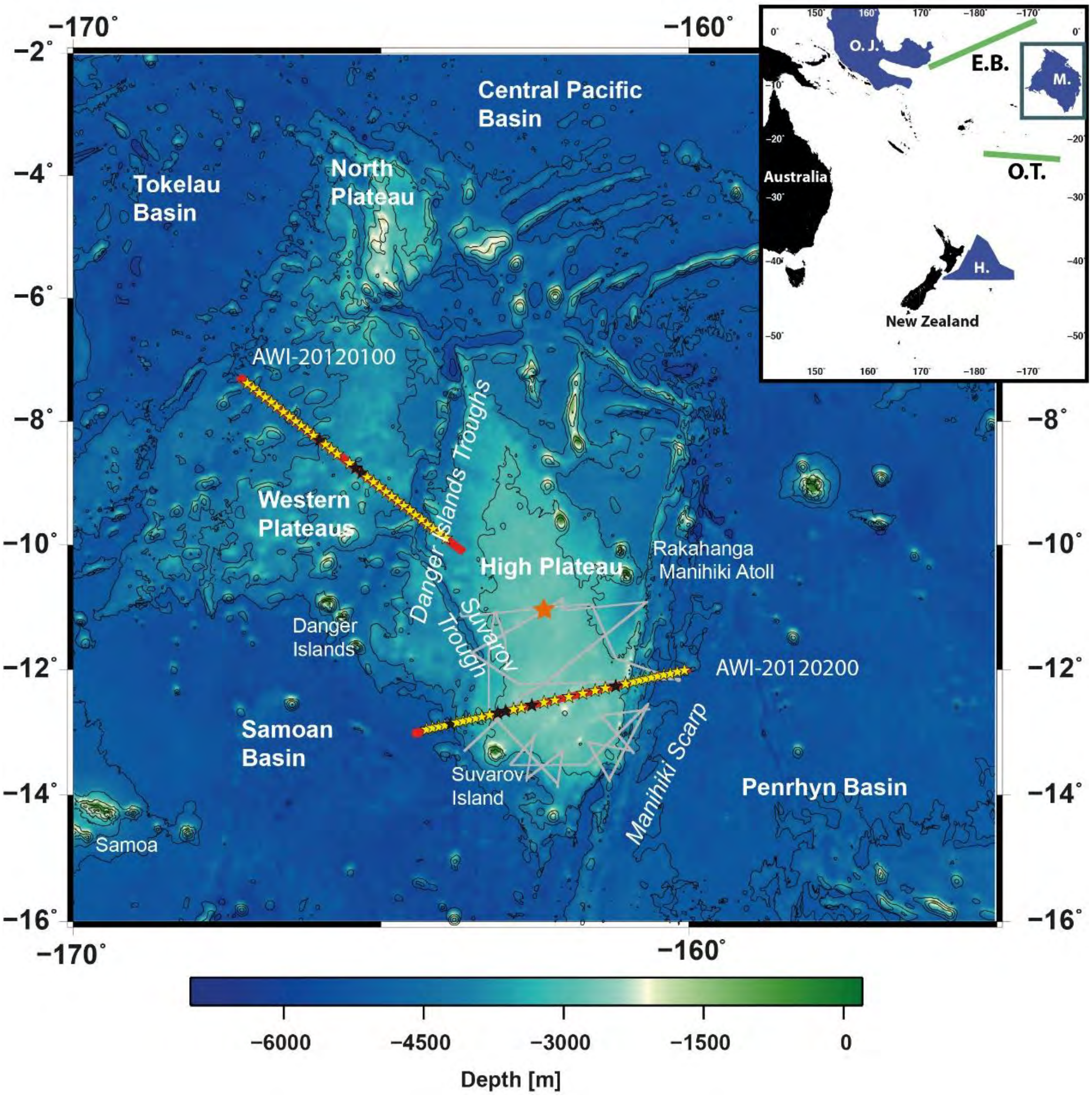

Figure. 1

Bathymetric map (30 arc-second grid GEBCO_2014) of the central western Pacific, refraction/wide-angle reflection seismic lines of cruise SO-224 are shown in red, reflection seismic lines of SO-224 are shown in grey. Yellow stars depict the positions of the OBS-stations.The black stars are stations, which did not return any data. The borehole DSDP Leg 31 Site 317 is marked by the orange star. In the inlet map the three main components of the Ontong Java Nui related LIPs are shown at their current position (Ontong Java (O. J.), Hikurangi (H.) and Manihiki (M.)). The green lines indicate the former spreading centers of the Osbourn Trough (O.T.) and the Ellice Basin (E.B.). The dark green box indicates the study area shown in the larger map.

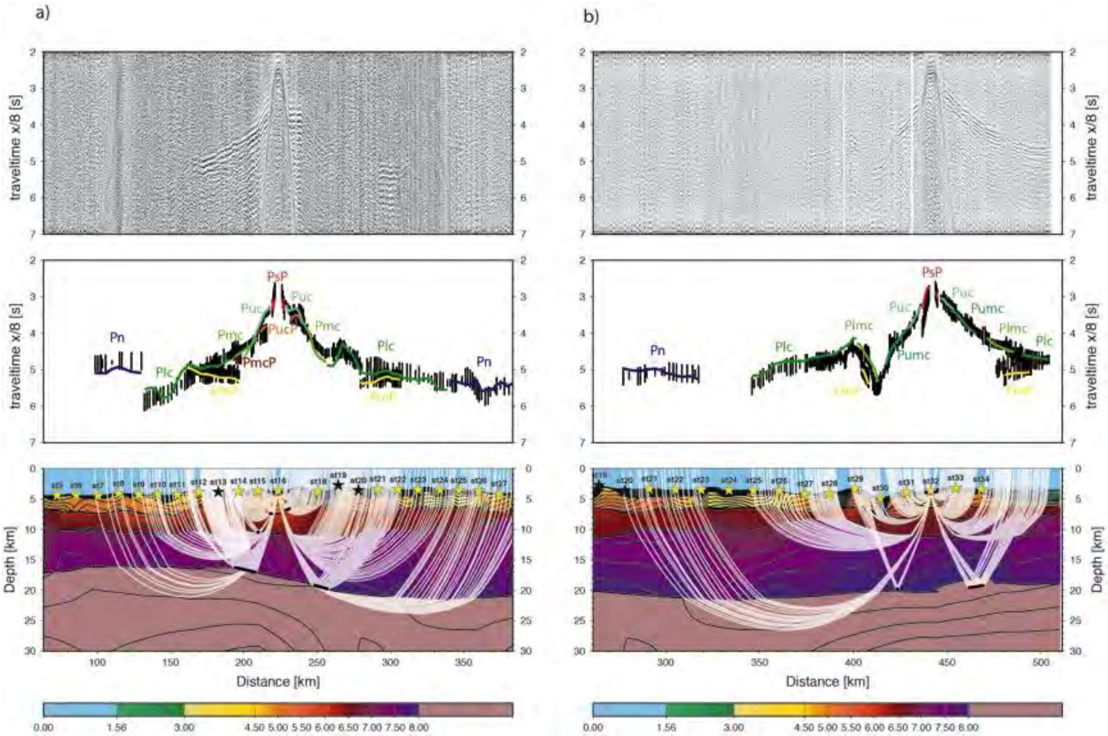

Figure 2:

Data examples for P-wave arrivals (AWI-20120100) at a) st16 representing the central Western Plateaus and b) st32 representing the Danger Islands Troughs area

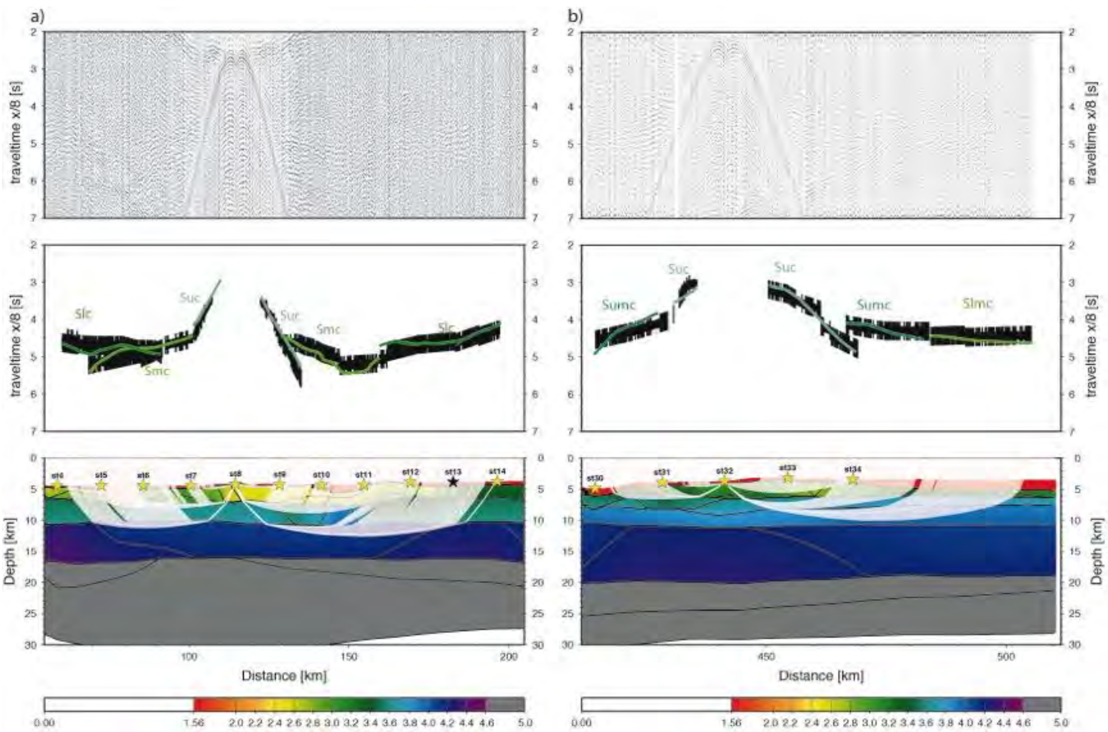

Figure 3:

Data examples for S-wave arrivals (AWI-20120100) at a) st08 representing the western part of the Western Plateaus and b) st32 representing the Danger Islands Troughs area

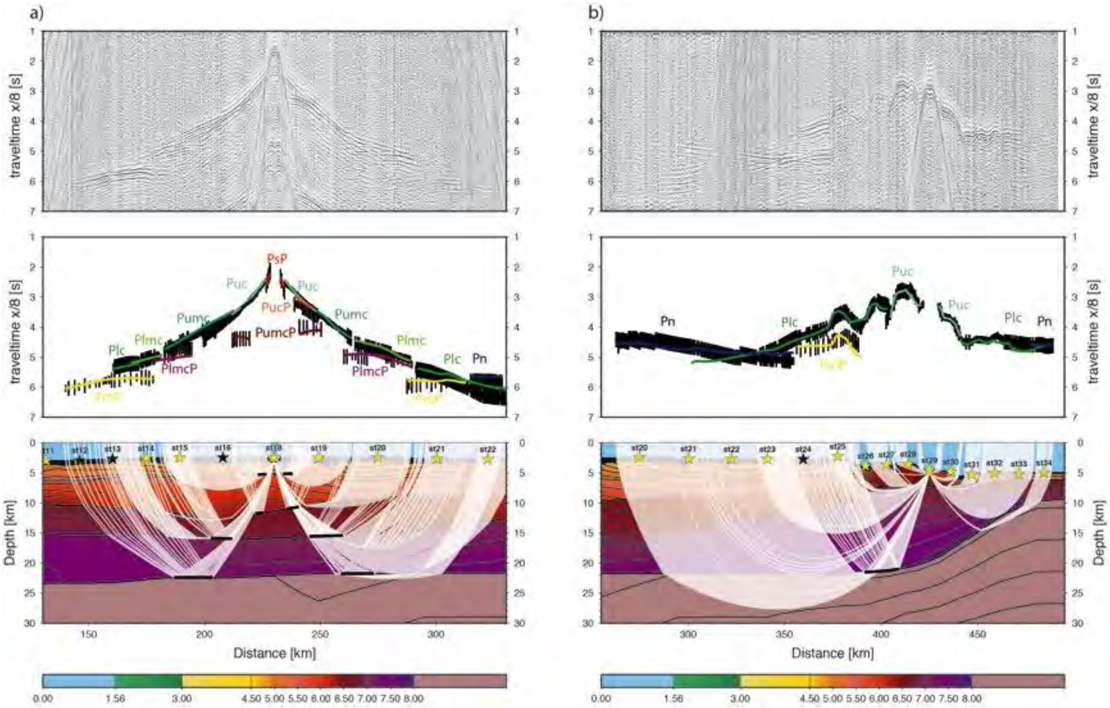

Figure 4:

Data examples for P-wave arrivals (AWI-20120200) at a) st18 representing the central High Plateau and b) st29 representing the Manihiki Scarp area

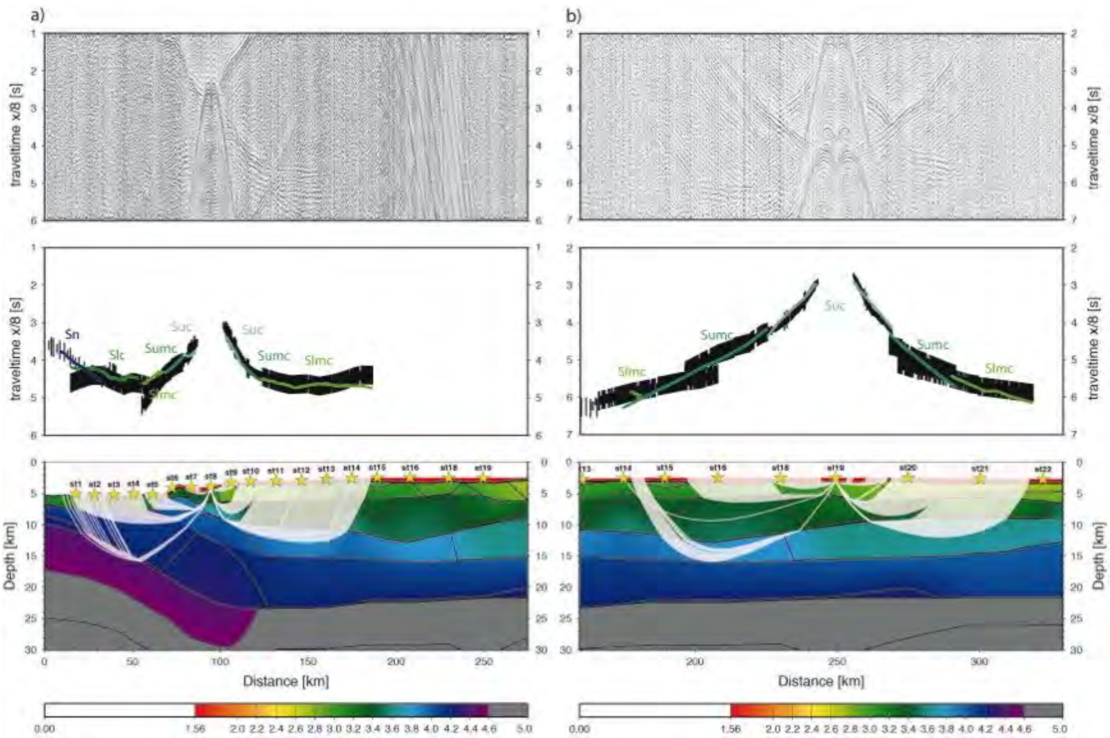

Figure 5:

Data examples for S-wave arrivals (AWI-20120200) at a) st08 representing the western margin of the High Plateau and b) st19 representing the central High Plateau

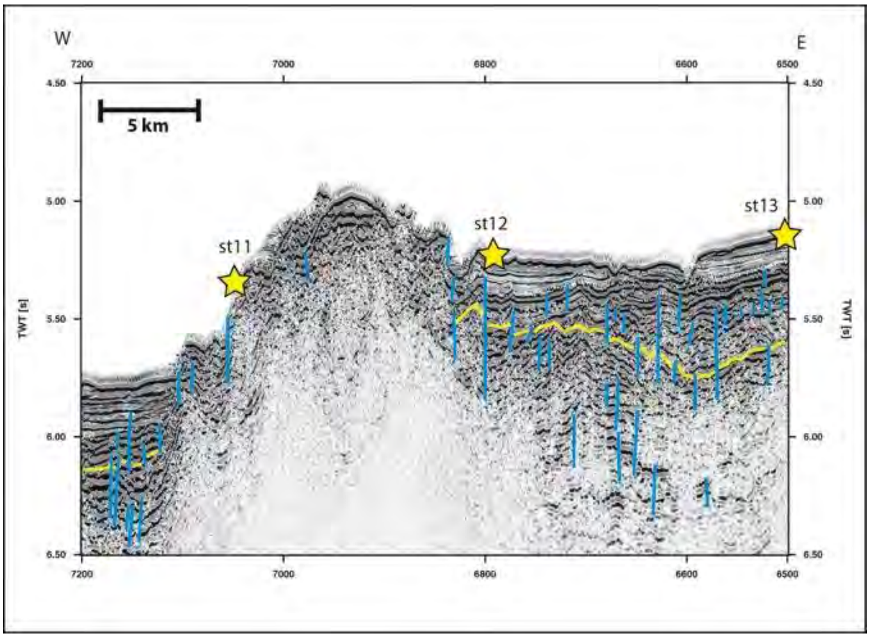

Figure 6:

Reflection seismic data from AWI-20120100 between OBS-station 11 and OBS-station 13, the blue lines indicate faultsystems, the yellow reflector indicates the top of basalt. The faults were identified in an enlarged section.

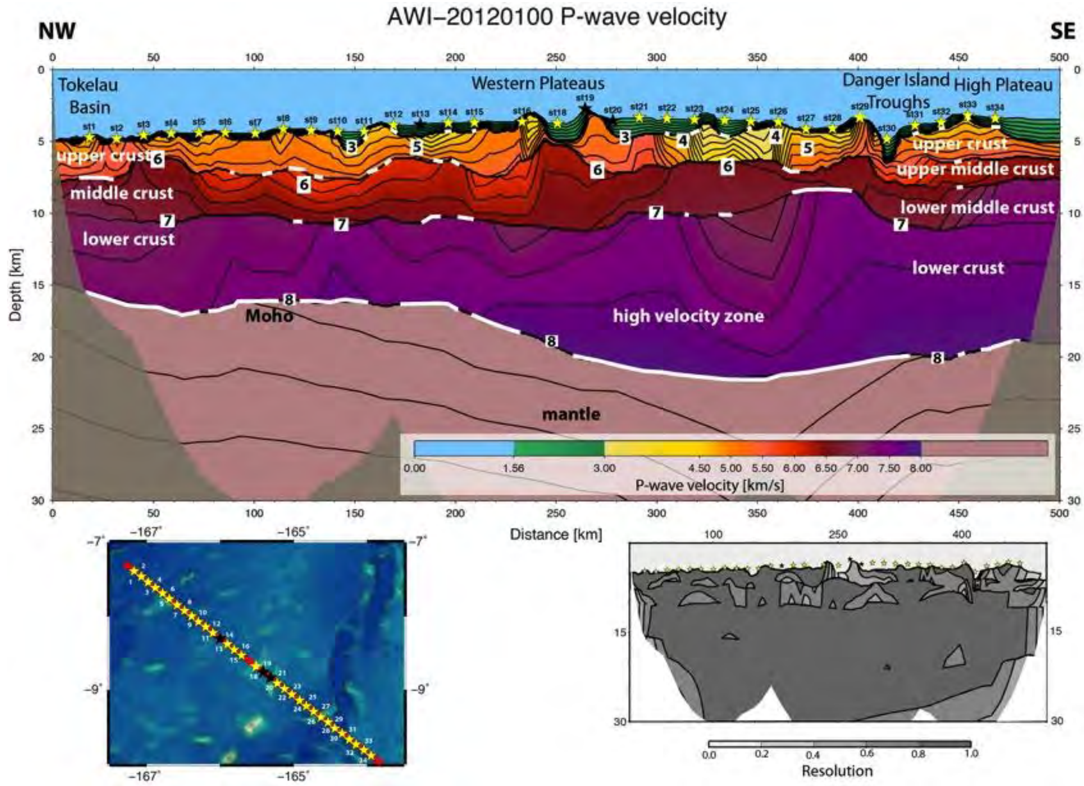

Figure 7:

P-wave velocity model of AWI-20120100; the grey transparent areas are not covered by rays. White layer boundaries are constraint by wide-angle reflections. The location of OBS stations is indicated with the yellow (functioning station) and black (malfunctioning station) stars. The resolution calculated for the velocity nodes is shown in the lower right corner. The RMS-value of this model is 0.189 and the χ^2 is 1.552. The vertical exaggeration is 7:1.

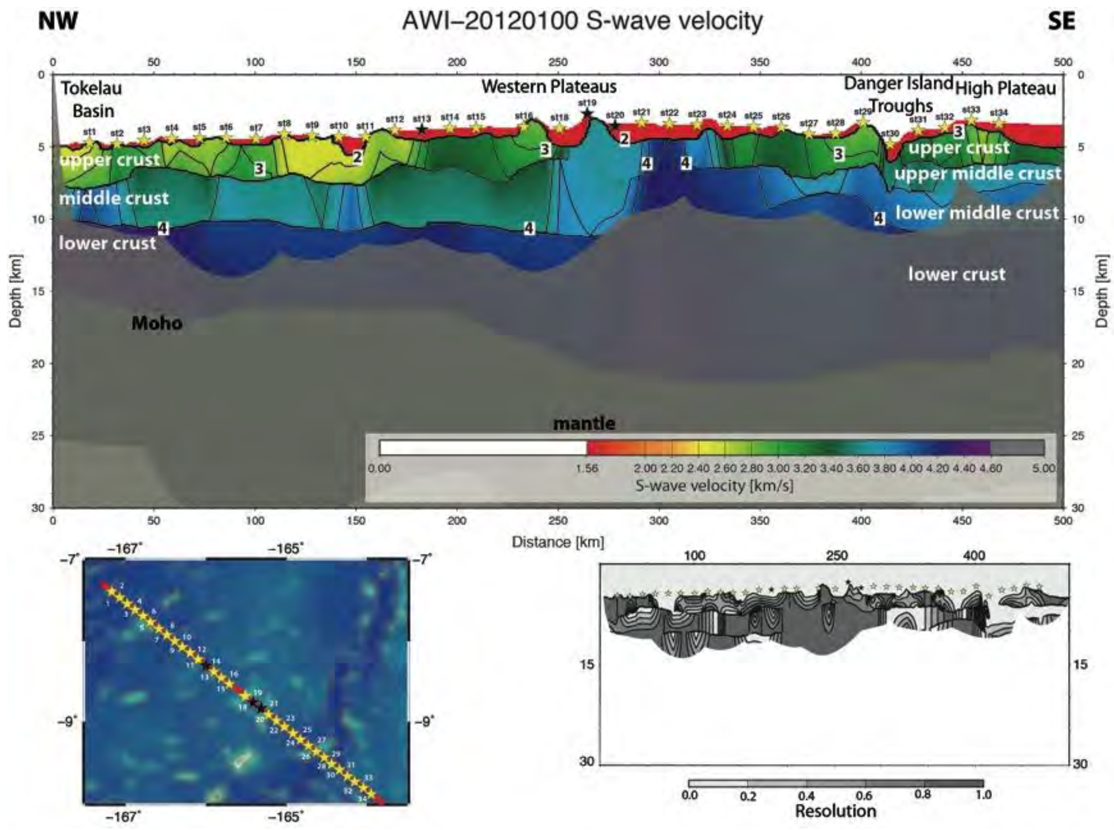

Figure 8:

S-wave velocity model of AWI-20120100; the grey transparent areas are not covered by rays. The locations of OBS stations are indicated with the yellow (functioning station) and black (malfunctioning station) stars. The resolution calculated for the velocity nodes is shown in the lower right corner. The RMS-value of this model is 0.280 and the χ^2 is 2.751. The vertical exaggeration is 7:1.

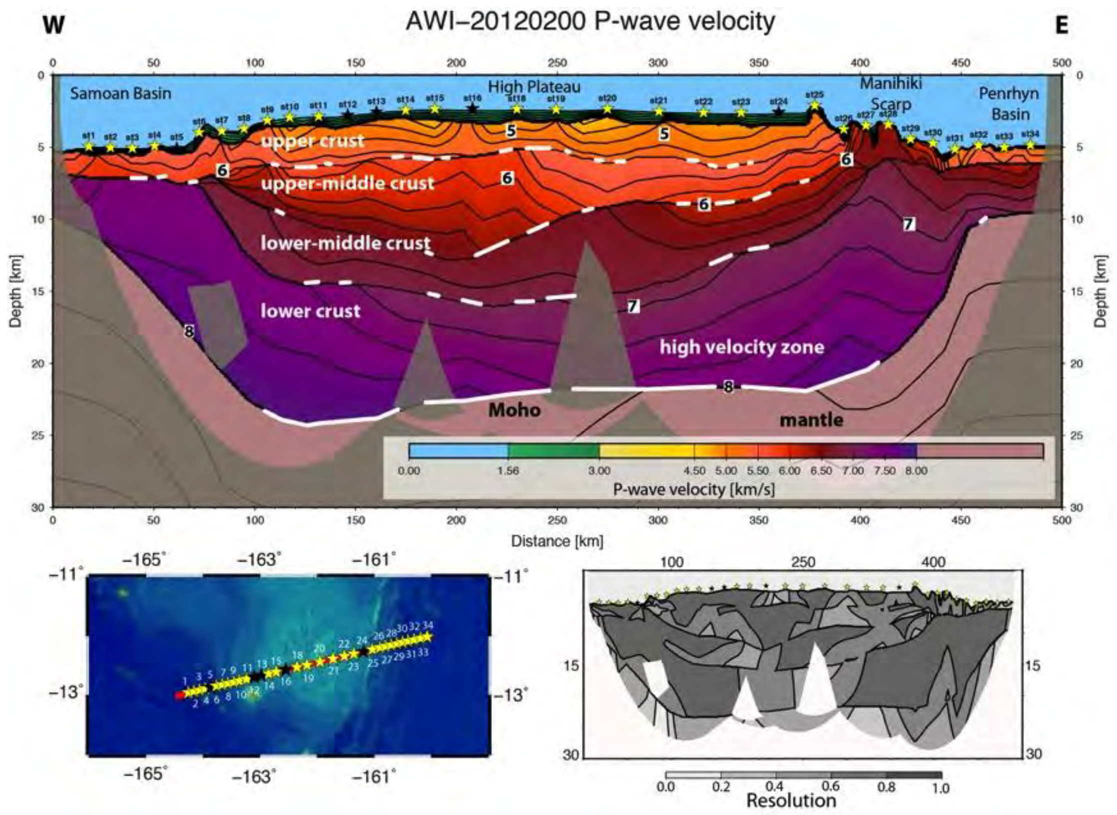

Figure 9:

P-wave velocity model of AWI-20120200; the grey transparent areas are not covered by rays. White layer boundaries are constrained by wide-angle reflections. The locations of OBS stations are indicated with the yellow (functioning station) and black (malfunctioning station) stars. The resolution calculated for the velocity nodes is shown in the lower right corner. The RMS-value of this model is 0.186 and the χ^2 is 1.674. The vertical exaggeration is 7:1.

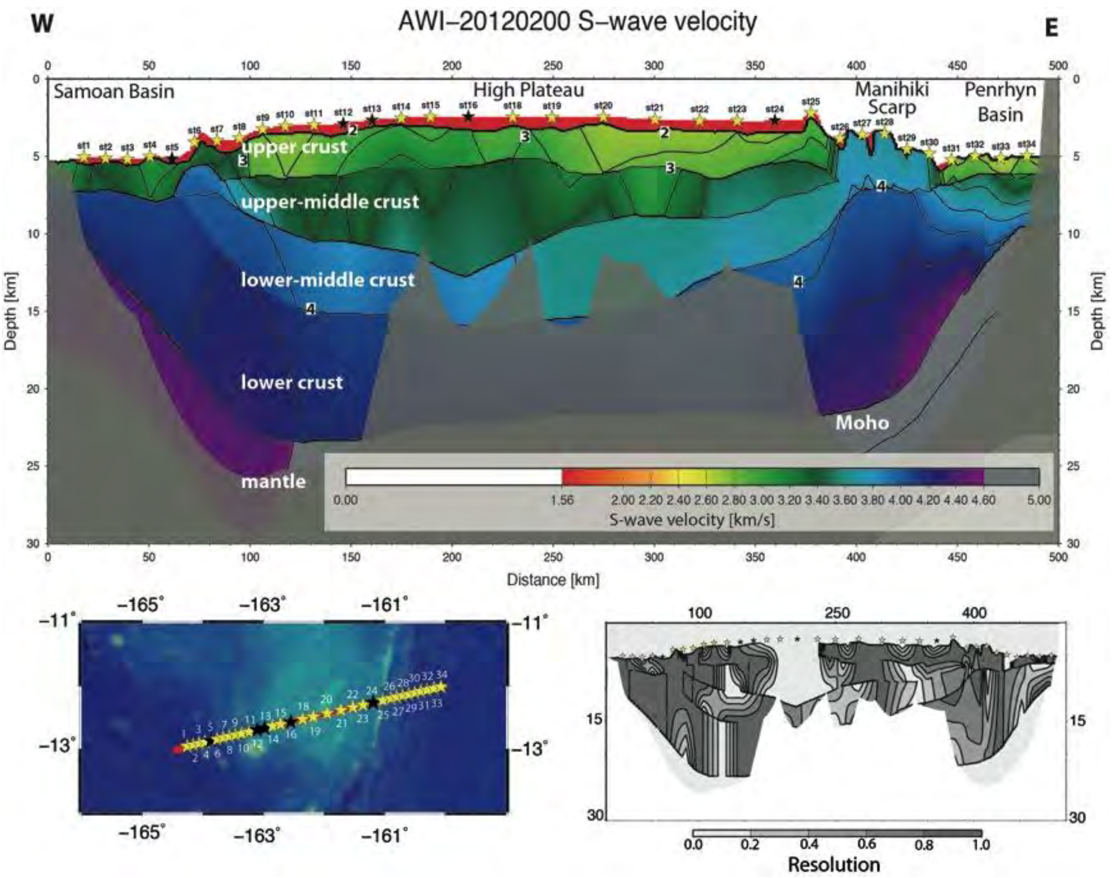

Figure 10:

S-wave velocity model of AWI-20120200; the grey transparent areas are not covered by rays. The locations of OBS stations are indicated with the yellow (functioning station) and black (malfunctioning station) stars. The resolution calculated for the velocity nodes is shown in the lower right corner. The RMS-value of this model is 0.223 and the χ^2 is 2.11. The vertical exaggeration is 7:1.

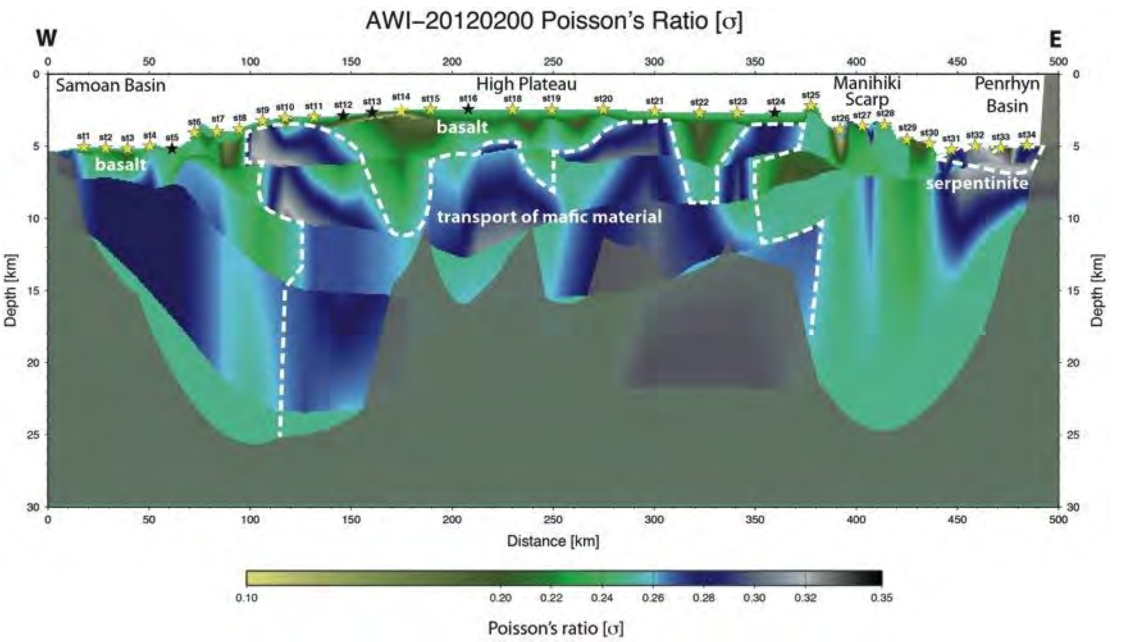

Figure 11:

Calculated Poisson's ratio for AWI-20120200; grey shaded areas are not covered by rays in the S-wave model and therefore cannot return reliable values for the Poisson's ratio. White dashed lines indicate areas of higher Poisson's ratio. The vertical exaggeration is 7:1.

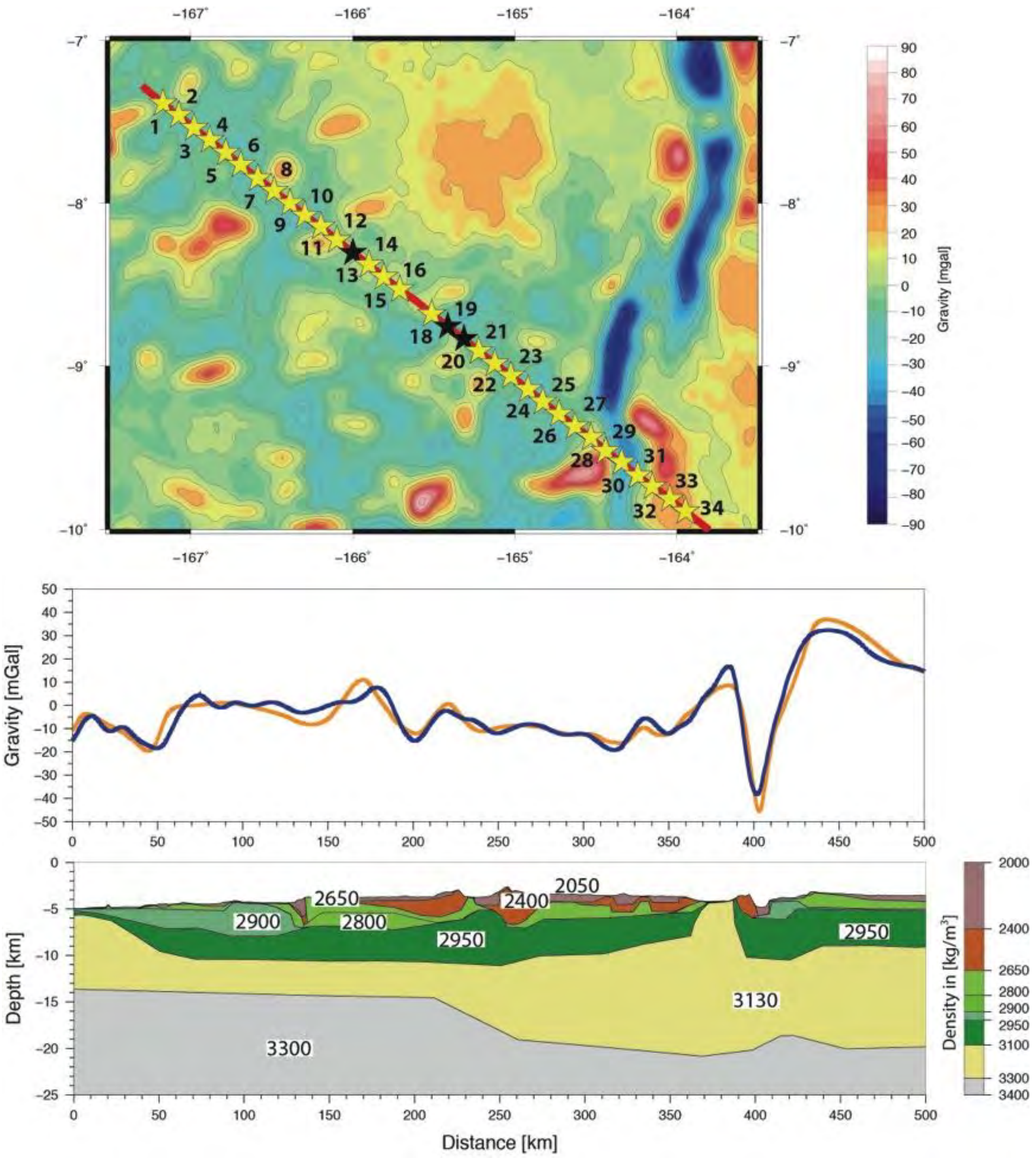

Figure 12:

Gravity model of AWI-20120100 upper panel: free-air gravity anomaly (Sandwell and Smith, 1997 v.23) middle panel: measured anomaly in blue calculated anomaly in orange, lower panel: gravity model for AWI-20120100

The mean residual value for this model is 3.6 mgal.

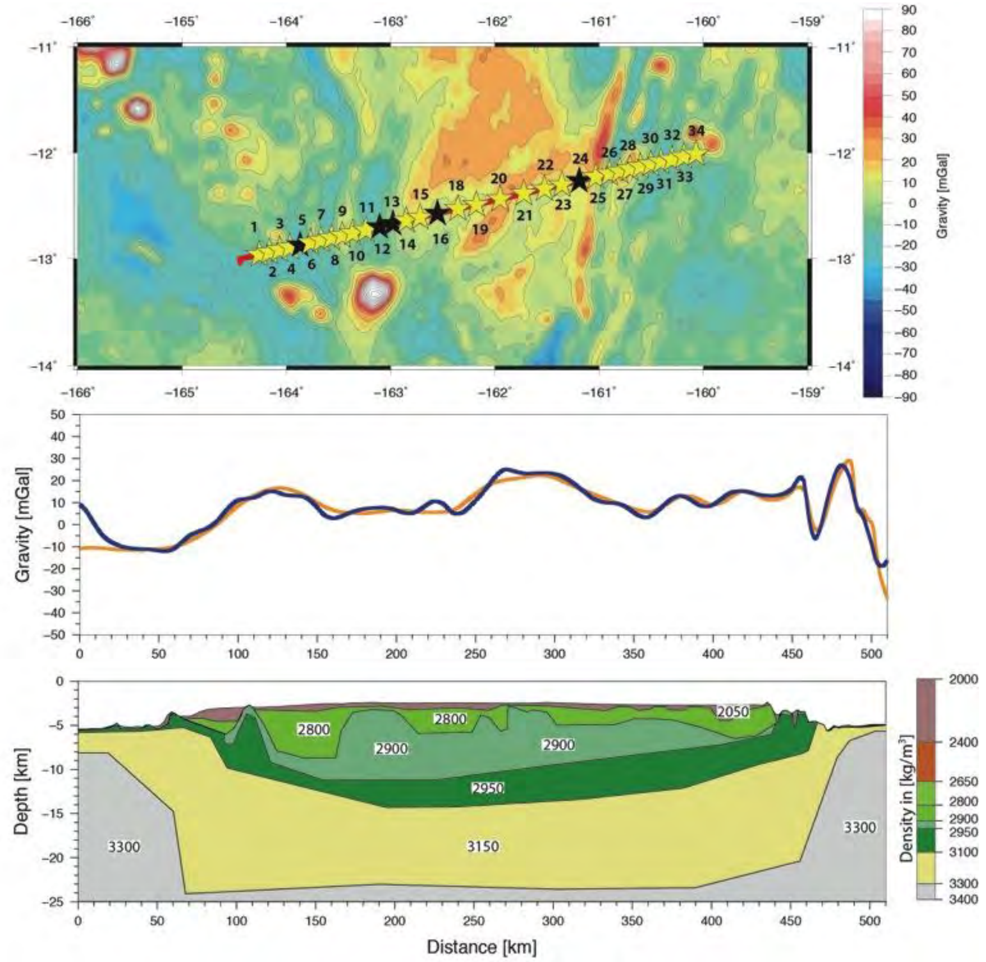

Figure 13:

Gravity model of AWI-20120200 upper panel: free-air gravity anomaly (Sandwell and Smith, 1997 v.23) middle panel: measured anomaly in blue calculated anomaly in orange, lower panel: gravity model for AWI-20120200.

The mean residual value for this model is 5.8 mgal (mean residual value of 2.6 mgal between 40 and 490 km).

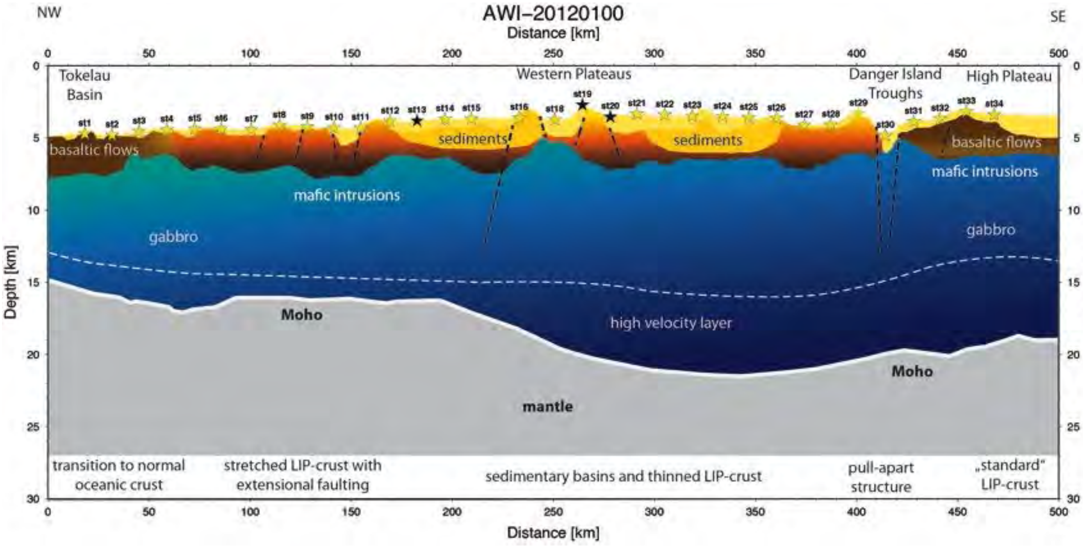

Figure 14:

Sketch of geological interpretation of AWI-20120100, black dashed lines indicate fault systems, the white dashed lines indicate the presence of the HVZ. The vertical exaggeration is 7:1.

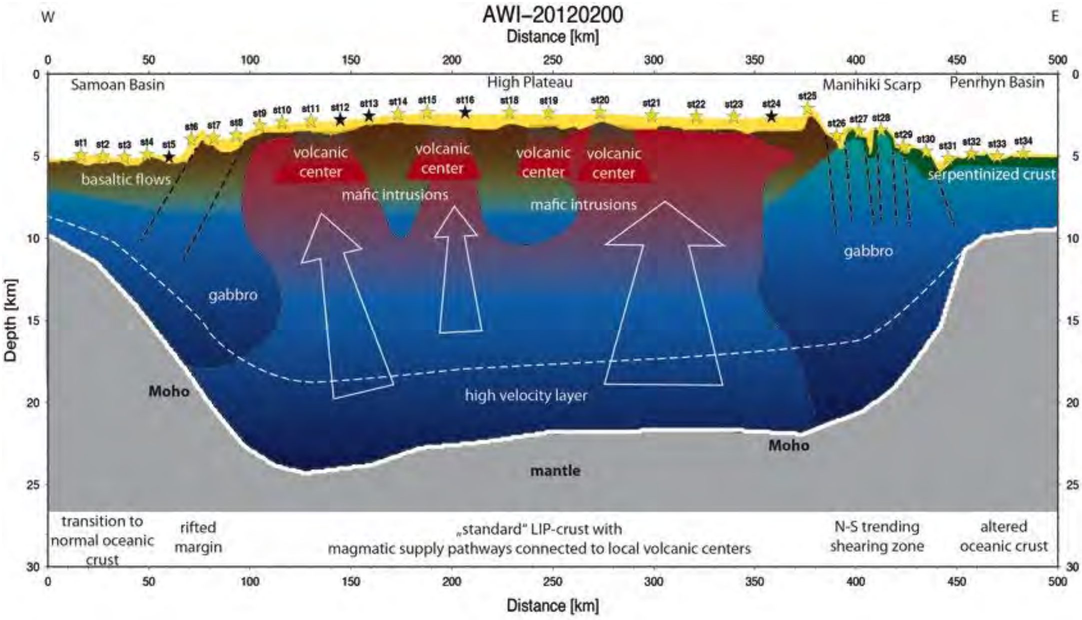

Figure 15:

Sketch of geological interpretation of AWI-20120200, white arrows indicate areas of magmatic upwelling, black dashed lines indicate fault systems, dashed white lines indicate the presence of the HVZ. The vertical exaggeration is 7:1.

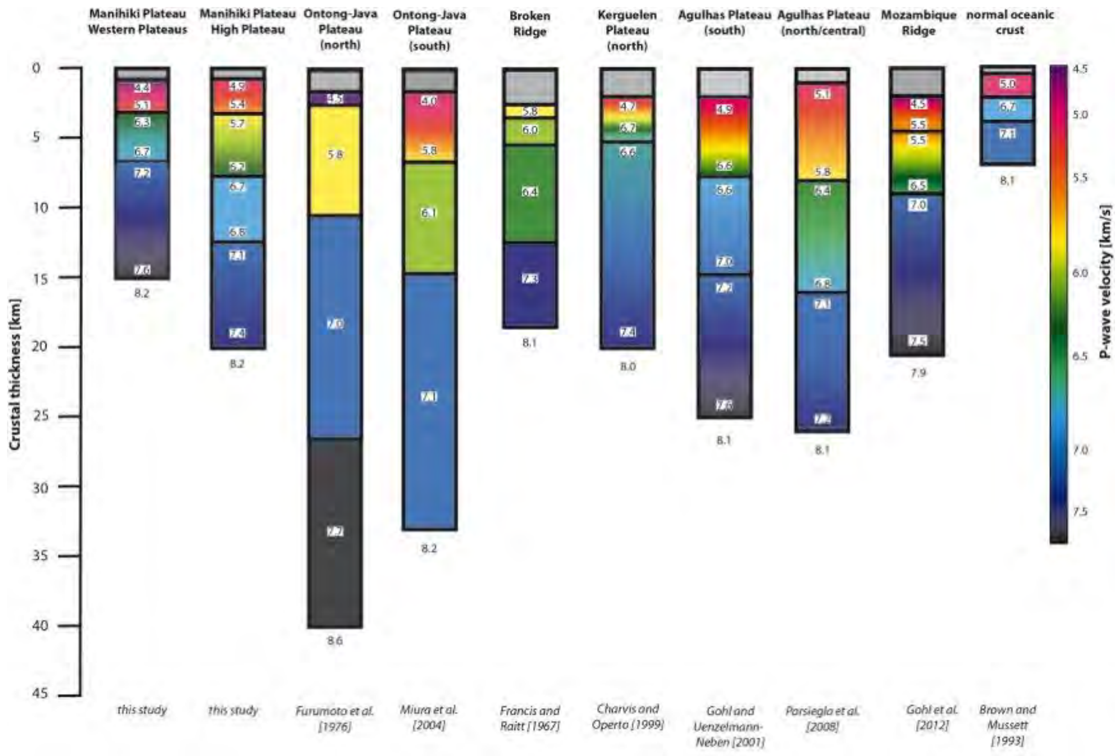

Figure 16:

Comparison of the crustal structure of the Western Plateaus and the High Plateau with other oceanic LIPs; The grey shaded areas represent the sedimentary cover. The high velocity zone of the lower crust is represented in the blue colors. The transitional crustal layer is depicted in the yellow and orange colors.

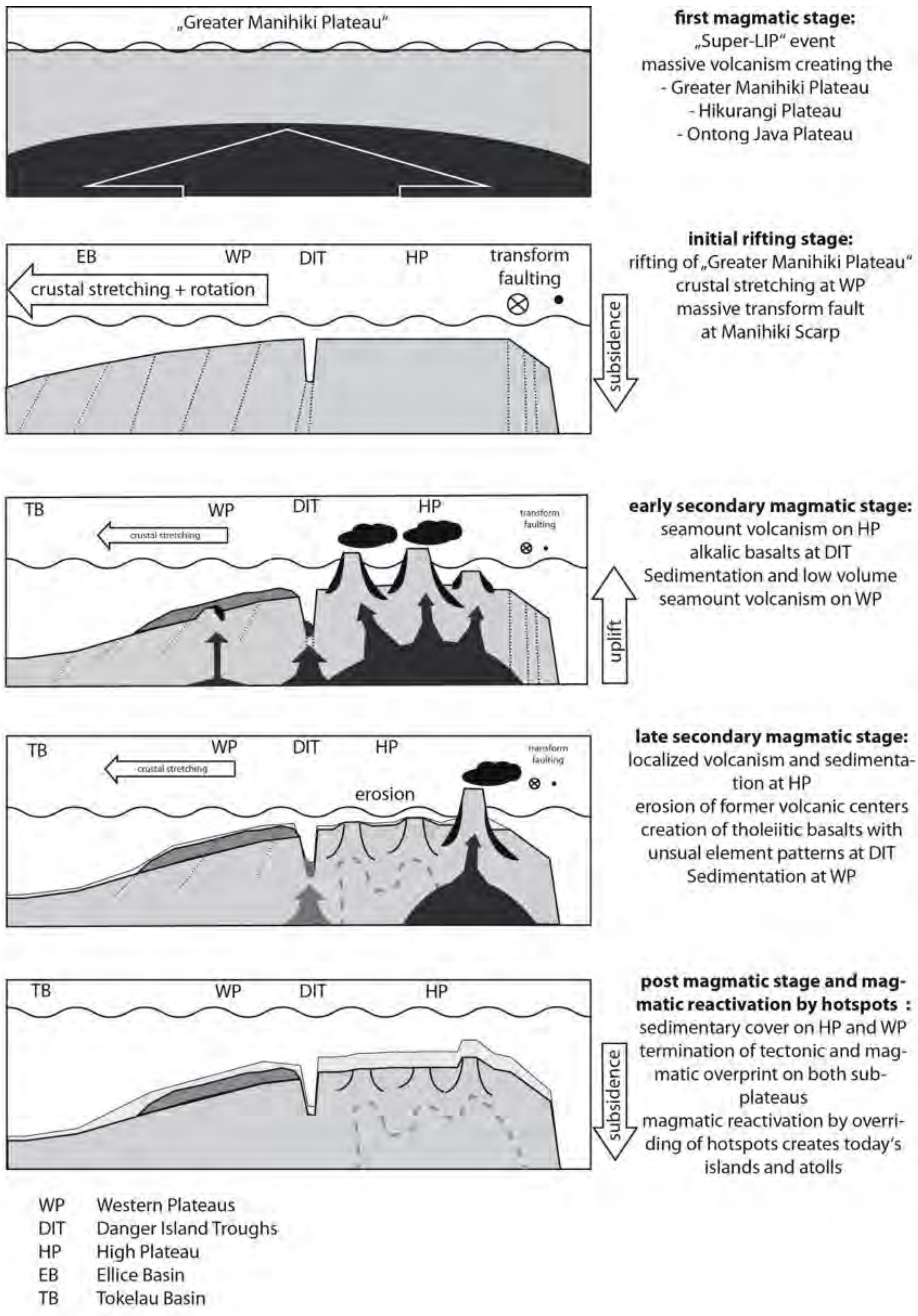

Figure 17:

Sketch of the evolution of the Manihiki Plateau with special focus on the High Plateau and the Western Plateaus from the early Cretaceous to the current setting

Hochmuth, K., Gohl, K., Uenzelmann-Neben, G. (submitted): Playing jigsaw with large igneous provinces – a plate tectonic reconstruction of Ontong Java Nui. Solid Earth. - Anlage 3

Playing jigsaw with large igneous provinces – a plate tectonic reconstruction of Ontong Java Nui

Katharina Hochmuth, Karsten Gohl, Gabriele Uenzelmann-Neben
Alfred-Wegener Institut Helmholtz-Zentrum für Polar- und Meeresforschung
Am Alten Hafen 26, 27568 Bremerhaven

Abstract
The plate tectonic framework of the western Pacific during the Cretaceous is a complicated interplay between short-lived spreading centers, microplates and continental and oceanic plateau fragments. We investigate the Large Igneous Provinces of the area to infer possible indications for tectonic mechanisms. The Large Igneous Provinces of the western Pacific might have developed as a single "Super" Large Igneous Province Ontong Java Nui and have a great impact on the plate kinematic behavior of the tectonic plates of the western Pacific. By the examination of seismic reflection data and refraction/wide-angle reflection seismic data, we reconstruct break-up mechanisms at all the margins of the Manihiki Plateau, the proposed centerpiece of the Ontong Java Nui. In our plate tectonic model, we predict an interaction of the arriving plume head with the spreading center, separating the Pacific and the Phoenix Plate during the initial emplacement of the Large Igneous Province. By accounting for crustal stretching and secondary magmatic activity, it is possible to decipher the complicated interplay between the emplacement of a Large Igneous Province and the tectonic forces such as extensional and shearing forces. This reconstruction sheds light on a timeframe, which is poorly understood but seems to be crucial for understanding global aspects such as the cessation of subduction at the western Gondwana margin.

1. Introduction

The plate tectonic set-up of the central and western Pacific, since the Cretaceous is a mosaic of multiple small and short-lived oceanic plates and continental fragments. Plate kinematic reconstructions [e.g by *Davy et al.*, 2008; *Chandler et al.*, 2012; *Seton et al.*, 2012] struggle to explain all the features of the difficult interplay between Large Igneous Provinces (LIP), relict spreading centers, subduction and hot spot volcanism overprinting the area. Since the creation of most of the oceanic crust of the Western Pacific takes place during the Cretaceous Normal Superchron, there are no seafloor-spreading anomalies to guide plate tectonic reconstructions (Fig.1). The remnants of the proposed "Super"-LIP emplacement during the early Cretaceous [*Taylor*, 2006; *Chandler et al.*, 2012; 2013] play an important role in this set-up, since former parts of this proposed Super-"LIP" - the Ontong Java Plateau and the Hikurangi Plateau - interact both with subduction trenches bordering the Australian Plate (Fig. 1) and were possibly individual oceanic plates during the Cretaceous. The termination of the subduction at the eastern Gondwana Margin can also be linked to the arrival of the Hikurangi Plateau at the subduction zone [*Davy and Wood*, 1994; *Billen and Stock*, 2000; *Davy et al.*, 2008; 2012; *Reyners*, 2013; *Davy*, 2014; *Timm et al.*, 2014], a process which initiated a global plate reorganization event [*Matthews et al.*, 2012]. The third major LIP, the Manihiki Plateau has currently no direct interaction with tectonically active plate boundaries, but indicates at its margins as well as in its internal fragmentation the tectonic forces experienced during the Cretaceous. The internal deformation of the Manihiki Plateau into three sub-provinces has previously been ignored by all published plate tectonic reconstructions. Recently published findings revealed distinct differences in the tectonic and magmatic evolution between the main two sub-provinces the Western Plateaus and the High Plateau.
In this paper, we analyze the role of LIPs in the plate tectonic framework of the western Pacific Ocean and revisit the hypothesis of the coupled emplacement of the major LIPs of the western

Pacific as proposed by Taylor (2006) and Chandler et al. (2012). By re-examining available refraction/wide-angle reflection seismic data along with seismic reflection data and global gravity and bathymetry grids, we present a more detailed reconstruction of the emplacement of the Pacific's LIPs, possible break-up scenarios and the role of internal fragmentation of the Manihiki Plateau.

Figure 1 current plate tectonic set-up of the western Pacific Ocean; Active subduction zones are shown in orange, transform faults in red and mid-ocean ridges in black. Black dashed lines are tectonic lineations tracked from magnetic anomaly maps [Maus et al., 2009] and gravity anomaly maps [Sandwell and Smith, 1997]. Green and turquois lines indicate fault zones within Jurassic seafloor [Nakanishi et al., 1992] and Cretaceous seafloor respectively. The former spreading center at the Osbourn Trough is marked in blue and the Tongareva Triple Junction Trace is marked in magenta. Isochrons (thin black lines) are taken from Seton et al. [2012] and are shaded in orange for the M-Series and in yellow for the C-Series on the Pacific Plate. Seafloor emplaced during the Cretaceous Normal Superchron is shown in white.

2. Geological Setting

2.1 The large igneous provinces of the western Pacific

The crustal structure and therefore the plate tectonic behavior of LIPs differ greatly from normal oceanic crust. Although the igneous material (basalt in the upper crust, gabbros in the lower crust) is the same, LIPs exhibit in average three times thicker crust than normal oceanic crust [*Coffin and Eldholm*, 1994; *Ridley and Richards*, 2010]. The Ontong Java Plateau even has a crustal thickness of up to 40 km [*Furumoto et al.*, 1976; *Miura et al.*, 2004]. The High Plateau of the Manihiki Plateau has a crustal thickness of 20 km [*Hochmuth et al.*, in review]. The Western Plateaus show a thinning of the crust from 17 km in the east to 9 km in the west [*Hochmuth et al.*, in review]. The crustal thickness of the Hikurangi Plateau is interfered from gravity modeling to be approximately between 17 km and 23 km [*Davy et al.*, 2008].
All those LIPs experienced phases of secondary magmatic and volcanic activity, which partly overprint tectonic sutures [*Davy et al.*, 2008; *Inoue et al.*, 2008; *Hoernle et al.*, 2010; *Pietsch and Uenzelmann-Neben*, in review].
An important key feature of LIPs is the high velocity zone (HVZ) (P-wave velocities between 7.2 and 7.7 km/s) within its lower crust. The HVZ consists of olivine and pyroxene crystal fractionation, which is trapped above the mantle – crust boundary (Moho) [*Ridley and Richards*, 2010; *Karlstrom and Richards*, 2011]. The presence of the HVZ indicates the influence of hot mantle upwelling, e.g. due to the presence of a hotspot or a mantle plume. The configuration of the HVZ along LIP margins allows an insight into tectonic deformation as well as LIP formation processes (Fig. 2). The presences of a HVZ can be interfered by refraction/wide-angle reflection seismic experiments, which have been carried out on the Ontong Java Plateau [*Furumoto et al.*, 1976; *Miura et al.*, 2004] and on the Manihiki Plateau [*Winterer et al.*, 1974; *Hochmuth et al.*, in review] (Figs. 2 and 3). Additionally to the LIPs of the western Pacific, numerous smaller magmatic events shape the area. The overriding of multiple active hotpots such as for example the Samoa, Tahiti, Pitcairn, Louisville and Society Islands hotspots altered the area along with the relicts of former hotspots traces such as the Gilbert ridge and the Tokelau seamounts.
The omnipresence of volcanically altered oceanic crust has an important impact on the plate tectonic mechanisms of the western Pacific. For example, buoyancy calculations by Cloos [1993] predict, that oceanic plateaus as thick as 17 km can be subducted and subduction orogenesis due to oceanic plateaus requires a broad volcanic edifice (approx.>100 km long and 50 km wide) with a crustal thickness of 30 km. Those calculations indicate that LIPs can play a significant role in the plate tectonic framework especially in the Pacific Ocean and influence the behavior of oceanic plates by arc polarity reversal [*Musgrave*, 1990; *Mann and Taira*, 2004] or altering subduction patterns [*Gutscher et al.*, 1999; *Liu et al.*, 2010].

2.2 The plate tectonic framework of the Pacific during the Cretaceous

The plate kinematics of the Cretaceous Pacific area are still highly debatable and include countless microplates and now subducted plates [e.g. *Seton et al.*, 2012]. During the Jurassic, the Pacific Triangle developed, which is the birthplace of today's Pacific Plate. The Pacific Triangle was formed by the Izanagi Plate in the northwest, the Farallon Plate in the northeast and the Phoenix Plate in the south, which were connected by triple junctions. Magnetic seafloor spreading anomalies can be identified from M27 (155 Ma) and M0 (120 Ma) within the Phoenix lineation northeast of the Ontong Java Plateau [*Nakanishi et al.*, 1992]. After the Cretaceous Normal Superchron, magnetic seafloor spreading anomalies can be traced from C34n (83.5 Ma) to C1 (0.8 Ma) to the east of the Manihiki Plateau (Fig.1). The whole tectonic re-organization of the Ontong Java Nui LIPs occurred during a time of a relatively stable magnetic field, which does not allow to trace the movement of individual plates. We introduce two competing models deciphering this time period presented within the literature, and will re-examine those models in the light of newly acquired data from the Manihiki Plateau: The "Super"-LIP Ontong Java Nui and the separated formation of the Ontong Java Plateau and the coupled emplacement of Manihiki and Hikurangi.

2.2.1 "The Ontong Java Nui hypothesis"

Taylor (2006) hypothesized that the three major LIPs of the western Pacific Plate were emplaced as one "Super" Large Igneous Province. The "Super"-LIP Ontong Java Nui was formed in the vicinity of the Farallon-Phoenix-Pacific triple junction approximately at 125 Ma. The trace of this triple junction is imprinted on today's Pacific Plate by a gravity anomaly called the Tongareva Triple Junction Trace, which is traceable from the Manihiki Plateau to the Pacific-Antarctic Ridge striking NW-SE [*Larson et al.*, 2002; *Viso et al.*, 2005] (Fig.1). The break-up of Ontong Java Nui was initiated at all margins of the Manihiki Plateau between 120 Ma and 118 Ma. To the south the Osbourn Trough developed separating the Hikurangi Plateau from Manihiki [*Billen and Stock*, 2000; *Worthington et al.*, 2006; *Davy et al.*, 2008]. The Ontong Java Plateau drifted away to the west by spreading in the Ellice Basin and the Nova Canton Trough [*Taylor*, 2006; *Chandler et al.*, 2012]. The northeastern fragment of the Manihiki Plateau rifted northeastwards on the Farallon Plate and the eastern part of the Manihiki Plateau was integrated into the Phoenix Plate in a southward direction [*Larson et al.*, 2002; *Viso et al.*, 2005]. The movement between the Hikurangi Plateau and the Manihiki Plateau stopped at 100 Ma with the jamming of the subduction zone at the Chatham Rise and the subsequent cessation of spreading at the Osbourn Trough [*Davy et al.*, 2008; *Davy*, 2014]. Other authors [e.g. *Billen and Stock*, 2000; *Sutherland and Hollis*, 2001; *Worthington et al.*, 2006] argue for a longer lifespan of the Osbourn Trough, but all agree on a cessation of spreading has to lay within the Cretaceous Normal Superchron (125 – 83.5 Ma). Around 80 Ma the spreading in the Nova Canton Trough between the Ontong Java Plateau and the Manihiki Plateau terminated, although the Ontong Java Plateau did not yet reach the Solomon Trench. The opening of the Ellice Basin by the Nova Canton Trough included a rotational component [*Chandler et al.*, 2013] between 37° and 52° obtained by paleomagnetic reconstructions. This rotation calls either for a decoupeling of the Ontong Java Plateau from the Pacific Plate or a so far unrecoginzed large-scale rotation of the Pacific Plate between 125 Ma and 83.5 Ma.

The main objections towards this coupled emplacement of the three LIPs include the different crustal thicknesses between the Ontong Java Plateau (< 30 km of crust) [*Furumoto et al.*, 1970; *Gladczenko et al.*, 1997; *Richardson et al.*, 2000; *Klosko et al.*, 2001; *Miura et al.*, 2004] and the conjugate margin at the Manihiki Plateau, the Western Plateaus, which present a gradual decrease of crustal thickness from 17 km to 9 km towards the Ellice Basin [*Hochmuth et al.*, in review]. If the emplacement was coupled the Western Plateaus should have a similar crustal thickness as its conjugate plateau. Additionally, the tectonic fit between the different plateaus cannot be achieved easily since secondary volcanism and tectonic activity altered the plateau's margins. A further complication of the plate kinematic reconstruction is that the Nova Canton Trough does not show a clear spreading axis, but seems to consist of multiple small ridges and fracture zones, which give indications for a scissor-like opening of the basin [*Taylor*, 2006; *Chandler et al.*, 2012].

2.2.2 Individual emplacement of Ontong Java and Manihiki-Hikurangi

Whereas the coupled emplacement of the Hikurangi Plateau and the Manihiki Plateau seems to be a well-established factor in the plate kinematics of the western Pacific, the fit between the Ontong Java Plateau and the Manihiki Plateau is still under debate. Therefore, we give an overview on alternative scenarios, which do not include a coupled emplacement between the Ontong Java Plateau and the Manihiki and Hikurangi Plateaus. Larson et al. [1972] and Winterer et al. [1974] propose a plate tectonic set-up, where the oceanic plateaus of Ontong Java Nui are situated on the spreading between the Pacific and the Antarctic Plate. The different sub-provinces of the Manihiki Plateau are created by a spreading segment jump [*Winterer et al.*, 1974] or the presence of the Farallon – Antarctic spreading on the High Plateau [*Larson and Chase*, 1972]. The Nova Canton Trough, which separates the Manihiki Plateau and the Ontong Java Plateau can be explained by the presence of the Pacific – Phoenix spreading center during the emplacement of the LIPs by individual plumes [*Larson*, 1997]. For the main magmatic stage, Larson [1997] proposes an interaction of the two plumes with the Pacific – Phoenix ridge. The Nova Canton Trough was created after the primary magmatism by reheating and extension of the young lithosphere. This concept highlights the importance of a possible ridge – plume interaction creating the LIPs of the western Pacific.

Some world-wide plate kinematic reconstructions [e.g. *Müller et al.*, 2008] do not include LIPs in their model and therefore give no indication of a possible scenario. Those plate tectonic scenarios suggest the creation of the Ellice Basin as part of the spreading between the Pacific and the Antarctic Plate.

3. Overview on published data and additional data

Before we can re-evaluate emplacement mechanisms and further tectonic activity, it is necessary to provide a condensed overview on the relevant data, which is currently available in the western Pacific region.

As previously stated, the main phase of tectonic evolution within this region occurs during the Cretaceous Normal Superchron. Small variations within the magnetic field strength within this time period have been detected in the Atlantic Ocean offshore North Africa [*Granot et al.*, 2012], but unfortunately those variations cannot be recognized within the western Pacific. The Nova Canton Trough shows no clear spreading axis, and the Osbourn Trough is magnetically overprinted by the Louisville Hotspot in the south and the smaller Austral-Cook and MacDonald Hotspots in the north [*Billen and Stock*, 2000]. Therefore magnetic data can only help to restrict a very wide timeframe for the plate re-organization.

Chandler et al. [2013] compiled all available paleolatitude data from the Deep Sea Drilling Project (DSDP), Ocean Drilling Project (ODP) and the International Ocean Drilling Project (IODP) cores on the Ontong Java Plateau (Table 1). Their findings point to an emplacement latitude of the Ontong Java Plateau between 17˚ S and 33˚ S with a rotation of the plateau between 37˚ and 52˚. This rotation is currently not integrated in any plate kinematic reconstruction and might indicate a large-scale rotation of the Pacific Plate or an individual movement of the Ontong Java Plateau during the Cretaceous Normal Superchron.

Drilled cores, reaching the crystalline basement, are rare in the area and only a very small number have published basement ages. But along with dredges e.g. from the Wishbone Scarp [*Mortimer et al.*, 2006] or the Danger Islands Troughs [*Ingle et al.*, 2007] and rocks exposed on islands e.g. Malaita [*Ishikawa et al.*, 2005; 2007; *Musgrave*, 2013] they can be used as a valuable references for the timing of local tectonic events (Table 1).

	Latitude	Longitude	Age	Paleolatitude	Reference:
ODP leg 130 - 807	3.6000	156.620	122.3	-17.9 ± 3.3	[*Mahoney et al.*, 1993]
ODP leg 192 -1183	-1.177	157.015	121	-27.9 ± 7.2	[*Riisager et al.*, 2003]

ODP leg 192 -1184	-5.011	164.223	123.5	-34.4 ± 5	[*Chambers et al.*, 2004]
ODP leg 192 -1185	-0.358	161.668	121	-23.3 ± 2.2	[*Riisager et al.*, 2003]
ODP leg 192 -1186	-0.680	159.844	121	-25.2 ± 3.5	[*Riisager et al.*, 2003]
ODP leg 192 -1187	0.943	161.451	121	-22.2 ± 2.3	[*Riisager et al.*, 2003]
DSDP leg 33 - 317	-11.0015	-165.263	116.8 ± 3.7	-47.5	[*Cockerham and Jarrard*, 1976]
So168 DR55	-40.7508	-160.916	115		[*Mortimer et al.*, 2006]
Malaita	-8.772	160.916	160		[*Ishikawa et al.*, 2005]

Table 1 additional dated locations and paleolatitude data

A further constraint, that needs to be considered are the tectonic lineations traceable in satellite gravity anomaly maps and bathymetric maps. Large scale anomalies such as the Tongareva Triple Junction trace [*Larson et al.*, 2002], the East and West Wishbone Ridges [*Mortimer et al.*, 2006] and the Manihiki Scarp [*Viso et al.*, 2005] are relicts of former plate boundaries (Fig.1). Additional information of the plate movement can be extracted from intraplate fracture zones. Taylor [2006] and Chandler [2012] examined the fracture zones within the Ellice Basin (Nova Canton Trough), which strike in an East-West direction. North-south striking fracture zones can be observed north and south of the Osbourn Trough (Fig. 1). Additionally, fracture zones dissect the Ellice Basin and the Phoenix lineations [*Nakanishi et al.*, 1992]. Another guide for the evolution of the Pacific Plate and the Pacific-Farallon spreading are the large Pacific Fracture zones e.g. Galapagos Fracture Zone.

Additionally an important factor in reconstruction the Cretaceous western Pacific is the LIPs itself. The current state of those magmatic bodies is altered by tectonic forces and volcanism of later magmatic stages and do not necessarily resemble the LIP at its emplacement. We try to account for crustal extension due to crustal stretching or massive emplacement of magmatic material, as well as for “lost” fragments to the east and north of the Manihiki Plateau [*Larson et al.*, 2002; *Viso et al.*, 2005; *Pietsch and Uenzelmann-Neben*, in review].

4. Indications of possible break-up mechanisms as seen on the Manihiki Plateau

The Manihiki Plateau plays an important role in the plate tectonic set-up of the Pacific during the Cretaceous, since it potentially exposes break-up margins towards the other LIPs of the region and the seafloor emplaced during the Cretaceous Normal Superchron. A close examination of the crustal structure along with the magmatic and tectonic activity displayed in high resolution seismic reflection data [*Pietsch and Uenzelmann-Neben*, in review] and refraction/wide-angle reflection seismic data [*Hochmuth et al.*, in review] acquired in 2012 [*Uenzelmann-Neben*, 2012] allows us to identify possible break-up mechanisms on the Manihiki Plateau. Additionally it is important for further reconstructions to incorporate the amount of crustal growth created by later magmatic stages and tectonic stress (Fig. 2).

Fig. 2 Relicts of tectonic stress on the Manihiki Plateau as seen in seismic reflection lines from So-224 and KIWI-98 (thin black lines) and refraction seismic lines (thick black lines) from So-224. Dashed areas indicate faulted basement. The dotted areas show no to little faulting within the basement. The yellow line indicates a series of troughs (e.g. Danger Islands Troughs) active during the supposed break-up of Ontong Java Nui. The orange line marks the Suvarov Trough, which was active after the initial break-up. The red lines on the refraction seismic profiles indicate the presence and thickness of the High Velocity Zone, within the P-wave *models. The position of the refraction seismic profiles shown in Fig. 3 is indicated by the light grey boxes.*

The Manihiki Plateau was created by a first phase of extrusive volcanism with an approximated age of 125 Ma [*Timm et al.*, 2011]. Later magmatic stages can be traced throughout the High Plateau [*Pietsch and Uenzelmann-Neben*, in review] and contributed in some areas (e.g. the southern High Plateau) to the spatial extent of the LIP. Although the LIP did not grow significantly in area due to later stage magmatism (< 65 Ma), the amount and mode of the secondary magmatic phases differ between the High Plateau, where multiple outlets of secondary magmatic stages can be seen, and the Western Plateaus, which show low-volume secondary magmatism [*Pietsch and Uenzelmann-Neben*, in review; *Hochmuth et al.*, in review]. The boundary between those two magmatic regimes is formed by the Danger Islands Troughs.

A more important factor for assessing the extension of the crust after the initial emplacement of the LIP is the tectonic stress (Fig.2), which is visible by countless faults (e.g. High Plateau) and the decrease of crustal thickness (e.g. Western Plateaus). Additional information on the extent of the LIP influenced crust can be contributed from the presence of a high velocity zone (HVZ) with p-wave velocities above 7.3 km/s in the lower crust of the plateaus (Fig.3). The HVZ is a result of the upwelling of very hot mantle material and is a key feature for identifying oceanic LIPs.

We can identify four different areas of tectonic behavior on the Manihiki Plateau. On the central High Plateau tectonic activity is little and mainly induced by magmatic activity and the collapse of magmatic structures [*Pietsch and Uenzelmann-Neben*, in prep.] (Fig.2). The eastern flank of the High Plateau, the Manihiki Scarp exhibits a N-S trending sheared margin [*Larson et al.*, 2002; *Viso et al.*, 2005; *Pietsch and Uenzelmann-Neben*, in review; *Hochmuth et al.*, in review] with up to eight basement ridges exposing lower crust (Fig.3a). The HVZ terminates below the basement ridges and crustal thickness decreases from 15 km to 4.5 km within 60 km [*Hochmuth et al.*, in review]. Additional crustal material seems to be emplaced by the exposure of lower crustal material and not by stretching processes.

The southern High Plateau shows multiple normal fault systems, which can be related to rifting activity during the Cretaceous and later tectonic stress (40 - 1.8 Ma) [*Pietsch and Uenzelmann-Neben*, in prep.] (Fig.2). This area has also been influenced by secondary magmatic stages (> 65 Ma) and even younger magmatic activity (23 – 10 Ma). The HVZ in the lower crust of the Manihiki Plateau stretches into the Samoan Basin, which indicates the influence of the LIP on the initial spreading process (Fig.3b). The crustal stretching (β) is evident but relatively small (β=1,26).

The western High Plateau and the Western Plateaus show low volume secondary magmatism [*Pietsch and Uenzelmann-Neben*, in review; *Hochmuth et al.*, in review]. In refraction/wide-angle reflection data from the Western Plateaus, we can observe a constant presence of the HVZ and a decrease in crustal thickness from 18 km in the East to 9 km in the West (β=2) over 400 km (Fig.3c). This indicates a potential overlap with a conjugate margin of 200 km. Small and large offset faults are present throughout the sub-province and an almost undisturbed center comparable to the central High Plateau cannot be observed within the available data sets (Figs. 2 and 3).

Another significant feature of the Manihiki Plateau are its internal troughs, the S-N trending Danger Islands Troughs and the NE-SW trending Suvarov Trough (Fig. 2). Seismic reflection data reveals that the Suvarov Trough is younger than 65 Ma and can therefore not be a result of the initial tectonic activity within the Cretaceous Normal Superchron [*Pietsch and Uenzelmann-Neben*, in prep.]. Dated dredge samples from the Danger Island Troughs show an ^{40}Ar-^{39}Ar age of 117.9 ± 3.9 a and 99.5 ± 0.7 a [*Ingle et al.*, 2007] and indicate a multistage source history with multiple phases of magmatic activity. Refraction/wide-angle reflection seismic data reveals the lack of typical upper crustal material within the trough but a relatively undisturbed lower and middle crust [*Hochmuth et al.*, in review]. The Danger Islands mark a significant border between the two magmatic and tectonic regimes of the High Plateau and the Western Plateaus. By tracing the exposed fault systems in bathymetry (GEBCO 2014) and global satellite gravity anomaly maps [*Sandwell and Smith*, 1997] (V23), a rotational component from NNE-SSW striking features in the North to NNW-SSE striking features in the South can be observed [*Nakanishi et al.*, 2015]. This supports the hypothesis, that the Western Plateaus and the High Plateau acted as individual tectonic plates during a part of the Cretaceous.

Fig. 3 Examples for P-wave models crossing the different margins of the Manihiki Plateau. The position of the different profiles is indicated in Fig. 2. Figure 3 a) shows the Manihiki Scarp - a sheared margin, Fig. 3 b) shows the southern High Plateau - a stretched margin with volcanic overprint and Fig. 3 c) shows a part of the Western Plateaus - a stretched margin with little magmatic activity
For more information on the experimental set-up and a view of whole profiles see Hochmuth et al. [in review] and Uenzelmann-Neben [2012].

5. A joined emplacement of the Manihiki Plateau and the Ontong Java Plateau?

The emplacement of the Manihiki Plateau and the Hikurangi Plateau as a coupled LIP and the consecutive spreading at the Osbourn Trough has been well established in the literature [*Billen and Stock*, 2000; *Worthington et al.*, 2006; *Davy et al.*, 2008]. Plate tectonic concepts proposed for example by Larson [1997] explain the emplacement of the Manihiki Plateau and the Ontong Java Plateau by the contemptuous rising of two plume heads, which are separated by the Pacific-Phoenix Ridge. Further support for this scenario are the shortcomings of "Ontong Java Nui hypothesis" [*Taylor*, 2006] such as the different crustal thickness between the Manihiki Plateau and the Ontong Java Plateau. By the emplacement of the LIPs by two separated plumes, the differences in crustal thickness from >30 km at the Ontong Java Plateau [*Furumoto et al.*, 1970; *Gladczenko et al.*, 1997; *Klosko et al.*, 2001; *Miura et al.*, 2004] and 20 km on the High Plateau and 17 to 9 km on the Western Plateaus [*Hochmuth et al.*, in review] would be explained. The thinning of the crust on the Western Plateaus would be a result of the distance to the center of emplacement, which was located on the High Plateau of the Manihiki Plateau. Although the gradual decrease in crustal thickness along with numerous normal faults within the basement of the Western Plateaus, might point to crustal extension processes. The Nova Canton Trough has been overprinted by the interaction of the plumes with the spreading ridge between the Pacific Plate and the Phoenix Plate. Magnetic anomaly data [*Nakanishi et al.*, 1992] shows, that the spreading axis of the Phoenix-Pacific ridge does not coincide with the assumed main spreading direction of the Nova Canton Trough, mainly to the west, where the M10 lineation borders the Cretaceous seafloor created by the oblique spreading within the Nova Canton Trough. Furthermore, the rotation of the Ontong Java Plateau can be explained by the scissor-like opening of the Nova Canton Trough [*Taylor*, 2006; *Chandler et al.*, 2012]. If this spreading and the emplacement of the LIPs were contemptuous the rotation would not be inherited by the spreading, but be a relict of other processes, which have not been observed so far. Additionally it might be complicated to distinguish the effect of three different sources – an Ontong Java Plume, a Manihiki-Hikurangi Plume and the spreading center between Pacific and Phoenix Plate – on such a confined area. Interactions between those upwelling centers are so far undetermined. It seems possible, that the igneous volume emplaced could interact and intermingle within the crust.
Even though, the hypothesis by Taylor [2006] is still debated and not all the mechanisms of the emplacement are fully understood, we like to explore this concept of a joint emplacement of the Manihiki Plateau and the Ontong Java Plateau further. The interaction between the spreading ridge and the plume emplacing the LIP seems nevertheless a crucial aspect of the plate tectonic reconstruction.

6. Solving the jigsaw – the reassembling of Ontong Java Nui

6.1 Classification of the margins of the LIPs

We investigated the different margins of the Manihiki Plateau, by all available seismic reflection data and refraction/wide-angle reflection seismic data, to be able to characterize the break-up occurring at the different margins and possibly connect the multiple margin types to their corresponding margins on the Hikurangi Plateau and Ontong Java Plateau (Table 2). For the interpolation across margins and plateaus, we used global bathymetry (GEBCO_2014) and satellite gravity anomaly maps [*Sandwell and Smith*, 1997] (V23). We identified six margin types on all Ontong Java Nui related LIPs (Fig.4).

Margin	characteristica	regional example	citation	interfered from
tectonically inactive margin	slow decrease in depth, basalt flows into the oceanic basin, dip angle < 0.1 degrees	northern Ontong Java Plateau	[*Mochizuki et al.*, 2005]	seismic reflection data, bathymetry, gravity anomaly
subducting margin	subduction of LIP crust	Ontong Java Plateau – Solomon Trench Hikurangi – Chatham Rise	[*Miura et al.*, 2004; *Davy et al.*, 2008; *Davy*, 2014]	refraction seismic, bathymetry, gravity anomaly
sheared margin	rough topography with multiple ridges exposing lower crustal layers, sudden termination of HVZ	Manihiki Scarp	[*Larson et al.*, 2002; *Viso et al.*, 2005; *Ai et al.*, 2008; *Hochmuth et al.*, 2014; *Pietsch and Uenzelmann-Neben*, in review]	seismic reflection data. seismic refraction data, bathymetry, gravity anomaly
stretched margin with little magmatic activity	countless large and small offset faults, low volume secondary magmatism, constant HVZ, massive crustal stretching	Western Plateaus (Manihiki Plateau), Rekohu Embayment (Hikurangi Plateau)	[*Davy et al.*, 2008; *Hochmuth et al.*, in review]	refraction seismic data, reflection seismic data, bathymetry, gravity anomaly
stretched margin with magmatic overprint	multiple fault systems, massive magmatic activity during later magmatic stages, small amount of crustal stretching	southern Manihiki Plateau, Southeast High Plateau (Hikurangi Plateau)	[*Davy et al.*, 2008; *Hochmuth et al.*, 2014; *Pietsch and Uenzelmann-Neben*, in review]	refraction seismic, seismic reflection data, gravity anomaly, bathymetry
rifted margin	very short LIP - ocean basin transition area, sharp boundary, sudden depth decrease dip angle > 5 degrees	Rapuhia Scarp (Hikurangi Plateau)	[*Davy and Collot*, 2000; *Davy et al.*, 2008]	reflection seismic, bathymetry, gravity anomaly

Table 2. Overview on the different margins of Ontong Java Nui and their individual features

The Manihiki Plateau shows break-up related features on all its margins (Figs. 2, 3, 4). Those can be related to corresponding margins on the Hikurangi Plateau and the Ontong Java Plateau (Fig.4). It is important to note, that the possible break-up of Ontong Java Nui included rotational components on the Ontong Java Plateau [*Chandler et al.*, 2013] and on the Hikurangi Plateau [*Davy*, 2014]. Those rotational components have so far not been included in any plate tectonic reconstructions. Possible former adjacent margins could be subducted underneath the Australian Plate. For an improved plate tectonic model it is crucial to account for rotational components as well as for the crustal growth due to crustal stretching and secondary magmatism. Since the secondary magmatism was emplaced during and after the main tectonic activity, the initial LIP does not include the magmatic volume emplaced during those phases.

Fig. 4 Classification of the margins of Ontong Java Nui in their current setting (main figure) and during their emplacement (inlet figure).

6.2 Ontong Java Nui at the time of emplacement

We connected the presumably corresponding margins of the LIPs (Fig. 4) and removed the crustal extension introduced after the break-up. The crustal extension and the resulting overlap between the LIPs has been calculated by inferring the stretching coefficient (β) from refraction/wide-angle reflection data and calculating the overlap (o) by $o=w*(\beta-1)/\beta$. The width (w) of the stretched crust is taken from seismic reflection data. The rotational components such as the rotation of the Ontong Java Plateau [*Chandler et al.*, 2013] have also been included (Fig.4 inlet). The subducted parts of the LIPs are added to the initial Super-LIP. Here we used the traceable subducted slap of the Hikurangi Plateau [*Reyners*, 2013] below New Zealand and estimated extensions of the Ontong Java Plateau by Musgrave [2013]. Since the northeastern and eastern fragments of the Manihiki Plateau are subducted, we estimated the extent of those fragments with the assumption, that the emplacement mechanism was similar to the northern Ontong Java Plateau, where basalt flows border the Nauru Basin [*Mochizuki et al.*, 2005]. Those assumptions result in an initial size of Ontong Java Nui of 1.1% of the Earth's surface, which is larger then previously anticipated.

Fig. 5 Comparison of paleolatitude data (dashed area of possible emplacement of Ontong Java Nui) and calculated isochrones [Seton et al., 2012] and magnetic anomaly picks [Nakanishi et al., 1992]. The grey area indicates the Ontong Java Plateau in relation to the magnetic anomaly picks.

By comparing the reassembled LIP with recent global plate tectonic models [e.g. *Seton et al.*, 2012] and paleomagnetic anomaly picks [*Nakanishi et al.*, 1992] it becomes obvious, that the paleolatitudes calculated for the Ontong Java Plateau are approximately 400 km further north than the reconstructed position of the Ontong Java Plateau (Fig. 5). Even though, paleolatitude calculations are few and imply a large error margin, we investigate further possible factors for this significant offset. The Phoenix lineations show multiple fracture zones (FZ) within their sequence [*Nakanishi et al.*, 1992] including the Phoenix FZ and the Central Pacific FZ (Fig. 6). These induced a considerable offset between the magnetic lineations in the vicinity of the Ontong Java Plateau and are, along with other smaller FZ traceable within the eastern Nova Canton Trough [*Sandwell and Smith*, 1997; *Maus et al.*, 2009] (Fig. 6). The crust of the Nova Canton Trough was emplaced after the emplacement of the oceanic LIPs, which allows an emplacement of the LIP further north with later, post-emplacement movement towards the south. Koppers et al. [2005] inferred from asynchronous bents in the seamount chains of the Gilbert Ridge and the Tokelau seamounts, two short extensional phases within the Nova Canton Trough (67 Ma and 57 Ma), which might be related to a re-activation of the trough or to the activity of fracture zones. If those fracture zones were active after the Cretaceous spreading in the Nova Canton Trough, they can at least partly account for the offset between reconstructed paleolatitudes and calculated paleolatitudes. We, therefore infer an emplacement of Ontong Java Nui between 18 °S and 40 °S. The absolute plate motion of the Pacific Plate during the Cretaceous is vaguely constraint by direct measurements from basaltic flows, but a hook-like shape of the absolute polar wander path is proposed [*Sager*, 2006; *Wessel and Kroenke*, 2008]. Unfortunately the data from the Ontong Java Plateau does not fit this path. Sager [2006] proposes a decoupling of the northern and southern Pacific Plate – including the Ontong Java Plateau - during the Cretaceous. Paleo-plate boundaries are not observed within the Jurassic Pacific Plate, which makes this uncoupling rather unlikely. If the Pacific Plate and the Ontong Java Plateau were coupled during the early Cretaceous, it has to be inferred that the rotation of the Ontong Java Plateau was at least partly also performed by the Pacific Plate. Unfortunately the data to distinguish between those scenarios is not available yet. Additionally, the southern hemisphere Pacific is underrepresented in the calculations of the rotation poles [*Sager*, 2006], which might underestimate possible rotations.

Fig. 6 a) magnetic anomaly map of the Nova Canton Trough after Maus et al. [2009], grey areas indicate the oceanic LIPs b) tectonic interpretation with major fracture zones in green from Nakanishi et al. [Nakanishi et al., 1992] for the Phoenix lineations and additional smaller fracture zones at the convergence between the Nova Canton Trough and the Clipperton FZ , fracture zones within the Nova Canton Trough (yellow area) after Taylor [2006] in black and the magnetic isochrones of M10 (red) and M1 (orange) within the Phoenix lineations [Nakanishi et al., 1992].

7. Plate tectonic reconstruction of the western Pacific (120-80 Ma) with special focus on the Manihiki Plateau

The plate kinematic reconstruction presented here is based on the global tectonic model of Seton et al. [2012]. We additionally used the hotspot reference W&K08-D by Wessel and Kroenke [2008], which was published by Chandler et al. [*2013*]. The model comprises the time frame from 125 Ma to 80 Ma and translates directly into the model by Seton et al. [*Seton et al.*, 2012] for the development after the Cretaceous Normal Superchron. An overview on the modeled tectonic events is provided in Table 3.

7.1. Emplacement of Ontong Java Nui – different crustal thicknesses on the plateau – result of a plume ridge interaction? (>120 Ma)

The former spreading at the Pacific – Phoenix ridge cross cuts the projected emplacement area of the Manihiki Plateau. Larson [1997] implies that the presence of the mid-ocean ridge separated to individual plumes. This separation could cause two or more centers of main magmatic activity, which correlate with especially thick crust. In our model, areas of potentially thick crust – the High Plateau of the Manihiki Plateau and the Ontong Java Plateau – are separated by the spreading center at approximately the same distance from the ridge (Fig. 7a). Therefore, the High Plateau of the Manihiki Plateau and the High Plateau of the Ontong Java Plateau lay on crust of similar Jurassic age. The constant HVZ beneath the Manihiki Plateau shows that the fragmentation and thinning of the crust is an imprint of later crustal stress and not a relict of the initial emplacement of the LIP. The interaction between an arriving plume head and a mid-ocean ridge is not yet completely understood [*Dyment et al.*, 2007; *Ito et al.*, 2003; *Mittelstaedt and Ito*, 2005] A similar scenario can currently be observed in Iceland [*Flóvenz and Gunnarsson*, 1991; *Darbyshire et al.*, 1998; *Du and Foulger*, 2001] and relict examples can be seen e. g. at the Galapagos hotspot [*Sinton et al.*, 2003; *Thompson et al.*, 2004; *Kokfelt et al.*, 2005], but in all these scenarios the size of the upwelling mantle is smaller then the assumed size of a plume head. If the ridge plays a role in this scenario multiple possibilities have to be considered. First the plume-head could simply swallow the ridge and overprint the entire area by its massive outpourings. This should lead to even more thickened crust in the area of the ridge. Seafloor spreading should cease during the arrival of the plume. This scenario fails to explain the differences in crustal thickness between the plateaus, but points to a single centered emplacement. Another possible scenario could be similar to the mechanism proposed by Larson [1997], that the upwelling of mantle material at the ridge separated the plume head into smaller portions resulting in multiple main outlets. If the ridge was active between magmatic pulses, the crust emplaced at the ridge would further separate the magmatic main centers and therefore the areas of thickest crust. The interaction with the spreading ridge could account for the astonishing differences in crustal thickness between the Manihiki Plateau and the Ontong Java Plateau. Unfortunately we have only little constraints for the crustal thickness of the Hikurangi Plateau to test different hypotheses. Furthermore it would be interesting to be able to constrain the thickness of the subducted crust to map the areas of more and less magmatic input. The presence of very young crust beneath the Western Plateaus of the Manihiki Plateau could also explain the later establishment of the Nova Canton Trough. Younger and more ductile crust was first deformed and spreading was initiated later (see. 7.2 and Tab. 3).

7.2. Initial break-up of Ontong Java Nui – evaluation of the break-up mechanisms on an oceanic LIP (120 Ma – 116 Ma)

The development of break-up margins can be traced along the different margins of the Manihiki Plateau. This initial phase of break-up is rather chaotic and includes multiple tectonic components such as shearing and crustal stretching (Fig. 7b). Additionally, the magmatic activity is still strong on the LIPs [*Inoue et al.*, 2008; *Hoernle et al.*, 2010; *Pietsch and Uenzelmann-Neben*, in review], which leads to overprinting and alteration of the tectonic evidence. We present the different break-up mechanisms in a clock-wise fashion starting to the south, where the Osbourn Trough develops [*Billen and Stock*, 2000; *Worthington et al.*, 2006; *Downey et al.*, 2007]. First movement between the plateaus include a rapid separation between the Rapuhia Scarp (Hikurangi Plateau) and the southern Western Plateaus, whereas crustal stretching can be observed at the eastern Hikurangi Plateau and the High Plateau of the Manihiki Plateau (Fig.3b). The presence of a thin HVZ below the Samoan Basin seems not to be related to this break-up, but could be the result of a later overprint of the southern High Plateau by Tahiti – Society Islands Hotspot [*Pietsch and Uenzelmann-Neben*, in prep.]. On the southern tip of the High Plateau the Osbourn Spreading Center intersects with the Manihiki Scarp. This shearing zone (Fig.3a) can be traced along the eastern High Plateau and established itself as the eastern plate boundary of the Manihiki Plateau towards the Eastern Manihiki Plate, which comprises the seafloor between the Tongareva Triple Junction and the Manihiki Scarp (Fig. 7b). Seton et al. [2012] postulate that two plates develop during this time, the Chasca Plate (north) and the Catequil Plate (south), representing the former Phoenix Plate and the southern Farallon Plate (south of the Clipperton FZ). This reorganization facilitates the movement at the Tongareva Triple Junction and has been incorporated into our model. At the Manihiki Scarp, the eastern fragment of the Manihiki Plateau has been moved to the south and gets incorporated into the Phoenix Plate (Catequil Plate). The Tongareva Triple Junction, which consists of the spreading between the Pacific Plate, the Phoenix Plate and the Farallon Plate jumps to the northeastern corner of the Manihiki Plateau [*Larson et al.*, 2002; *Pockalny et al.*, 2002]. The northern margin of the Manihiki Plateau is unfortunately mostly unsurveyed, but bathymetry [*Nakanishi et al.*, 2015] and gravity data [*Sandwell and Smith*, 1997] indicates the presence of massive tectonic activity possibly related to shearing processes. We propose a fast clock-wise rotation of the northeastern fragment of the Manihiki Plateau, resulting in

multiple ridges (Fig. 1) and the possible extension of the crust on the northern High Plateau (Fig. 2). The northern fragment is incorporated into the Farallon Plate (Chasca Plate) and moves eastwards.
The Danger Islands Troughs are also a product of this large-scale reorganization of the plate boundaries. The rotation of the Ontong Java Plateau [*Chandler et al.*, 2013] and the rotation of the Western Plateaus create the Danger Islands Troughs as a series of pull – apart basins, which show also a clock-wise rotation (Fig. 7b). We propose, that some of the rotation calculated for the Ontong Java Plateau has also occurred on the Western Plateaus. This rotation could result in the decrease of crustal thickness and faulting on the Western Plateaus. The spreading in the Nova Canton Trough is not established yet, although it is possible, that faulting is already active in the area. The tectonic activity at the later Nova Canton Trough is not parallel to the former spreading center between the Pacific and the Phoenix Plate possibly also due to the rotational forces acting on the Pacific Plate. The Hikurangi Plateau and the Ontong Java Plateau are separated by the former spreading center between the Pacific and Phoenix Plate, which possibly developed a transform motion. The reconstruction in this area is very difficult and can only be achieved be crude estimations of the subducted seafloor. Musgrave [2013] proposes an additional triple junction in this area to account for the so called Malaita terranes. The rotation of the Ontong Java Plateau was not included in this study and possibly facilitates the combination of the Ontong Java Plateau and the Malaita terrane, which lay on 160 Ma old crust.

7.3. Dispersal of Ontong Java Nui over the western Pacific – plate motion during the Cretaceous Normal Superchron (116 Ma – 83 Ma)

After the initial break-up, that involved a tremendous amount of crustal stretching, short – lived spreading centers and rotational forces, the plate boundaries stabilize. The timing of the stabilization correlates with the fading of massive volcanic activity on the Manihiki Plateau [*Pietsch and Uenzelmann-Neben*, in review]. Therefore the influence of the plume ceases and secondary phases of magmatic stages show a clearly weaker and more localized volcanic emplacement. The Osbourn Trough develops a half spreading rate of 10 cm/a (116 Ma – 100 Ma) and slows down the production of new crust significantly after the soft-docking at the Chatham Rise [*Davy et al.*, 2008; *Davy*, 2014]. The interaction with the Chatham Rise also includes the rotation of the Hikurangi Plate, which can also be observed in the change in orientation of the Osbourn Trough. The morphology of the Osbourn Trough resembles a slow – spreading ridge [*Downey et al.*, 2007; *Billen and Stock*, 2000]. Therefore, we propose a change in orientation from NW-SE to W-E after the softdocking of the LIP crust at the continental Chatham Rise and a slowing of the spreading rate to 3 cm/a, which is consistent with previous publications [*Billen and Stock*, 2000; *Downey et al.*, 2007] calculating spreading rates . The Hikurangi Plate subducts beneath the Gondwana Margin at the location of the Chatham Rise (Fig. 7c). This docking event has a great impact on the whole western Pacific and leads Matthews et al. [2012] to propose that kinks within fracture zones can be correlated and dated to this event. In our reconstructions we can also link reorientations of fracture zones observed on the Hikurangi Plate, the Manihiki Plate and the Eastern Manihiki Plate to this time frame.
To the East, the Wishbone Scarp an interoceanic subduction zone develops [*Mortimer et al.*, 2006] representing the plate boundary between the Phoenix (Catequil) Plate and the Hikurangi Plate. At the Tongareva Triple Junction, the intersection of the Manihiki Plate, the Farallon Plate (Chasca) and the Phoenix Plate (Catequil), the mechanism swaps between a ridge-ridge-ridge and ridge-ridge-transform configurations [*Larson et al.*, 2002]. The Manihiki Plateau is decoupled from the Pacific Plate by the Clipperton Fracture Zone and moving slightly eastwards (Fig. 7c). The movement at the Danger Islands Troughs stops around 110 Ma due to the establishment of an oblique spreading within the Nova Canton Trough. This indicates, that the different sub-provinces of the Manihiki Plateau acted as individual plates for a short time, but still inherited a

significant different crustal structure during this time. For the Nova Canton Trough a scissor-like opening has been proposed by Taylor [2006] and Chandler et al. [2012] separating the Ontong Java Plateau from the Manihiki Plateau (Fig. 7c).

7.4. Incorporation of Ontong Java Nui into the Pacific Plate – Pacific plate motion after the Cretaceous Normal Superchron (< 83 Ma)

After the hard-docking of the Hikurangi Plateau with the Chatham Rise, subduction at the Gondwana Margin ceases, leading to one of the largest reorganizations within the plate tectonic framework of the Pacific. Seafloor-spreading stops around the Manihiki Plateau and between the different fragments of Ontong Java Nui (Fig. 7d). With the establishment of the spreading in the Bellingshausen Sea, the different plateaus are firmly integrated into the Pacific Plate. Younger tectonic activity can mainly be related to hotspot volcanism. Koppers et al. [2005] identified tectonic activity within the area of the Nova Canton Trough at 67 Ma and 57 Ma leading to the reorientation of the Gilbert Ridge and Tokelau seamount chain respectively. Multiple fracture zones could be identified at the junction between the Nova Canton Trough and the Clipperton FZ, as well as at the central Nova Canton Trough (Fig. 6). This motion could also be responsible for the southward movement of the Ontong Java Plateau and the re-activation of the Danger Islands Troughs and the creation of the Suvarov Trough on the Manihiki Plateau [*Pietsch and Uenzelmann-Neben*, in prep.].

8. Conclusions:

We present a revised version of the plate kinematic model of the Western Pacific during the Cretaceous Normal Superchron. In our reconstruction, we include available paleolatitude calculations and collected seafloor ages. Since no magnetic isochrones were created during this time the plate kinematic reconstruction has to rely on the tectonic lineations imprinted on today's seafloor. In the western Pacific, vast areas with overthickend oceanic crust, the large igneous provinces of the Manihiki Plateau, the Hikurangi Plateau and the Ontong Java Plateau are present. By re-evaluating of the different margins of the LIPs, we unraveled possible break-up mechanisms after their joint emplacement as Ontong Java Nui. The Manihiki Plateau built the centerpiece of this "Super"-LIP and shows the multi-faced nature of the LIP-break-up within its crust. Our plate kinematic model includes:

- The initial emplacement of Ontong Java Nui in vicinity of the Pacific-Phoenix spreading ridge, resulting in huge differences within the LIP's crustal thickness due to the interaction of the spreading ridge and the arriving plume head.
- The rotation of the Ontong Java Plateau along with the Pacific Plate and the Western Plateaus of the Manihiki Plateau opening a series of pull-apart structures, the Danger Islands Toughs.
- The initiation of spreading at the Osbourn Trough included crustal stretching on the High Plateau of the Manihiki Plateau and a rotational component.
- Strong sharing forces separated the Manihiki Plateau's eastern and northeastern fragments before their incorporation into the Farallon and Phoenix Plate.

Fig. 7 Overview on the development of the western Pacific at a) 125 Ma b) 117 Ma c) 110 Ma d) 83 Ma
Subduction zones are shown in light blue, transform faults in orange, spreading centers in blue, oblique spreading and rotation in green; The different parts of the Manihiki Plateau are labeled as follows:WP Western Plateaus, HP High Plateau, NE Northeastern fragment, E Eastern Fragment; B.P. stands for the Bellingshausen Plate the Tongareva Triple Junction is marked by the red star.
The figures are created with GPlates 1.5 and GMT 4.

Table 3. Overview on the evolution of the western Pacific (120 – 80 Ma) with special focus on the oceanic LIPs of the region

	Manihiki Plateau	Ontong Java Nui	Pacific
prior to 125 Ma	• Subaerial emplacement as a single crustal unit by massive volcanic outpourings	• emplacement by massive volcanic outpourings with multiple main centers of activity	• formation of the Pacific Triangle (180 Ma) • seafloor spreading at multiple triple junctions
120 Ma	• high magmatic activity on the High Plateau • limited magmatic activity on the Western Plateaus	• initial movement between Manihiki and Hikurangi rotation and crustal stretching	
118 Ma	• initiation of the fragmentation of the Manihki Plateau • creation of the Manihiki Scarp, spreading north of the High Plateau • creation of the Danger Islands Troughs • initiation of crustal stretching at the Western Plateaus	• Development of the Osbourn Spreading center between Manihiki and Hikurangi • rotation of the Ontong Java Plateau along with the Western Plateaus	• Triple Juction jump (PAC-FAR-PHO) (Tongareva Triple Junction), • reorganization of plate tectonic framework • possible initiation of the rotation of the Pacific Plate • southward migration of Farallon Phoenix spreading along Tongareva Triple Junction trace
115 Ma	• incorporation of NE – Manihiki into Farallon Plate	• initiation of spreading at Nova Canton Trough	• initiation of ocean – ocean subduction at West Wishbone Scarp
105 Ma		• first interaction between Hikurangi Plateau and Chatham Rise • rotation of Osbourn Trough • Soft – docking of Hikurangi Plateau with Chatham Rise	
100 Ma	• final development of the Danger Islands Troughs	• Establishment of oblique spreading at Nova Canton Trough	
80 Ma		• Cessation of southwards subduction of Hikurangi Plateau • Cessation of spreading at Nova Canton Trough • Incorporation into Pacific Plate	• Full establishment of spreading within the Bellingshausen Sea • Cessation of subduction at West Wishbone Scarp

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References:

Ai, H.-A., J. M. Stock, R. Clayton, and B. Luyendyk (2008), Vertical tectonics of the High Plateau region, Manihiki Plateau, Western Pacific, from seismic stratigraphy, *Marine Geophysical Research*, *29*(1), 13–26, doi:10.1007/s11001-008-9042-0.

Billen, M. I., and J. Stock (2000), Morphology and origin of the Osbourn Trough, *Journal of Geophysical Research*, *105*(B6), 13481–13489, doi:10.1029/2000JB900035.

Chambers, L. M., M. S. Pringle, and J. G. Fitton (2004), Phreatomagmatic eruptions on the Ontong Java Plateau: an Aptian 40Ar/39Ar age for volcaniclastic rocks at ODP Site 1184, *Geological Society*, *229*, 325–331, doi:10.1144/GSL.SP.2004.229.01.18.

Chandler, M. T., P. Wessel, and W. W. Sager (2013), Analysis of Ontong Java Plateau palaeolatitudes: evidence for large-scale rotation since 123 Ma? *Geophysical Journal International*, *194*(1), 18–29, doi:10.1093/gji/ggt075.

Chandler, M. T., P. Wessel, B. Taylor, M. Seton, S.-S. Kim, and K. Hyeong (2012), Reconstructing Ontong Java Nui: Implications for Pacific absolute plate motion, hotspot drift and true polar wander, *Earth and Planetary Science Letters*, *331-332*, 140–151, doi:10.1016/j.epsl.2012.03.017.

Cloos, M. (1993), Lithospheric buoyancy and collisional orogenesis: Subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts, *Geological Society of America Bulletin*, *105*(6), 715–737.

Cockerham, R. S., and R. D. Jarrard (1976), Paleomagmatism of some leg 33 sediments and basalts, *Inital Reports of the Deep Sea Drilling Project*, doi:doi:10.2973/dsdp.proc.33.1976.

Coffin, M. F., and O. Eldholm (1994), Large igneous provinces: crustal structure, dimensions, and external consequences, *Reviews of Geophysics*, *32*, 1–36, doi:10.1029/93RG02508.

Darbyshire, F. A., I. T. Bjarnason, R. S. White, and Ó. G. Flóvenz (1998), Crustal structure above the Iceland mantle plume imaged by the ICEMELT refraction profile, *Geophysical Journal International*, *135*(3), 1131–1149, doi:10.1046/j.1365-246X.1998.00701.x.

Davy, B. (2014), Rotation and offset of the Gondwana convergent margin in the New Zealand region following Cretaceous jamming of Hikurangi Plateau large igneous province subduction, *Tectonics*, doi:10.1002/2014TC003629.

Davy, B., and J. Y. Collot (2000), The Rapuhia Scarp (northern Hikurangi Plateau) — its nature and subduction effects on the Kermadec Trench, *Tectonophysics*, *328*(3-4), 269–295, doi:10.1016/S0040-1951(00)00211-0.

Davy, B., and R. Wood (1994), Gravity and magnetic modelling of the Hikurangi Plateau, *Marine Geology*, *118*(1-2), 139–151, doi:10.1016/0025-3227(94)90117-1.

Davy, B., K. Hoernle, and R. Werner (2008), Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history, *Geochemistry*, doi:10.1029/2007GC001855.

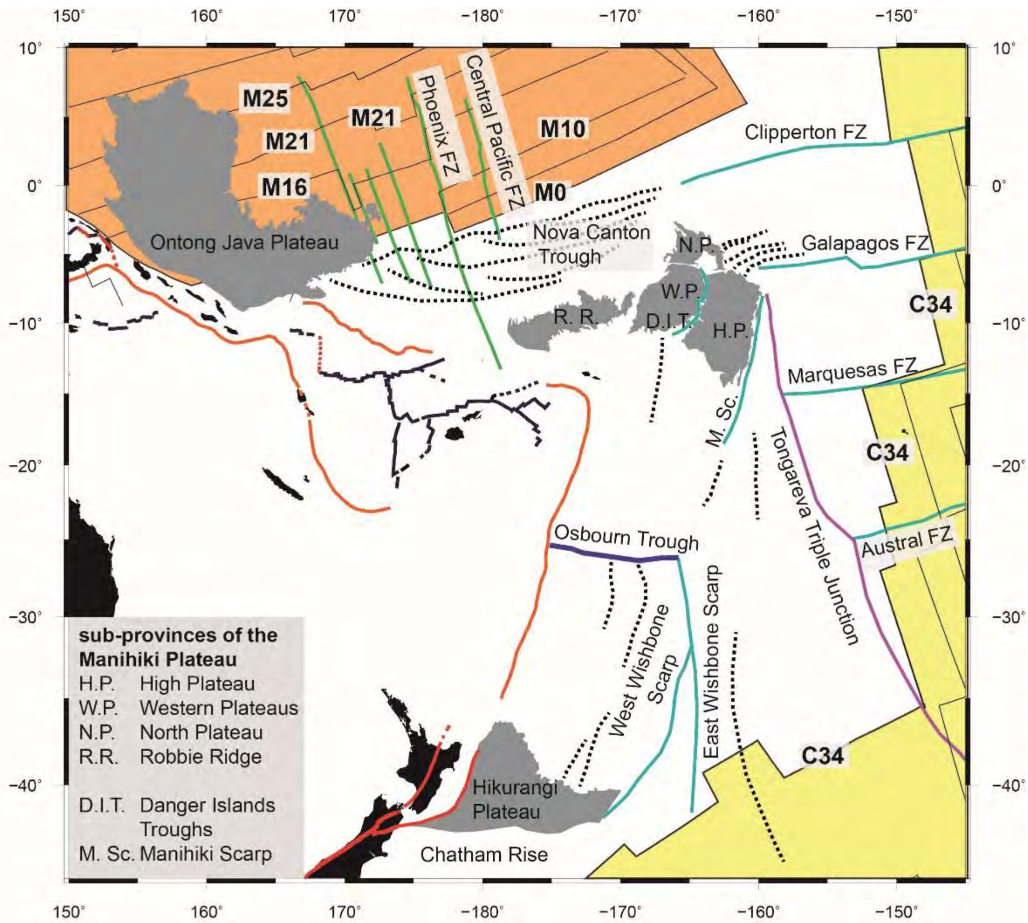
Davy, B., V. Stagpoole, D. Barker, and J. Yu (2012), Subsurface structure of the Canterbury region interpreted from gravity and aeromagnetic data, *New Zealand*

Journal of Geology and Geophysics, *55*(3), 185–191.
Downey, N. J., J. M. Stock, R. Clayton, and S. C. Cande (2007), History of the Cretaceous Osbourn spreading center, *Journal of Geophysical Research*, *112*, B04102.
Du, Z., and G. R. Foulger (2001), Variation in the crustal structure across central Iceland, *Geophysical Journal International*, *145*(1), 246–264, doi:10.1111/j.1365-246X.2001.00377.x.
Dyment, J., J. Lin, and E. T. Baker (2007), Ridge-hotspot interactions: What mid-ocean ridges tell us about deep Earth processes,, doi:10.1038/322137a0.
Flóvenz, Ó. G., and K. Gunnarsson (1991), Seismic crustal structure in Iceland and surrounding area, *Tectonophysics*, *189*(1-4), 1–17, doi:10.1016/0040-1951(91)90483-9.
Furumoto, A. S., D. M. Hussong, J. F. Campbell, G. H. Sutton, A. Malahoff, J. C. Rose, and G. P. Woollard (1970), Crustal and upper mantle structure of the Solomon Islands as revealed by seismic refraction survey of November-December 1966, *Pacific Science*, *24*, 315–332.
Furumoto, A. S., J. P. Webb, M. E. Odegard, and D. M. Hussong (1976), Seismic studies on the Ontong Java plateau, 1970, *Tectonophysics*, *34*(1-2), 71–90, doi:10.1016/0040-1951(76)90177-3.
Gladczenko, T. P., M. F. Coffin, and O. Eldholm (1997), Crustal structure of the Ontong Java Plateau: modeling of new gravity and existing seismic data, *Journal of Geophysical Research*, *102*, 22711–22729, doi:10.1029/97JB01636.
Granot, R., J. Dyment, and Y. Gallet (2012), Geomagnetic field variability during the Cretaceous Normal Superchron, *Nature Geoscience*, *5*(3), 220–223, doi:10.1038/ngeo1404.
Gutscher, M. A., J. L. Olivet, D. Aslanian, J. P. Eissen, and R. Maury (1999), The "lost Inca Plateau": cause of flat subduction beneath Peru? *Earth and Planetary Science Letters*, *171*(3), 335–341.
Hochmuth, K., K. Gohl, G. Uenzelmann-Neben, and R. Werner (in review), Multiphase magmatic and tectonic evolution - evidence from the crustal structure of the Manihiki Plateau, western Pacific, *Journal of Geophysical Research*.
Hoernle, K., F. Hauff, P. van den Bogaard, W. Reinhard, N. Mortimer, J. Geldmacher, D. Garbe-Schönberg, and B. Davy (2010), Age and geochemistry of volcanic rocks from the Hikurangi and Manihiki oceanic Plateaus, *Geochimica et Cosmochimica Acta*, *74*, 1–24, doi:10.1016/j.gca.2010.09.030.
Ingle, S. P., J. J. Mahoney, H. Sato, M. F. Coffin, J.-I. Kimura, N. Hirano, and M. Nakanishi (2007), Depleted mantle wedge and sediment fingerprint in unusual basalts from the Manihiki Plateau, central Pacific Ocean, *Geology*, *35*(7), 595, doi:10.1130/G23741A.1.
Inoue, H., M. F. Coffin, Y. Nakamura, K. Mochizuki, and L. W. Kroenke (2008), Intrabasement reflections of the Ontong Java Plateau: Implications for plateau construction, *Geochemistry Geophysics Geosystems*, *9*(4), 1–19, doi:10.1029/2007GC001780.
Ishikawa, A., E. Nakamura, and J. J. Mahoney (2005), Jurassic oceanic lithosphere beneath the southern Ontong Java Plateau: Evidence from xenoliths in alnöite, Malaita, Solomon Islands, *Geology*.
Ishikawa, A., T. Kuritani, A. Makishima, and E. Nakamura (2007), Ancient recycled crust beneath the Ontong Java Plateau: Isotopic evidence from the garnet clinopyroxenite xenoliths, Malaita, Solomon Islands, *Earth and Planetary Science Letters*, *259*(1-2), 134–148, doi:10.1016/j.epsl.2007.04.034.
Ito, G., J. Lin, and D. Graham (2003), Observational and theoretical studies of the dynamics of mantle plume–mid-ocean ridge interaction, *Reviews of Geophysics*, doi:10.1029/2002RG000117.
Karlstrom, L., and M. A. Richards (2011), On the evolution of large ultramafic magma chambers and timescales for flood basalt eruptions, *Journal of Geophysical*

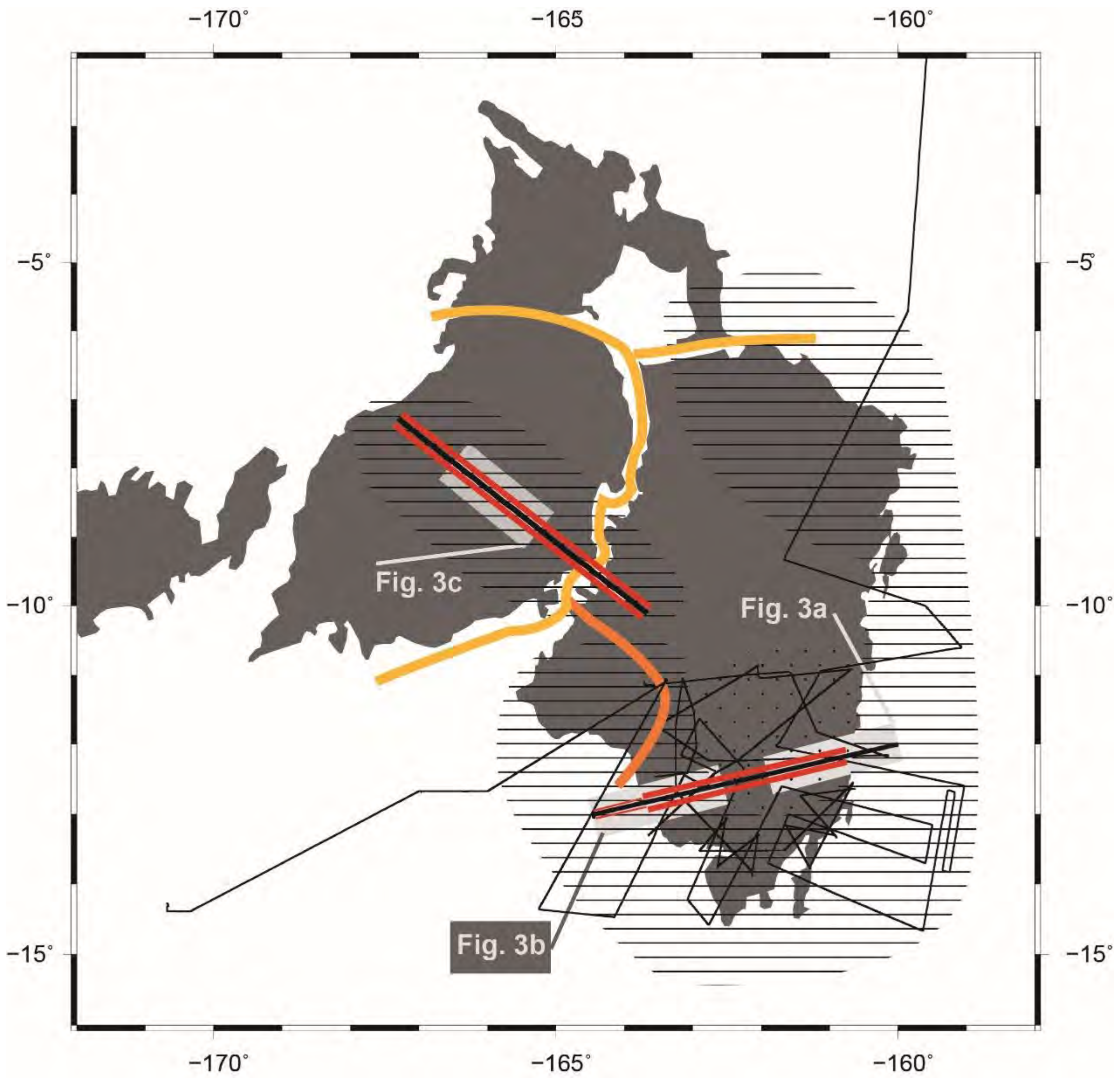
Research, *116*(B8), B08216, doi:10.1029/2010JB008159.
Klosko, E. R., R. M. Russo, E. A. Okal, and W. P. Richardson (2001), Evidence for a rheologically strong chemical mantle root beneath the Ontong–Java Plateau, *Earth and Planetary Science Letters*, *186*(3), 347–361, doi:10.1016/S0012-821X(01)00235-7.
Kokfelt, T. F., C. Lundstrom, K. Hoernle, and F. Hauff (2005), Plume–ridge interaction studied at the Galápagos spreading center: Evidence from 226 Ra–230 Th–238 U and 231 Pa–235 U isotopic disequilibria, *Earth and Planetary Science Letters*, doi:10.1016/j.epsl.2005.02.031.
Koppers, A. A. P., and H. Staudigel (2005), Asynchronous Bends in Pacific Seamount Trails: A Case for Extensional Volcanism? *SCIENCE-NEW YORK THEN WASHINGTON-*, *307*(5711), 904–907, doi:10.1126/science.1107260.
Larson, R. L. (1997), Superplumes and ridge interactions between Ontong Java and Manihiki Plateaus and the Nova-Canton Trough, *Geology*, *25*(9), 779–782, doi:10.1130/0091-7613(1997)025<0779:SARIBO>2.3.CO;2.
Larson, R. L., and C. Chase (1972), Late Mesozoic Evolution of the Western Pacific Ocean, *Geological Society of America Bulletin*, *83*(12), 3627, doi:10.1130/0016-7606(1972)83[3627:LMEOTW]2.0.CO;2.
Larson, R. L., R. A. Pockalny, R. F. Viso, E. Erba, L. J. Abrams, B. Luyendyk, J. Stock, and R. Clayton (2002), Mid-Cretaceous tectonic evolution of the Tongareva triple junction in the southwestern Pacific Basin, *Geology*, *30*, 67–70, doi:10.1130/0091-7613(2002)030<0067:MCTEOT>2.0.CO;2.
Liu, L., M. Gurnis, M. Seton, J. Saleeby, R. D. Müller, and J. M. Jackson (2010), The role of oceanic plateau subduction in the Laramide orogeny, *Nature Geoscience*, *3*(5), 353–357, doi:10.1038/ngeo829.
Mahoney, J. J., M. Storey, R. A. Duncan, K. J. Spencer, and M. Pringle (1993), Geochemistry and geochronology of Leg 130 basement lavas: nature and origin of the Ontong Java Plateau, *Proceedings of the Ocean Drilling Programm Scientific Results*, *130*, 3–22.
Mann, P., and A. Taira (2004), Global tectonic significance of the Solomon Islands and Ontong Java Plateau convergent zone, *Tectonophysics*, *389*(3), 137–190.
Matthews, K. J., M. Seton, and R. D. Müller (2012), Earth and Planetary Science Letters, *Earth and Planetary Science Letters*, *355-356*(c), 283–298, doi:10.1016/j.epsl.2012.08.023.
Maus, S., U. Barckhausen, H. Berkenbosch, N. Bournas, J. Brozena, V. Childers, F. Dostaler, J. D. Fairhead, C. Finn, and R. Von Frese (2009), EMAG2: A 2–arc min resolution Earth Magnetic Anomaly Grid compiled from satellite, airborne, and marine magnetic measurements, *Geochemistry Geophysics Geosystems*, *10*(8), doi:10.1029/,DanaInfo=dx.doi.org+2009GC002471.
Mittelstaedt, E., and G. Ito (2005), Plume-ridge interaction, lithospheric stresses, and the origin of near-ridge volcanic lineaments, *Geochemistry Geophysics Geosystems*, doi:10.1029/2004GC000860.
Miura, S., K. Suyehiro, M. Shinohara, N. Takahashi, E. Araki, and A. Taira (2004), Seismological structure and implications of collision between the Ontong Java Plateau and Solomon Island Arc from ocean bottom seismometer–airgun data, *Tectonophysics*, *389*(3-4), 191–220, doi:10.1016/j.tecto.2003.09.029.
Mochizuki, K., M. F. Coffin, O. Eldholm, and A. Taira (2005), Massive Early Cretaceous volcanic activity in the Nauru Basin related to emplacement of the Ontong Java Plateau, *Geochemistry Geophysics Geosystems*, *6*(10), 1–19, doi:10.1029/2004GC000867.
Mortimer, N., K. Hoernle, F. Hauff, J. M. Palin, W. J. Dunlap, R. Werner, and K. Faure (2006), New constraints on the age and evolution of the Wishbone Ridge, southwest Pacific Cretaceous microplates, and Zealandia–West Antarctica breakup, *….*
Musgrave, R. J. (2013), Evidence for Late Eocene emplacement of the Malaita

Terrane, Solomon Islands: Implications for an even larger Ontong Java Nui oceanic plateau, *J Geophys Res-Sol Ea*, doi:2169-9313/13/10.1002/jgrb.50153.
Musgrave, R. J. (1990), Paleomagnetism and tectonics of Malaita, Solomon Islands, *Tectonics*, *9*(4), 735–759.
Müller, R. D., M. Sdrolias, C. Gaina, and W. R. Roest (2008), Age, spreading rates, and spreading asymmetry of the world's ocean crust, *Geochemistry Geophysics Geosystems*, *9*(4), n/a–n/a, doi:10.1029/2007GC001743.
Nakanishi, M., K. Tamaki, and K. Kobayashi (1992), A new Mesozoic isochron chart of the northwestern Pacific Ocean: Paleomagnetic and tectonic implications, *Geophysical Research Letters*, *19*(7), 693–696, doi:10.1029/92GL00022.
Nakanishi, M., Y. Nakamura, M. F. Coffin, K. A. Hoernle, and R. Werner (2015), Topographic expression of the Danger Islands Troughs and implications for the tectonic evolution of the Manihiki Plateau, western equatorial Pacific Ocean, *GSA Special Papers*, *511*, doi:10.1130/2015.2511(11).
Pietsch, R., and G. Uenzelmann-Neben (in prep.), Can hotspot volcanism explain the Tertiary tectonic and volcanic reactivation of the Manihiki Plateau in the central Pacific?
Pietsch, R., and G. Uenzelmann-Neben (in review), The Manihiki Plateau – a multistage volcanic emplacement history, 1–35, *Geochemistry, Geophysics, Geosystems*
Pockalny, R. A., L. J. Abrams, and R. F. Viso (2002), Bathymetry and gravity data across a mid-Cretaceous triple junction trace in the southwest Pacific basin, *Geophysical Research Letters*, *29*(1), 1007, doi:10.1029/2001GL013517.
Reyners, M. (2013), The central role of the Hikurangi Plateau in the Cenozoic tectonics of New Zealand and the Southwest Pacific, *Earth and Planetary Science Letters*, *361*, 460–468, doi:10.1016/j.epsl.2012.11.010.
Richardson, W. P., E. A. Okal, and S. Van der Lee (2000), Rayleigh–wave tomography of the Ontong–Java Plateau, *Physics of the Earth and Planetary Interiors*, *118*(1), 29–51, doi:10.1016/S0031-9201(99)00122-3.
Ridley, V. A., and M. A. Richards (2010), Deep crustal structure beneath large igneous provinces and the petrologic evolution of flood basalts, *Geochemistry Geophysics Geosystems*, *11*(9), doi:10.1029/2009GC002935.
Riisager, P., J. Riisager, X. Zhao, and R. S. Coe (2003), Cretaceous geomagnetic paleointensities: Thellier experiments on Pillow lavas and Submarine basaltic glass from the Ontong Java Plateau, *Geochemistry Geophysics Geosystems*, *4*(12), doi:10.1029/2003GC000611.
Sager, W. W. (2006), Cretaceous paleomagnetic apparent polar wander path for the Pacific plate calculated from Deep Sea Drilling Project and Ocean Drilling Program basalt cores, *Physics of the Earth and Planetary Interiors*, *156*(3-4), 329–349, doi:10.1016/j.pepi.2005.09.014.
Sandwell, D. T., and W. H. Smith (1997), Marine gravity anomaly from Geosat and ERS 1 satellite altimetry, *Journal of Geophysical Research*, *102*(B5), 10039–10–054, doi:10.1029/96JB03223.
Seton, M. et al. (2012), Global continental and ocean basin reconstructions since 200Ma, *Earth-Science Reviews*, *113*(3-4), 212–270, doi:10.1016/j.earscirev.2012.03.002.
Sinton, J., R. Detrick, J. P. Canales, G. Ito, and M. Behn (2003), Morphology and segmentation of the western Galápagos Spreading Center, 90.5°–98°W: Plume-ridge interaction at an intermediate spreading ridge, *Geochemistry Geophysics Geosystems*, *4*(12), doi:10.1029/2003GC000609.
Sutherland, R., and C. Hollis (2001), Cretaceous demise of the Moa plate and strike-slip motion at the Gondwana margin, *Geology*, *29*(3), 279, doi:10.1130/0091-7613(2001)029<0279:CDOTMP>2.0.CO;2.
Taylor, B. (2006), The single largest oceanic plateau: Ontong Java-Manihiki-Hikurangi, *Earth and Planetary Science Letters*, *241*, 372–380,

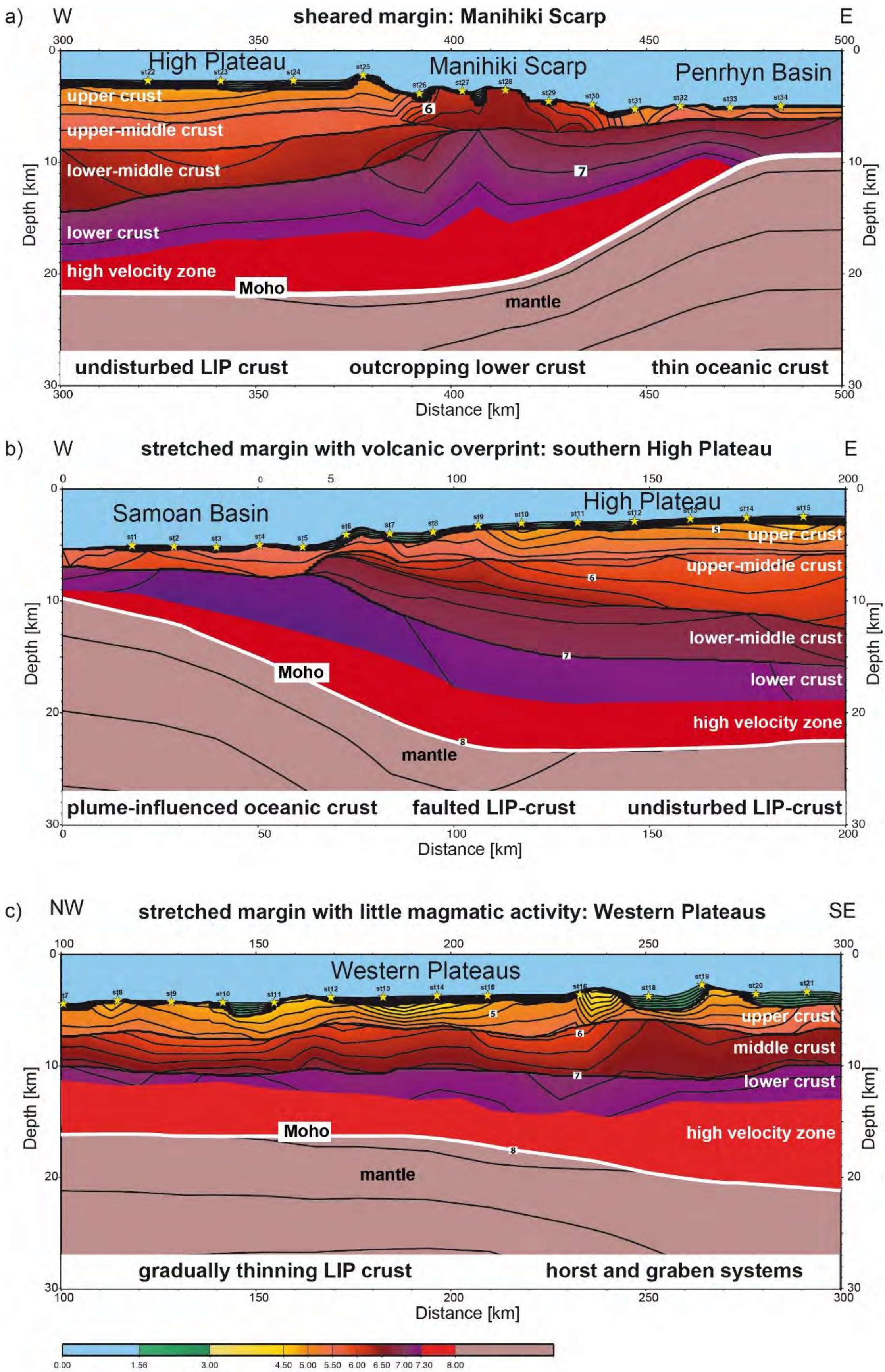
doi:10.1016/j.epsl.2005.11.049.
Thompson, P. M. E., P. D. Kempton, R. V. White, A. C. Kerr, J. Tarney, A. D. Saunders, J. G. Fitton, and A. McBirney (2004), Hf–Nd isotope constraints on the origin of the Cretaceous Caribbean plateau and its relationship to the Galápagos plume, *Earth and Planetary Science Letters*, *217*(1-2), 59–75, doi:10.1016/S0012-821X(03)00542-9.
Timm, C. et al. (2014), Subduction of the oceanic Hikurangi Plateau and its impact on the Kermadec arc, *Nature Communications*, *5*, 4923, doi:10.1038/ncomms5923.
Timm, C., K. Hoernle, W. Reinhard, F. Hauff, P. van den Bogaard, P. Michael, M. F. Coffin, and A. A. P. Koppers (2011), Age and geochemistry of the oceanic Manihiki Plateau, SW Pacific: New evidence for a plume origin, *Earth and Planetary Science Letters*, *304*, 135–146, doi:10.1016/j.epsl.2011.01.025.
Uenzelmann-Neben, G. (2012), The expedition of the research vessel“ Sonne” to the Manihiki Plateau in 2012 (So 224), *EPIC3Berichte zur Polar- und Meeresforschung = Reports on polar and marine research, Bremerhaven, Alfred Wegener Institute for Polar and Marine Research, 672, 111 p., ISSN: 1866-3192*.
Viso, R. F., R. L. Larson, and R. A. Pockalny (2005), Tectonic evolution of the Pacific–Phoenix–Farallon triple junction in the South Pacific Ocean, *Earth and Planetary Science Letters*, *233*(1-2), 179–194, doi:10.1016/j.epsl.2005.02.004.
Wessel, P., and L. W. Kroenke (2008), Pacific absolute plate motion since 145 Ma: An assessment of the fixed hot spot hypothesis, *Journal of Geophysical Research*, *113*(B6), B06101, doi:10.1029/2007JB005499.
Winterer, E. L., P. L. Lonsdale, J. L. Matthews, and B. R. Rosendahl (1974), Structure and acoustic stratigraphy of the Manihiki Plateau, *Deep Sea Research and Oceanographic Abstracts*, *21*(10), 793–813, doi:10.1029/JB084iB11p06003.
Worthington, T. J., R. Hekinian, P. Stoffers, T. Kuhn, and F. Hauff (2006), Osbourn Trough: Structure, geochemistry and implications of a mid-Cretaceous paleospreading ridge in the South Pacific, *Earth and Planetary Science Letters*, *245*(3-4), 685–701, doi:10.1016/j.epsl.2006.03.018.

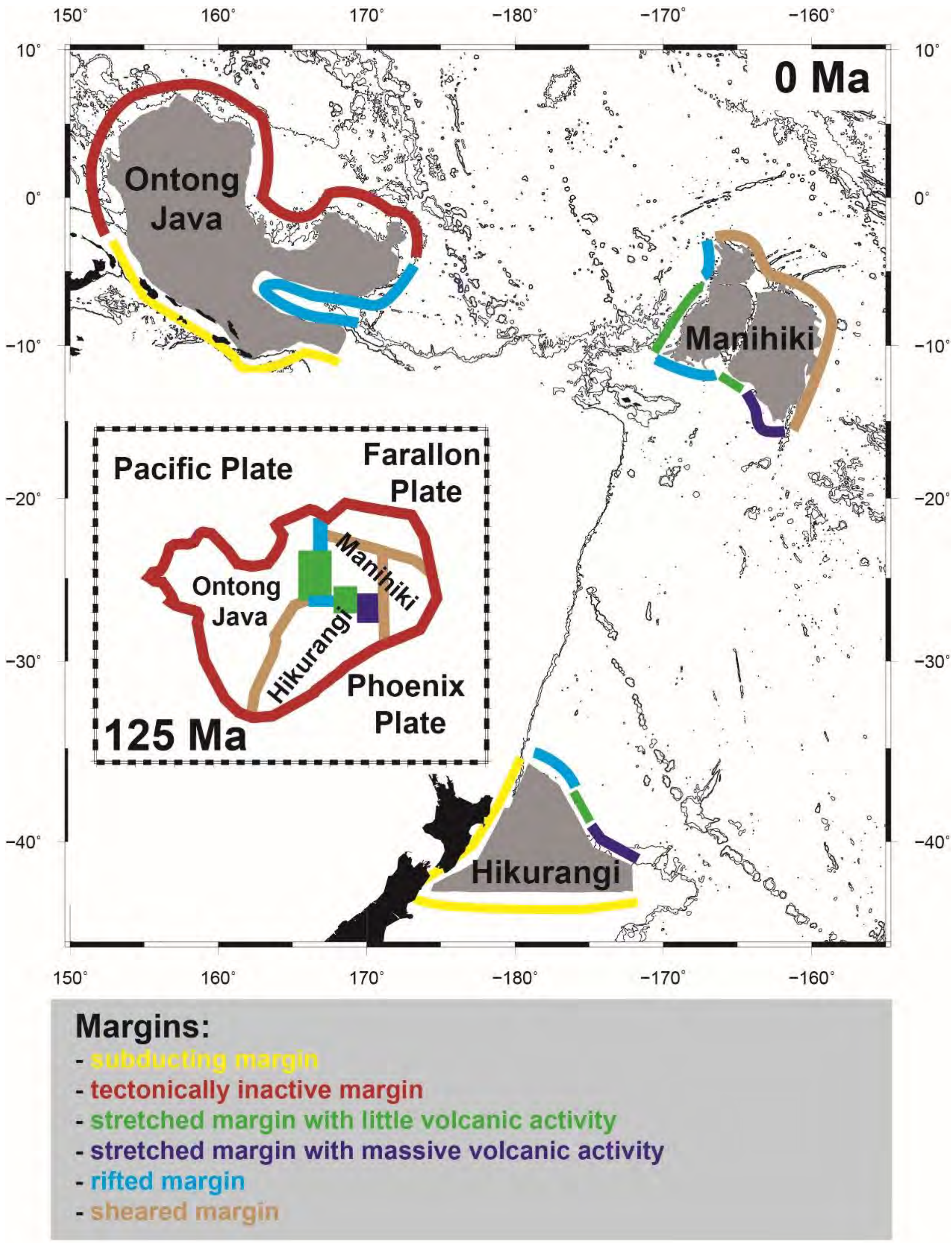


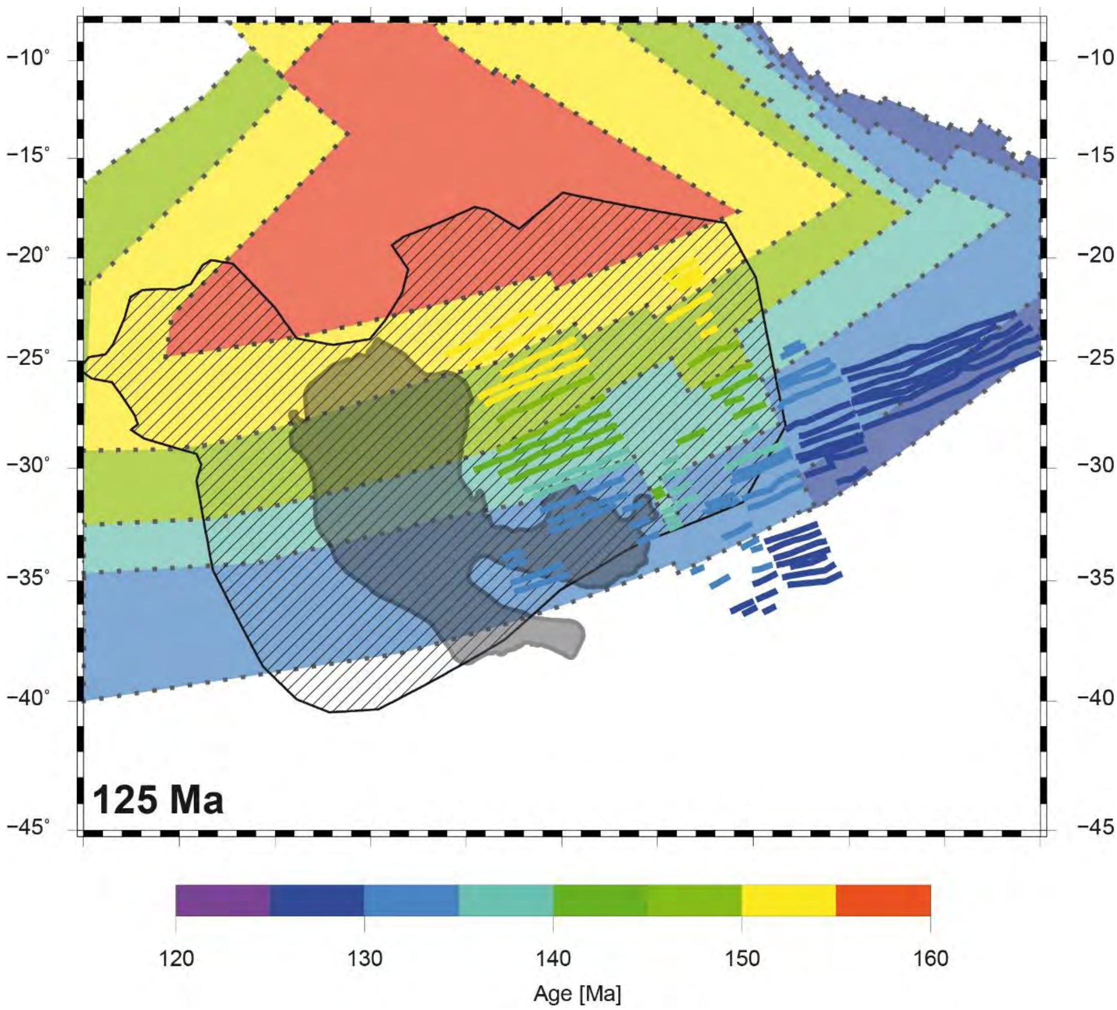
M25
M21
M21
M16
M10
M0
Phoenix FZ
Central Pacific FZ
Ontong Java Plateau
Nova Canton Trough
Clipperton FZ
Galapagos FZ
N.P.
W.P.
D.I.T.
H.P.
R. R.
C34
Marquesas FZ
M. Sc.
Tongareva Triple Junction
C34
Austral FZ
Osbourn Trough
West Wishbone Scarp
East Wishbone Scarp
C34
Hikurangi Plateau
Chatham Rise
sub-provinces of the Manihiki Plateau
H.P. High Plateau
W.P. Western Plateaus
N.P. North Plateau
R.R. Robbie Ridge
D.I.T. Danger Islands Troughs
M. Sc. Manihiki Scarp



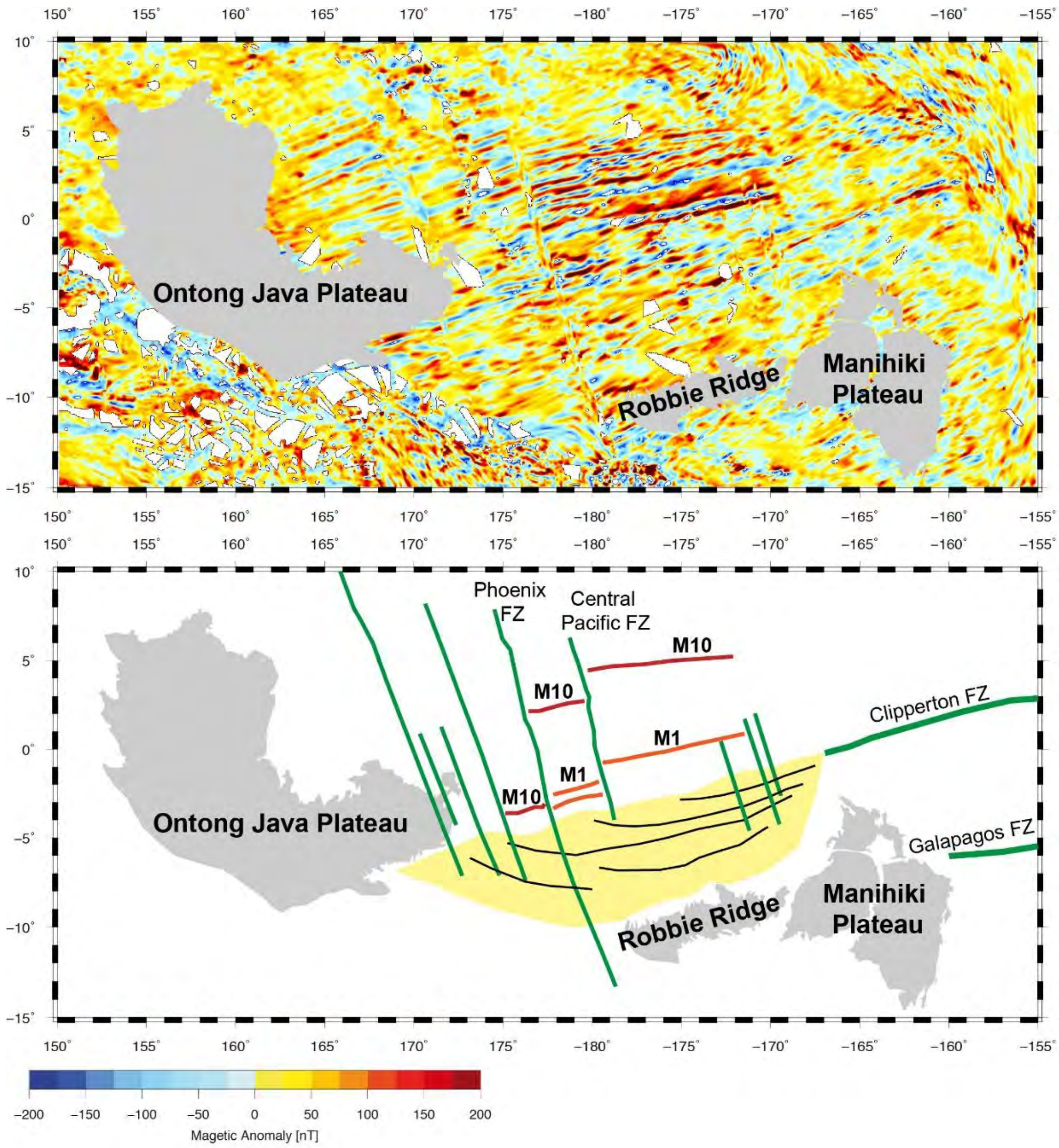
−170°
−165
−160°
−5°
−10°
−15°
Fig. 3c
Fig. 3a
Fig. 3b


a)
W
sheared margin: Manihiki Scarp
E
High Plateau
Manihiki Scarp
Penrhyn Basin
upper crust
upper-middle crust
lower-middle crust
lower crust
high velocity zone
Moho
mantle
undisturbed LIP crust
outcropping lower crust
thin oceanic crust
Depth [km]
Distance [km]
b)
W
stretched margin with volcanic overprint: southern High Plateau
E
Samoan Basin
High Plateau
upper crust
upper-middle crust
lower-middle crust
lower crust
high velocity zone
Moho
mantle
plume-influenced oceanic crust
faulted LIP-crust
undisturbed LIP-crust
Depth [km]
Distance [km]
c)
NW
stretched margin with little magmatic activity: Western Plateaus
SE
Western Plateaus
upper crust
middle crust
lower crust
high velocity zone
Moho
mantle
gradually thinning LIP crust
horst and graben systems
Depth [km]
Distance [km]
0.00
1.56
3.00
4.50
5.00
5.50
6.00
6.50
7.00
7.30
8.00

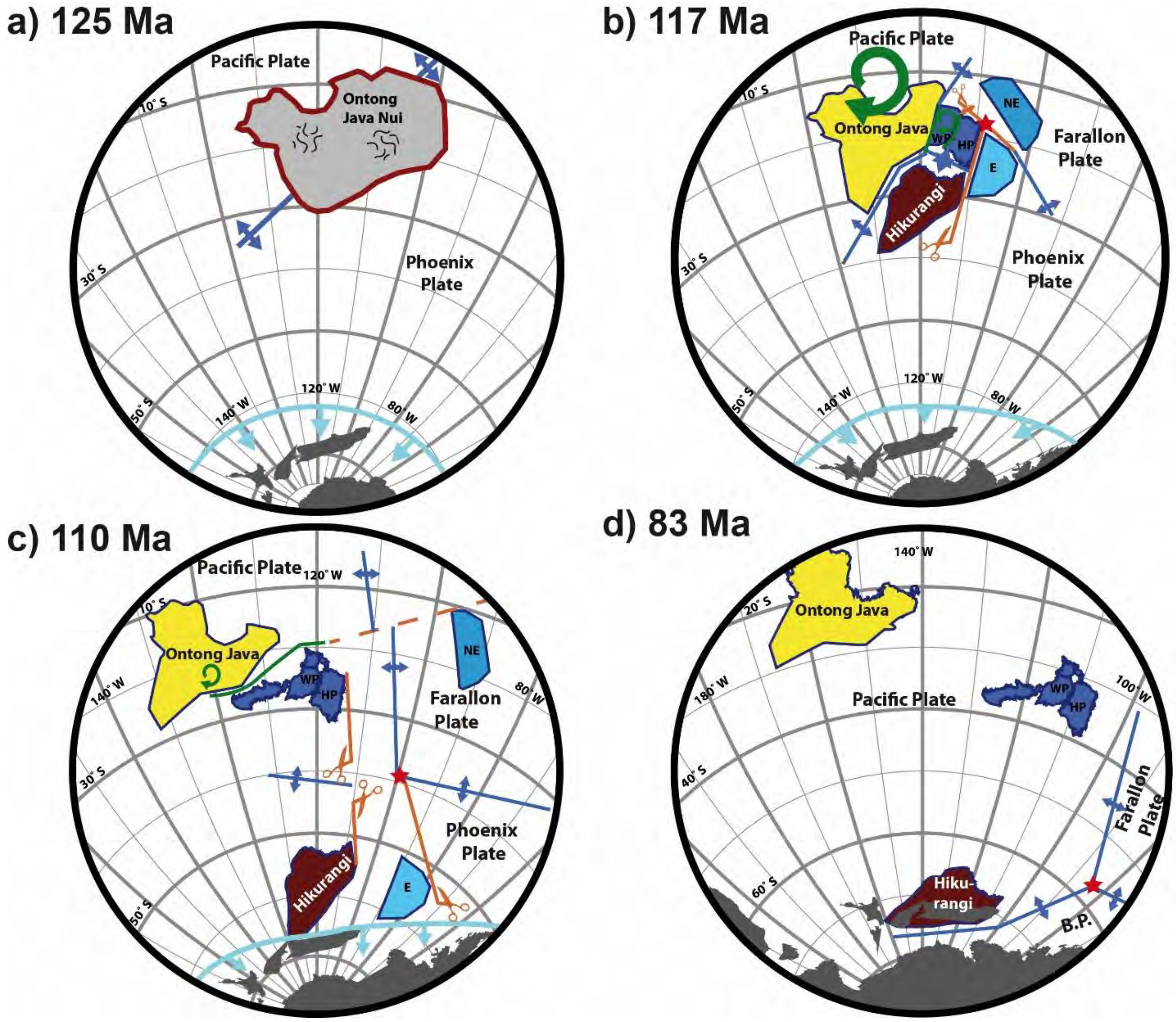

0 Ma
Ontong Java
Manihiki
Hikurangi
Pacific Plate
Farallon Plate
Phoenix Plate
125 Ma
Margins:
- subducting margin
- tectonically inactive margin
- stretched margin with little volcanic activity
- stretched margin with massive volcanic activity
- rifted margin
- sheared margin



−10°
−15°
−20°
−25°
−30°
−35°
−40°
−45°
−10
−15
−20
−25
−30
−35
−40
−45
125 Ma
120
130
140
150
160
Age [Ma]



Ontong Java Plateau
Robbie Ridge
Manihiki
Plateau
Phoenix
FZ
Central
Pacific FZ
M10
M10
M1
M1
M10
Clipperton FZ
Galapagos FZ
Magetic Anomaly [nT]



a) 125 Ma
Pacific Plate
Ontong Java Nui
10° S
30° S
50° S
Phoenix Plate
140° W
120° W
80° W
b) 117 Ma
Pacific Plate
Ontong Java
WP
HP
NE
E
Farallon Plate
Hikurangi
Phoenix Plate
10° S
30° S
50° S
140° W
120° W
80° W
c) 110 Ma
Pacific Plate
120° W
10° S
Ontong Java
WP
HP
NE
Farallon Plate
140° W
80° W
30° S
Phoenix Plate
Hikurangi
E
50° S
d) 83 Ma
140° W
20° S
Ontong Java
180° W
WP
HP
100° W
Pacific Plate
40° S
Farallon Plate
60° S
Hiku-
rangi
B.P.

PIETSCH, R., UENZELMANN-NEBEN, G. (IN REVISION): THE MANIHIKI PLATEAU – A MULTISTAGE VOLCANIC EMPLACEMENT HISTORY. GEOCHEMISTRY, GEOPHYSICS, GEOSYSTEMS - ANLAGE 4

The Manihiki Plateau – a multistage volcanic emplacement history

R. Pietsch*[1], G. Uenzelmann-Neben[1]
[1]Alfred Wegener Institute Helmholtz Center for Polar and Marine Research, Postfach 120161, 27515 Bremerhaven, Germany
*Corresponding author: Ricarda Pietsch, Email: ricarda.pietsch@awi.de, Tel.: +49 (0)471 4831 1206

Abstract
The emplacement history of the Manihiki Plateau, a Large Igneous Province, so far is poorly understood. We show evidence for multistage magmatic emplacement periods using new, high resolution seismic reflection data covering the High Plateau, which forms part of the Manihiki Plateau. Improved data quality allows for the identification of an intra-basalt reflection (IB1) sequence, interpreted to have formed prior to the main emplacement phase during the early Cretaceous (>125 Ma). It could reveal the onset of excessive extrusive volcanism associated with a first plateau formation within existing oceanic crust. Extrusion centers associated to the main eruption period (~125-110) and a later stage volcanic period (~90-65 Ma) can clearly be identified and distinguished. Extrusion centers formed within the late stage volcanic active period (ending ~65 Ma) are mainly concentrated along the margins of the High Plateau, suggesting that sources shifted from those being related to the initial emplacement of the High Plateau to tectonic induced volcanism at its margins. At the Manihiki Scarp, break-up characteristics are identified which differ from the ones observed along the south-western margin. Intrusion centers occur in the south-west-central part of the High Plateau which suggest a tectonic and volcanic reactivation during the Cenozoic.

Keywords
Reflection seismic, Manihiki Plateau, Cretaceous active period, Large Igneous Provinces, volcanic extrusion centers

Introduction

Large Igneous Provinces (LIPs) have been an important focus of recent studies due to their great impact on the earth's climate and geodynamic systems (Ganino and Arndt, 2009; Larson and Erba, 1999) as a consequence of massive volcanic outpouring during igneous pulses of short duration. They are usually associated with a rising head of a mantle plume (Bryan and Ernst, 2008; Coffin and Eldholm, 1994). In the Pacific Ocean, these flood basalt provinces, such as the Shatsky Rise, the Ontong-Java, Hikurangi and Manihiki Plateaus, formed during the late Jurassic- early Cretaceous (Nakanishi et al., 1999) and the early Cretaceous respectively (Hoernle et al., 2010; Mahoney et al., 1993; The Shipboard Scientific Party, 1976; Timm et al., 2011). Several tectonic reconstruction scenarios exist; however, a comprehensive emplacement model remains unclear.
The Manihiki Plateau, which is located in the western-central Pacific, has a yet to date unresolved multistage magmatic evolution history. Based on borehole data of DSDP Leg 33 Site 317, the plateau is supposed to have formed in a shallow water environment during the early Cretaceous (Jackson et al., 1976; Jenkyns, 1976). Different emplacement hypotheses exist, for example continental fragmentation (Heezen et al., 1965), impact related volcanism (Ingle and Coffin, 2004), formation at the Phoenix-Farallon-Pacific Plates triple-junction (Larson et al., 2002) and its combination with raising plumeheads (Larson, 1997). The most

agreed upon emplacement scenario suggests a formation of the Manihiki Plateau together with the Ontong-Java and Hikurangi Plateaus in a large single plateau and a direct break-off after ~115 Ma during the Cretaceous magnetic quiet period (Chandler et al., 2012; Downey et al., 2007; Taylor, 2006; Worthington et al., 2006). The Hikurangi Plateau, probably located at the west, south-west to south-east margins of the High Plateau (HP) rifted southward at the Osbourn spreading center (Downey et al., 2007). A missing piece of the composite LIP was supposedly located at the Manihiki Scarp in the east, likely rifted to the north-east and subducted beneath the Farallon Plate (Larson et al., 2002). The occurrence of second stage volcanism at the margins of the HP could have been triggered by these tectonic processes. Tectonic plate reconstruction studies (Chandler et al., 2012; Downey et al., 2007) suggest that the Ontong-Java Plateau was connected to the Manihiki Plateau at the Western Plateau (Winterer et al., 1974) and drifted towards the east.

The main objective of this study is to resolve the structure and evolution of the Manihiki Plateau by focusing on the questions of emplacement and tectonic related volcanism. Therefore, we use multi-channel, high resolution seismic reflection data gathered during cruise So224 of RV Sonne in 2012 (Uenzelmann-Neben, 2012) with an incorporation of borehole information of DSDP Leg 33 Site 317 (Ai et al., 2008; The Shipboard Scientific Party, 1976; Winterer et al., 1974). This allows us to confirm and extend the existent seismostratigraphic model. We aim to illuminate the complex multi-phase emplacement history by 1) a detailed analysis and interpretation of magmatic extrusion centers (EC) associated with the end of the tholeiitic volcanic main phase (ending ~120 Ma) and alkalic late stage volcanism (ending ~65 Ma) (Timm et al., 2011), 2) a discussion of an additional intra-basalt reflection IB1 which has not been drilled, 3) interpretation of spatial distribution of extrusion and intrusion centers, and 4) investigation of tectonic related volcanism at the margins.

Data acquisition and processing

During RV Sonne cruise So224 30 high resolution seismic reflection profiles (totaling ~4200 line km) were collected, focusing on the HP with the aim to study the structure of sedimentary sequences and basement on its central part and its eastern, southern and western margins (Fig. 1) (Uenzelmann-Neben, 2012). For the pulse generation of the seismic reflection data, a cluster of four GI-guns (volume 45 in^3 + 105 in^3) were fired with a shot interval of 10 s (~25 m). For profile AWI-20120201, a cluster of eight G-Guns (total volume 4160 in^3) was used with a shot interval of 60 s. Data were recorded with a 3 km long, 240-channel streamer (Sercel SEAL^{TM}) at a sampling rate of 1 ms. A frequency range up to 250 Hz allowed for a maximum vertical resolution of about ~3 m at the ocean bottom.

Results

Seismostratigraphic model

A seismostratigraphic model for the HP was first presented by Winterer et al. (1974) based on single channel seismic reflection data, and this was extended by Shipboard Scientific Party (1976) and Ai et al. (2008), naming the reflections from top to bottom (recent to old) (Fig. 2). In the previous works the seismic reflections were correlated visually to lithology and age information from DSDP Leg 33 Site 317 (The Shipboard Scientific Party, 1976). We correlated the seismic stratigraphy of the profiles gathered during cruise So224 to the lithology of borehole DSDP Leg 33 Site 317 by generating a synthetic seismogram. The impedance contrasts are calculated based on rock physical properties (wet-bulk density and interval velocity measured at the cores) convolved with a Ricker-wavelet (5-250 Hz) with a central frequency of 65 Hz (Fig. 2).

As a detailed lithological classification has been provided by the Shipboard Scientific Party (The Shipboard Scientific Party, 1976), we present a summary of the correlated lithology and seismic stratigraphy in Fig. 2 and Table 1, showing four units, separated by unconformities U1 (early/middle Eocene) and U2 (Mastrichtian/Campanian), which correspond to reflections R5 and R6, respectively, and reflection R7. Units 1 and 2 comprise Quaternary to middle Eocene sediments. Herein the description of the seismic stratigraphy focuses on Units 3 and 4 which show an abrupt increase of interval velocities to 2.5-3.0 km/s in Unit 3 and up to 4.0-5.5 km/s in Unit 4. These units have previously been identified to either represent deposition of volcaniclastic sediments (Unit 3) or basaltic flows (Unit 4) (Jackson et al., 1976; The Shipboard Scientific Party, 1976) of two different volcanic eruption episodes (Rea and Vallier, 1983; Timm et al., 2011).

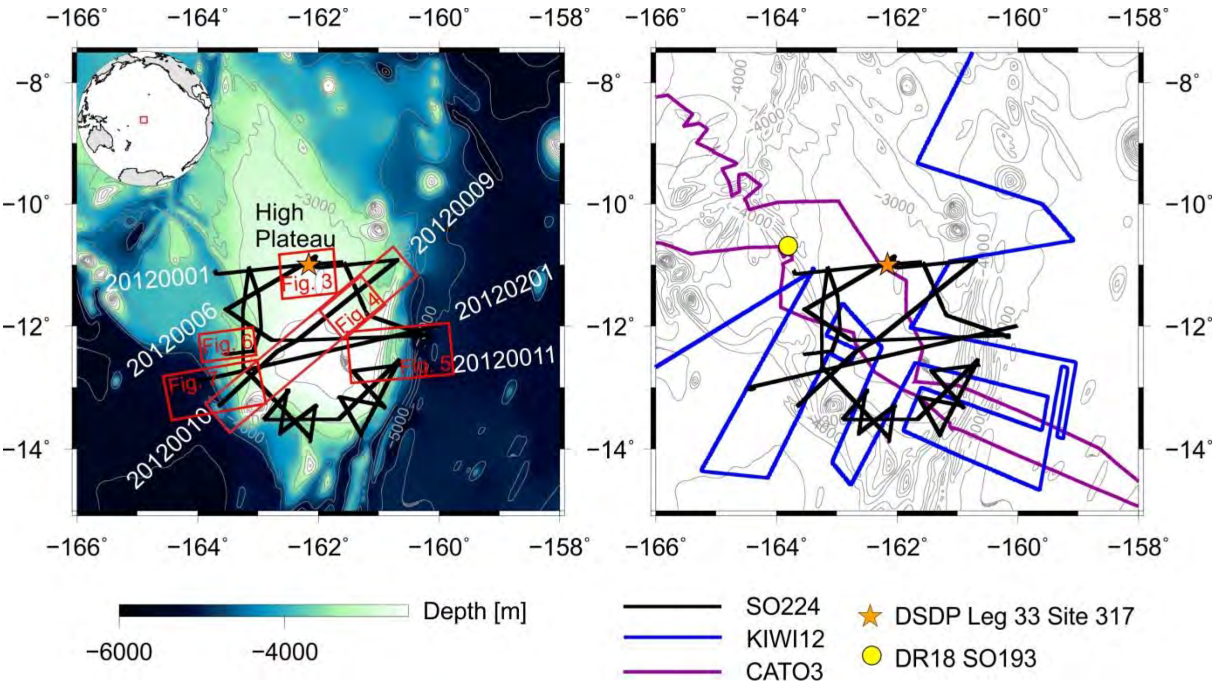


Fig. 1 The multi-channel seismic reflection profiles of cruise So224 (black lines) on top of the bathymetric map (GEBCO) expand over the central part of the High Plateau, part of the Manihiki Plateau, and its south-west and south-east margins. The orange star gives the location of DSDP Leg 33 Site 317 (The Shipboard Scientific Party, 1976), yellow circle of dredge sample of SO193 (Timm et al., 2011; Werner and Hauff, 2007) and single channel seismic reflection profiles of CATO3 (purple) and KIWI12 (blue) are shown for comparison.

The data processing steps included a set-up of geometry and a common-depth-point (CDP) sorting with interval spacing of 25 m, and 50 m for profile AWI-20120201. Afterwards, the CDP-sorted data were corrected for spherical divergence. A detailed velocity analysis was conducted every 50th CDP, including a coherency dependent set-up of velocity-depth profiles. A normal-move-out (NMO) correction, stacking and omega-x-migration in the time domain completed the basic data processing. For visualization the data were band-pass filtered within a range of 5-250 Hz. Additional velocity information for deep structures was made available via seismic wide angle/refraction data (Hochmuth et al., 2014).

We defined three additional reflections R5a, R6a and IB1 (Fig. 2 and Table 1). Reflection R5a is an internal reflection within Unit 2. It marks the change from an upper sequence (3.88-3.94 s TWT) characterized by internal, parallel reflections with strong amplitude to a more transparent sequence below (3.94-4.05 s TWT). The upper sequence comprises cherty chalk followed by a drilling gap of 129.5 m which extends to the transparent layer. Reflection R5a lies within this drilling gap. Using the borehole data of the upper and lower cores of DSDP Leg 33 Site 317, reflection R5a could represent an abundance of chert and could mark the onset of sedimentation after the volcanic emplacement period.

Reflection R6a (4.22 s TWT) is an internal reflection within Unit 3. It marks the change of an upper well-layered seismic sequence (4.05-4.22 s TWT) to a lower more transparent seismic sequence (4.22-4.31 s TWT). Interval velocities within this unit range from 2.8 km/s in the well-layered sequence to 2.0 km/s in the transparent sequence. In accordance to the lithology, the upper sequence comprised flow units of the later stage alkali volcanic phase (~110-65 Ma) (Hoernle et al., 2010; Timm et al., 2011). The transparent layer may well be the result of a redeposition of volcaniclastic deposits and sediments.

Reflection IB1 is an intra-basalt reflection within Unit 4. It is characterized by an increase of internal velocities up to 5-5.8 km/s. Its appearance is mostly continuous and shows lateral amplitude variations. It forms the top of a sub-parallel reflection sequence, which descends towards the south-western margin of the HP (Fig. 3).

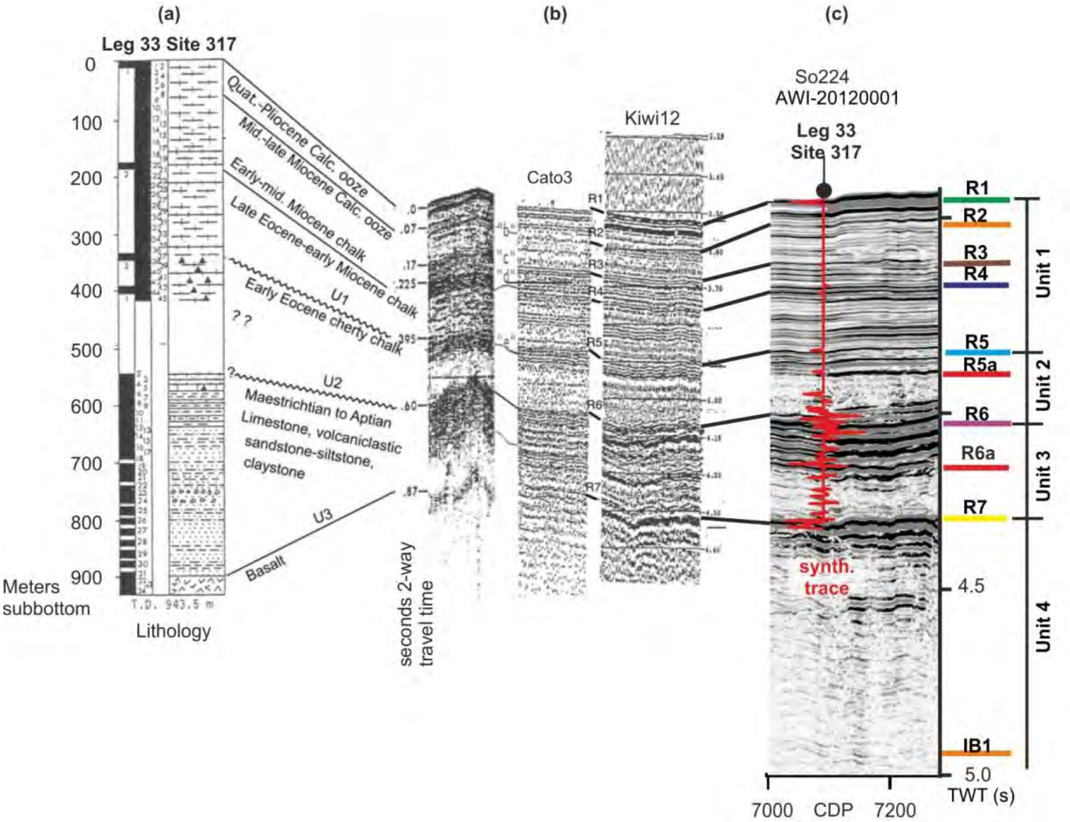


Fig. 2 Correlation of (a) Lithology of DSDP Leg 33 Site 317 (The Shipboard Scientific Party, 1976) to seismostratigraphy of (b) single channel seismic reflection data of cruise CATO3, KIWI12 (Ai et al., 2008) and (c) multi-channel high resolution data of cruise So224 (Uenzelmann-Neben, 2012). Reflections R1-R7 (recent-old) and newly defined reflections R5a and R6a are correlated to DSDP Site 317 lithology using a synthetic seismogram calculated of DSDP Leg 33 Site 317 wet-bulk density and velocity values. The additionally defined intra-basalt reflection IB1 lies below the penetration depth of borehole DSDP Leg 33 Site 317.

Identification of volcanic extrusion centers on the High Plateau

As the Manihiki Plateau was emplaced in an intraplate setting by a major amount of magma during short eruption pulses (Bryan and Ernst, 2008; Coffin and Eldholm, 1994), numerous volcanic features can be identified throughout the whole plateau. Extrusion centers are characterized by distinct seismic facies depending on the flow nature.

One of the most prominent volcanic structures is located on profile AWI-20120001 between CDPs 7200-9000 (Fig. 3a and b). Over a distance of 25 km reflection R7 (age ~120 Ma) domes up, develops a central, asymmetric main vent with an elevation of approx. ~1 km and slopes with a gradient of up to ~1.6°. Within the main vent multiple extrusion centers can be distinguished. Reflection R6 (age ~65 Ma) marks the end of a second, later volcanic period and parallels the topographic trend of reflection R7. Flows originate from the same extrusion centers as seen in R7 and additionally develop a second vent at CDP 7700 which forms a parasitic cone.

Further examples of extrusion centers exhibiting characteristic seismic patterns are seen on the south-western HP (Fig. 4). Reflection R7 marking the top of basalt has been chosen as the reference reflection because it can be clearly traced over the whole plateau. Below the top of basalt, an intra-basalt reflection sequence (IB1) can be identified and traced over long distances. Lateral amplitude variations occur, represented by strong amplitudes between CDPs 6000-9200 of Profile AWI-20120009 and CDPs 2200-3200 of Profile AWI-20120010 (Fig. 4b), as well as multiple zones of weaker amplitudes in between. The topography of IB1 is elevated towards the south-western HP and descends towards the flanks of the HP (Fig. 4c). Local elevation maxima are developed as seen, for example at CDPs 11200, 10400, 8800 and 6800 of profile AWI-20120009 (Fig. 4b).

	DSDP Leg 33 Site 317					Kiwi 12	So224			
Units	Lithology	Depth (mbsf)	Accum. rate m/m.y.	Age (Ma)	Refl. Name	Refl. Name	Refl. Name	Depth TWT (s)	Interval velocities (km/s)	Characteristics of reflector
Unit 1 (0-303.5m)	Calc. ooze	0		1.8		Q-R1	R1	3.46		Seafloor
	Calc ooze	58		5.3	"b"	lM-R2	R2	3.53	1.6-1.7	
	Ooze to Chalk	140	10	10	"c"	mM-R3	R3	3.63	1.7-1.9	
	Chalk	197		23	"d"	eM-R4	R4	3.68	2.0	
Unit 2 (303.5-647m)	Ooze to chalk limestone chert	**358** (/~377?)		45	"e"	eE-R5	R5	3.88		(358m late Eocene-mid Eocene unconformity corresponds to accumulation rate change)
	Drilling gap	377?	?	~55?			R5a	3.94		Eocene to Cretaceous chalks
Unit 3 (647-910m)	sandstone-siltstone, claystone, limestone, volcaniclastic	**576/** 592	1	65		IC-R6/U2	R6	4.05	2.5-3.5	Maestrichtian-Campanian boundary (576m) 595 hiatus
	Cretaceous limestone with minor chalk, claystone and chert	~602-645?								
		~645-680	87 ?	85? 110?			R6a	4.22	decrease	Santonian-Turonian boundary (83.5-89 Ma) Beginning of late stage volcanic active period
Unit 4 (> 910m)	Basalt	910	?	~120		eC-R7	R7	4.33	3.5-4.0	Top of basalt
				>125?			IB1	4.95	5-5.7	

mbsf: meters below sea floor

Table 1 Description of lithological Units of DSDP Leg 33 Site 317 and correlation to seismic reflection data of cruise KIWI12 and So224.

Seismic structures at the margins of the High Plateau

In contrast to the smooth topographic appearance of the basaltic flow (Unit 4) and the volcaniclastic deposition (Unit 3) in the central part of the HP, the margins of the HP show a different seismic reflection pattern; a rough topography. For the eastern margin, this transition is marked by an abrupt ridge-type seamount structure parallel to the Manihiki Scarp visible in profiles crossing the eastern margin, e.g. AWI-20120004, AWI-20120011 and AWI-20120201 (Fig. 5). For the ridge-type seamount (e.g. CDP 2200 in profile AWI-20120011) at least two different eruption phases can be distinguished (Fig. 5). The initial reflection IB1 is disrupted by reflection R7 (~120 Ma) and therefore terminates abruptly. Reflection R7 documents the first formation of the ridge-type seamount structures. Reflections R6a- R6 (~65 Ma) mark a second overprinting stage (CDP 2000-2400 in profile AWI-20120011). Across the Manihiki Scarp, which has previously been identified to form a transform fault (Viso et al., 2005), multiple seamount-type volcanic structures occur showing a strong and rough topography with steep sides and deep, buried troughs in between. A multi-stage formation of the volcanic seamounts can be identified, as subparallel flow units occur. However, dating and tracing of the reflections remains difficult due to the rugged topography and the missing sedimentary column (Fig. 5).

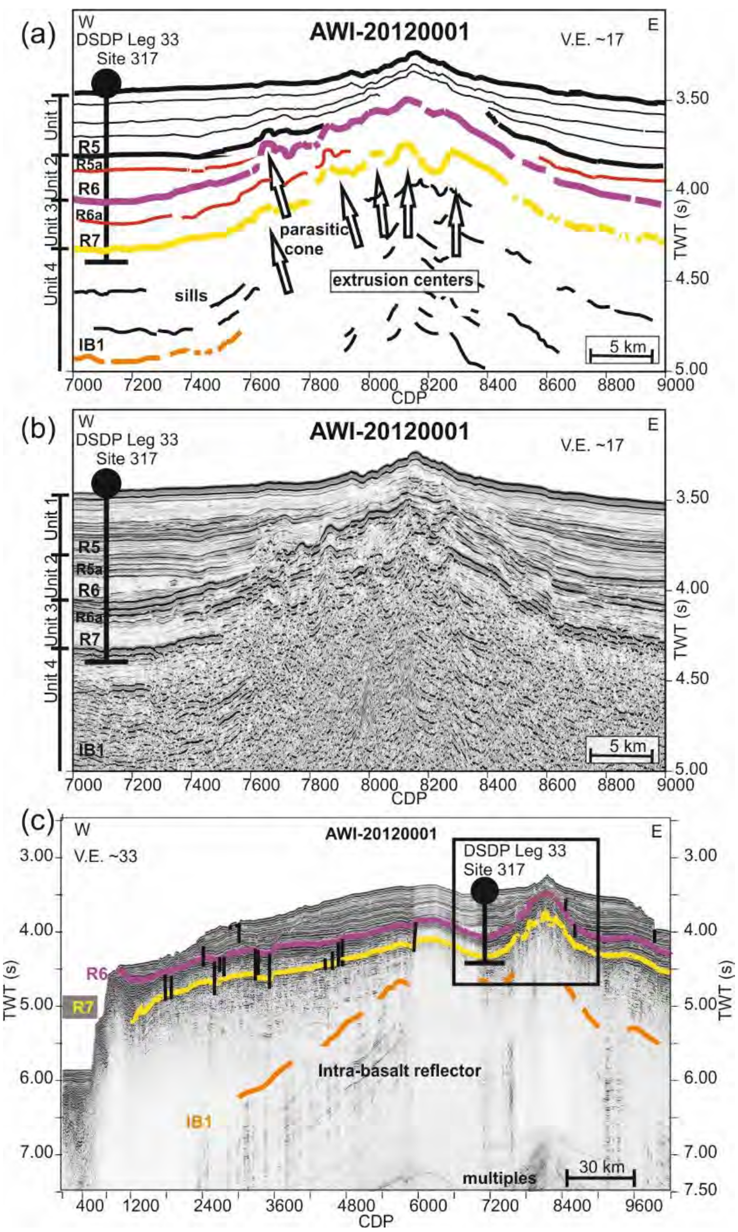


Fig. 3 The prominent complex volcanic extrusion structure (R7) located at CDP 7000-9000 in profile AWI-20120001 is fed by a deep reaching chimney and exhibits multiple vents. A parasitic cone occurs within the second stage volcanic period (R6) at the western flank of the

main vent right next to a deep depression in which DSDP Leg 33 Site 317 (The Shipboard Scientific Party, 1976) is located.

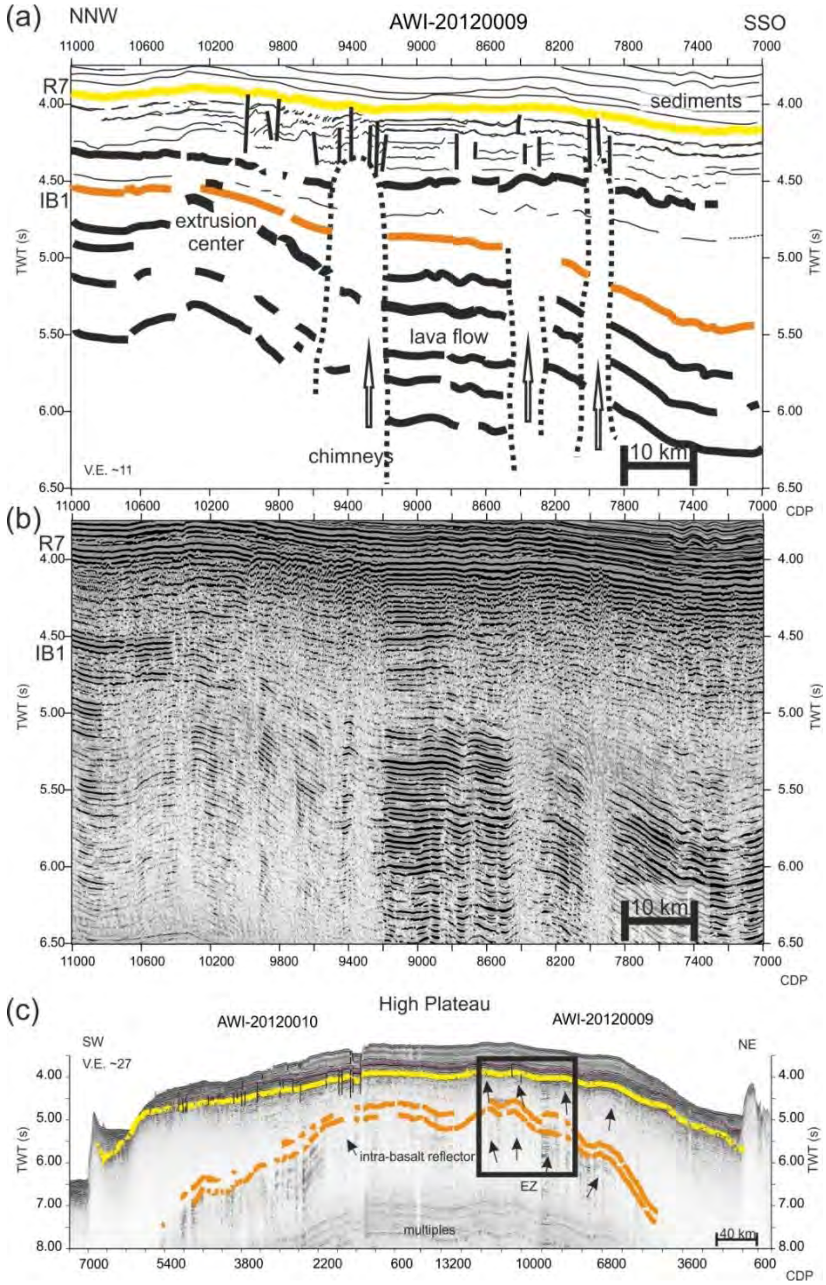


Fig. 4 Interpretation (a) and seismic section (b) of an intra-basalt (IB1) extrusion center developed along profiles AWI-20120009 and AWI-20120010 (c) crossing the High Plateau in a north-east south-west direction. Sub-parallel stratified reflections and strong amplitude variations within IB1 point towards subaerial lava flows. The upper reflections R7 and R6 are fed by deep reaching chimneys.

In the south-western and western parts of the HP the seismic image is characterized by numerous faults and disturbed zones. Some of these faults penetrate the complete sedimentary column from reflection R7 to R1, which can be seen e.g. in CDPs 1200-1400 of Profile AWI-20120010 (Fig. 4) where a graben structure is shown. The majority of the faults are offsetting Unit 3, pointing towards a second volcanic period on the HP.

In the south-western part of the HP, intrusion centers are visible which vary in age. Those magmatic intrusion centers show complex structures with multiple active periods and mostly developed domes with small topographic elevations. For example, between CDPs 1000-1300 of Profile AWI-20120028 at the western flank of the HP (Fig. 6), weakening and disruption of the reflection's amplitudes occur. A deep chimney can be identified feeding the volcanic flows of reflection R7 and R6. Reflection R7 has been disturbed and here is interpreted to represent a sill or dike structure, which feeds the upper sequence bounded by reflection R6. Disturbances of reflections are clearly visible throughout the whole sedimentary column, from reflection R7 to the ocean bottom and hint on a quite recent reactivation of the HP.

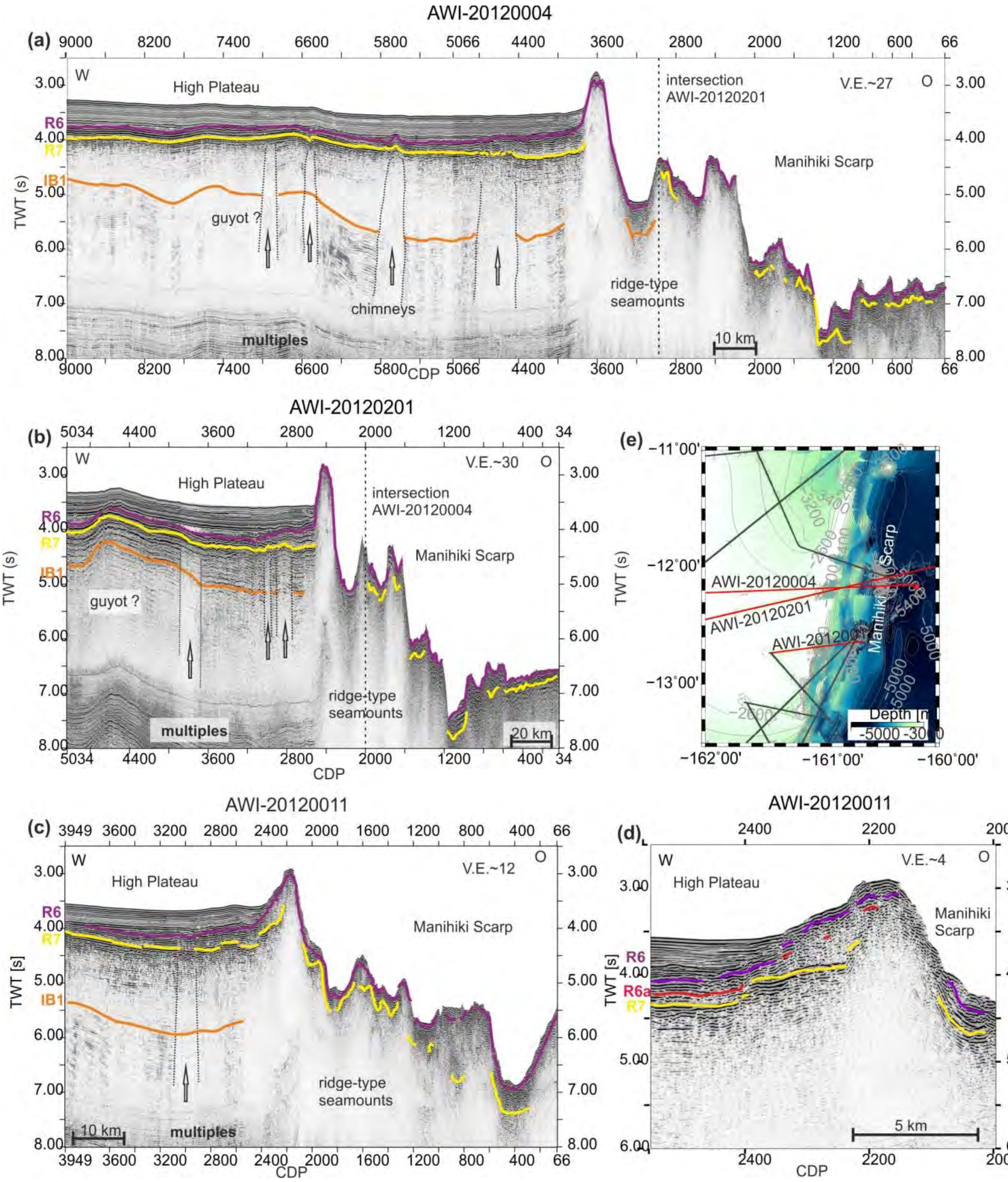


Fig. 5 Bathymetry (e) and seismic sections (a-d) across the transform-faulted Manihiki Scarp, the eastern margin of the HP, show ridge-type and seamount-type volcanic structures.

The southern and western margins of the HP flanks descend stepwise from the top of the plateau (~4 s TWT) to the deep sea ocean bottom at ~6.5 s TWT (Fig. 7). A deep trough, which develops into a graben structure in the northern part of the Suvarov Trough, separates a steep seamount-type volcanic structure (e.g. Fig. 7c, CDP 9000 on profile AWI-20120201). The eastern, deep-sea facing flank of the volcanic structure develops a normal fault structure with a vertical displacement of ~1.5 s TWT. Beyond, the seafloor appears smooth with local dome-like volcanic extrusion centers, e.g. CDP 9300 on profile AWI-20120201 (Fig. 7c). Reflections R7 and R6 can be traced across the south-western margin of the HP and show volcanic flow patterns and extrusion centers.

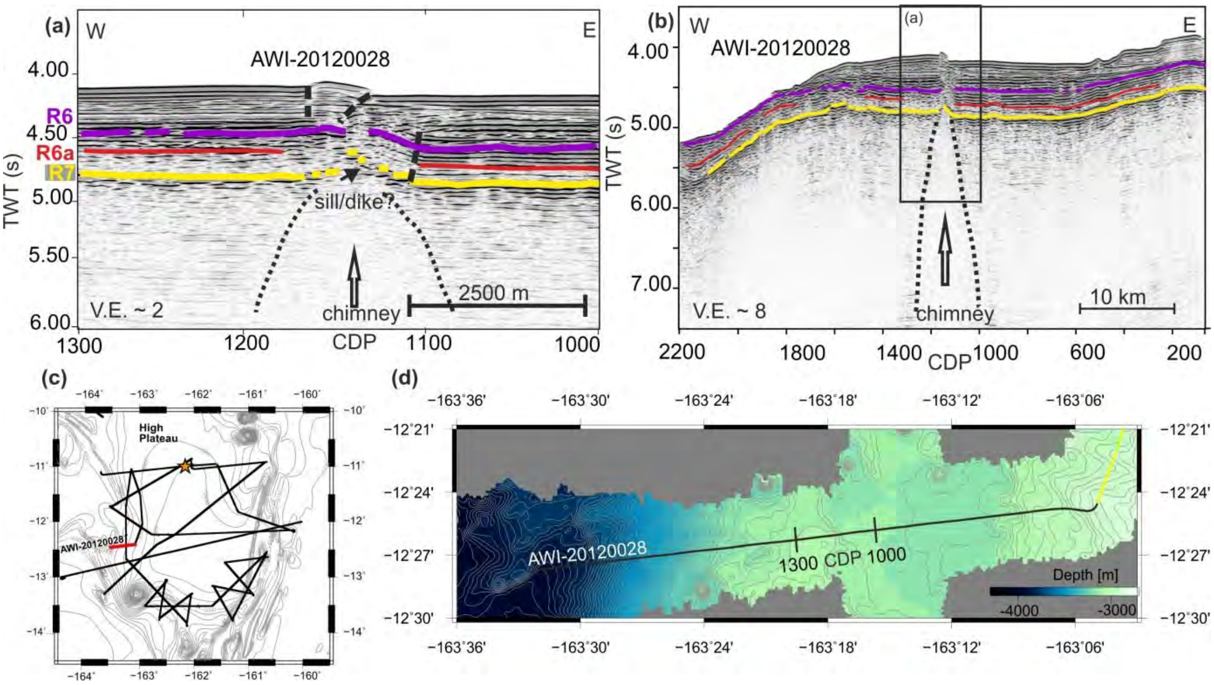


Fig. 6 Seismic interpretation section (a, b), location (c) and bathymetry (d) of a complex volcanic structure located on profile AWI-20120028 at the western part of the High Plateau. Fed by a deep reaching chimney, the top of basalt, reflection R7, is disrupted and exhibits a dyke structure to feed the upper reflection R6, which is associated with a later state volcanic period (The Shipboard Scientific Party, 1976). The disturbances of the upper units show that this extrusion center has recently been magmatically active.

Discussion

Intra-basalt reflection

An intra-basalt reflection R7 has been identified within the basement. Its existence has been proofed by analyzing and comparing traveltimes, reflection angles and velocity-depth profiles to seafloor and pegleg multiples. Additionally, the values of the depth migrated reflections (2300-2900 mbsf) agree well to the boundary between the upper crust and the upper-middle crust (2700 mbsf) as seen in seismic refraction data within the 20 km thick crust along profile AWI-20120201 (Hochmuth et al., 2014), where the P-wave velocity jump to values ≥6 km/s.

A few intra-basalt reflections also referred to as intra-basement reflections, have already been identified in seismic data on LIPs like the Ontong-Java Plateau (Inoue et al., 2008) or the Kerguelen Plateau (Rotstein et al., 1992; Schaming and Rotstein, 1990). Inoue et al. (2008) calculated synthetic seismograms and revealed that the appearance of strong amplitudes in seismic data can be explained by the alteration of pillow and massive lavas. Drilling data from the Kerguelen Plateau showed that intra-basement reflections represent subaerial lava flows (Frey et al., 2000). On the Manihiki Plateau, the seismic images of the intra-basalt reflection band (IB1) show evidence for volcanic extrusion origins, as displayed e.g. in Figure 4c at CDPs 11600 and 10400 in Profile AWI-20120009, where several prominent volcanic features are exhibited. Emanating from a center at CDP 10400 (Fig. 4a), subparallel reflection sequences are visible. We interpret these reflections to be volcanic flows emanating from the extrusion center in accordance to the interpretation of the seismic reflection studies of the comparable Agulhas LIP (Gohl and Uenzelmann-Neben, 2001; Uenzelmann-Neben et al., 1999) and Shatsky Rise (Sager et al., 2013). Lateral amplitude variations occur within IB1, showing strong amplitudes between CDP 8400-9200 and 7200-7900 and zones with weaker amplitudes in between. We interpret these sharply defined features with strong amplitudes to represent volcanic flows emerging from various vents. Comparing these flow units to the study conducted by Klarner and Klarner (2012), we conclude that they could have developed in a subaerial environment. Additionally, we interpret the zones of weak amplitudes (e.g. CDPs 9300, 8400 and 8000) to result from magmatic chimneys (Klarner and Klarner, 2012), which fed the later stage volcanism (R7). We suggest that the intra-basalt reflection sequence represents a volcanic flow unit deposited prior to R7.

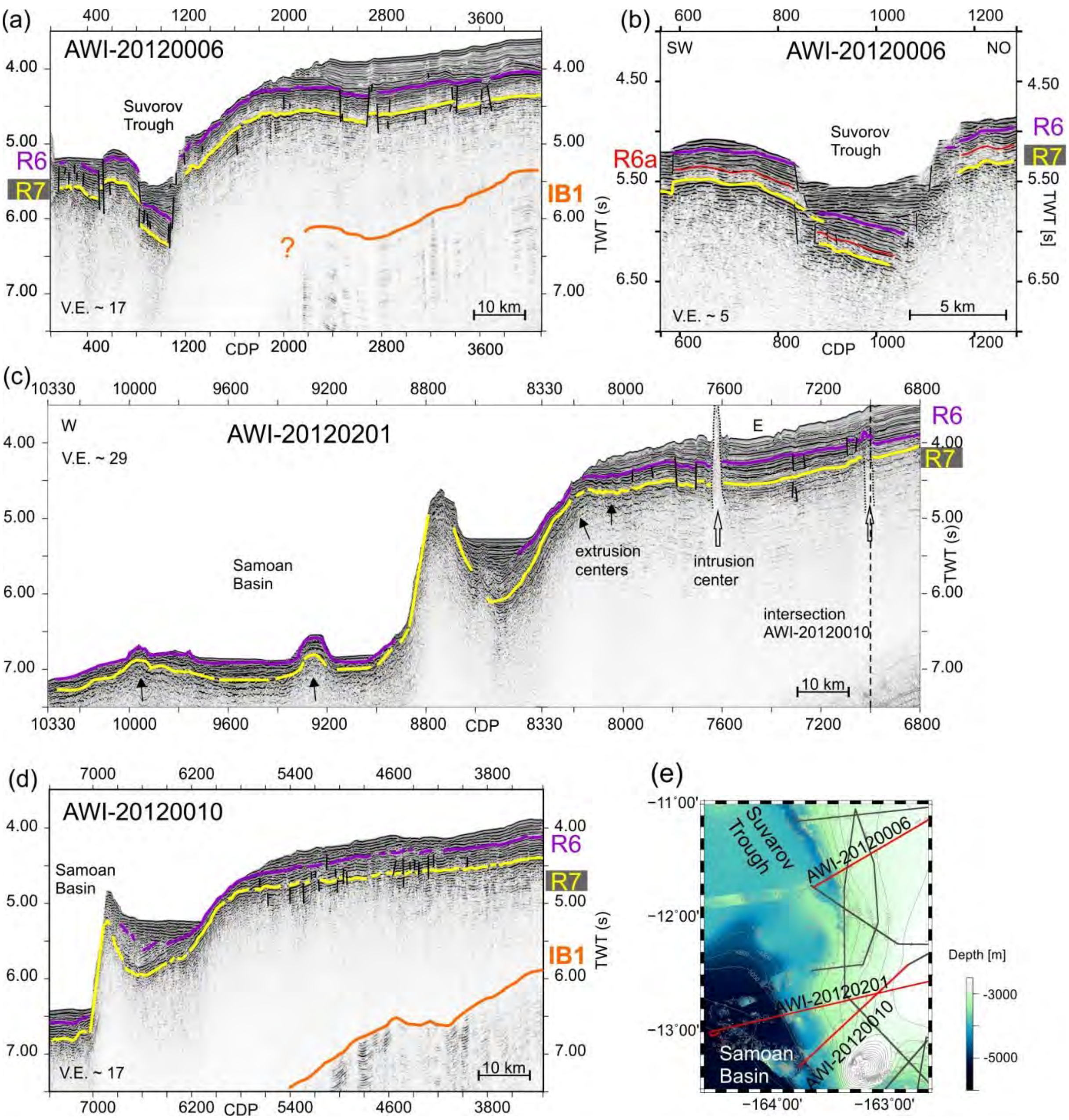


Fig. 7 Bathymetry (e) and seismic sections of the south-western and western margin of the HP as seen in profiles AWI-20120006 (a,b), AWI-20120201 (c) and AWI-20120010 (d), show tectonic disturbed zones in the north and terminates in a normal fault structure to the south.

Numerous volcanic extrusion centers are shown and the intra-basalt reflection IB1 descends towards the west. A recent intrusion is displayed in profile AWI-20120201 at CDP 7600.

The depth of reflection IB1 varies laterally across the HP (Fig. 4c and Fig. 8e) from ~7.5 s TWT at the margins to an elevation maximum of ~4.5 s TWT at the south-central part of the HP with a rather flat top e.g. CDPs 66-2200 of Profile AWI-20120010 (Fig. 4c). The radially rising topography hints on a massive volcanic extrusion center and could represent a shield volcano or/and a guyot structure. The inclination of reflection IB1 and the slowly rising slopes are similar to the seismic data of Shatsky Rise, another LIP (Sager et al., 2013) where a shield volcano (area of 3.1 x 105 km^2) of about twice the size of the MP has been identified. Thus, we conclude that it could form a shield volcano. The flat top could also display a guyot structure, as we suppose that reflection IB1 represents subaerial lava flows and erosion and subsidence of the Manihiki Plateau has occurred afterwards. Guyot structures on the HP are supported by the seismic refraction data collected parallel to profile AWI-20120201 Hochmuth et. al (2014) and have also been identified within the Hikurangi Plateau (Hoernle et al., 2004), which is assumed to have formed simultaneously with the Manihiki Plateau (Davy et al., 2008; Taylor, 2006). We suggest that this massive volcanic structure documents the nucleus of the HP's emplacement.

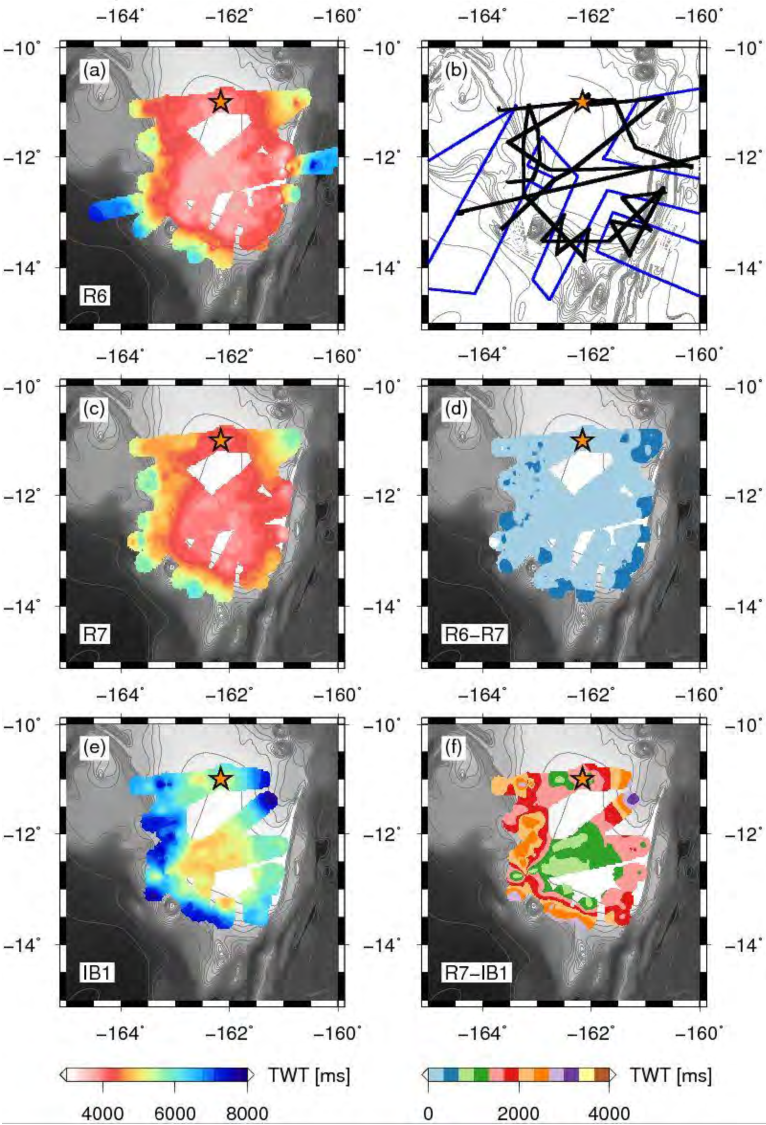


Fig. 8 Isopatch map of depth below seafloor of reflections R6 (a), R7 (c) and IB1 (e) and thickness of (d) Unit 3 (R6-R7) and (f) Unit 4 R7-IB1 are calculated from KIWI12 (blue) and

So224 (black) seismic reflection data. For all reflections IB1, R7 and R6 a summit height in the south-western part of the HP is visible and descending slopes to the north, east and south. Thicknesses of Unit 4 (f) show that the major emplacement occurred prior to R7 and a second smaller emplacement occurred within Unit 3 (d).

Confirmation of lithology and correlation of age for reflection IB1 remains difficult due to the lack of petrological data. As known from lithological and petrological studies of borehole DSDP Leg 33 Site 317, the tholeiitic basalt above IB1 shows an intra-plate morphology and was emplaced after the initial formation of the oceanic crust (~116-126 Ma) by a large amount of magma associated with a mantle plume head (Hoernle et al., 2010; Lanphere and Dalrymple, 1976; The Shipboard Scientific Party, 1976; Timm et al., 2011). This process led to an overprinting of the initial oceanic crust and sediments with smoother magmatic flow units. For the Ontong-Java Plateau intra-basalt reflections were considered to represent the alternation of pillow and massive lavas (Inoue et al., 2008). A joined emplacement of the Ontong-Java Plateau and the Manihiki Plateau (Taylor, 2006) thus supports our conclusion that reflection IB1 marks an earlier volcanic flow period.

Age estimation for IB1 is difficult because the 34 m of drilled basalt have been poorly recovered and age determination remained with wide uncertainties (Jackson et al., 1976; Lanphere and Dalrymple, 1976; The Shipboard Scientific Party, 1976). For the volcaniclastic deposition (Unit 3) bounded by reflections R6 and R7, a deposition rate (dr) of 18 m/My was calculated (The Shipboard Scientific Party, 1976). Below the top of basalt (Age_{R7} ~120 Ma) the interpolated average deposition rate could be as high as 87 m/my (The Shipboard Scientific Party, 1976). For an approximate age estimation, a measured P-wave velocity of v_i ~5.0 km/s for Unit 4 and a minimum travel time difference ΔTWT=0.4 between R7 and IB1 are used to calculate a minimum age of >132 Ma for the top of the reflection IB1 using $Age = Age_{R7} + \Delta\mathrm{TWT}/2 \cdot v_i/dr$. The reflection age, however, remains with big uncertainties as volcanic emplacement is periodic. Therefore, interfingering of sediments (Jackson et al., 1976) and alliterations of different flow types like pillow and massive lava (Inoue et al., 2008; Keszthelyi and Self, 1998) have to be taken into account. This would result in a lower deposition rate and an older age for IB1.

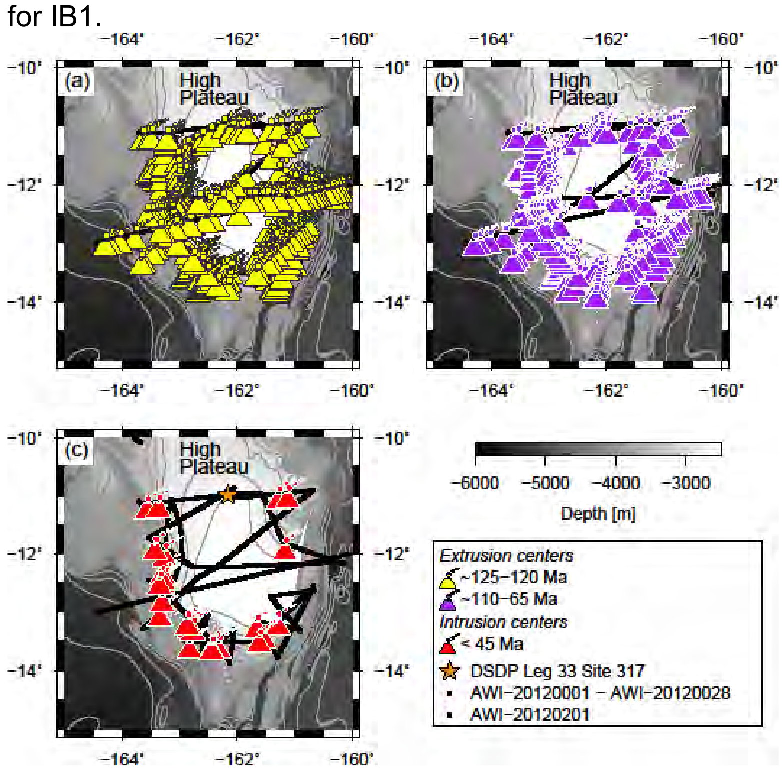


Fig. 9 Map of identified volcanic extrusion (a,b) and intrusion (c) centers located across the HP. For the different volcanic periods, associated with reflections R7 (~120 Ma) and R6 (~65

Ma) a different distribution of the extrusion centers hint on either volcanism related to emplacement of the High Plateau or tectonic-induced volcanism at the margins. Magmatic intrusions occur, suggesting that the south-western part was tectonically more active than the north-estern part.

Main volcanic emplacement phase

A massive volcanic enlargement of the HP has occurred after the emplacement of the IB1 reflection as documented by reflection R7 and the thickness of Unit 4 in Figures 8 (c) and (f). Radiating from the southern-central part of the HP to its margins, the thickness of Unit 4 varies from ~700ms TWT to ~3600ms TWT, which corresponds to a calculating thickness of 1.5 km for the main volcanic eruption in the central High Plateau. An overview of the identified volcanic extrusion centers is given in Figure 9. For reflection R7 (Fig. 9a) extrusion centers can be observed on every profile, distributed in the central part as well as on the eastern, southern and western flanks of the HP. We conclude that this period represents the main emplacement phase of the HP, which is supported by petrological data (Hoernle et al., 2010; Timm et al., 2011) as the lava is tholeiitic basalt. The identified multitude of extrusion centers (Fig. 9) matches well with the identification of strong volcanic activities on other LIPs in the Pacific Ocean during the early Cretaceous (Rea and Vallier, 1983) as well as reconstruction studies of the joined Ontong-Java-Nui (Taylor, 2006).

Late-stage emplacement volcanism phase

In the western Pacific Ocean a second volcanic period has occurred around 90 Ma (Rea and Vallier, 1983). The end of this volcaniclastic deposition period around 65 Ma is marked by reflection R6 (The Shipboard Scientific Party, 1976). The topography of the reflection R6 follows the overall trend of reflection R7 (Fig. 8a). The thickness of Unit 3, bounded by reflections R7 and R6 shows a constant maximum increase of ~400 ms TWT (Fig. 8d). Extrusion centers seen within reflection R6 show a clustered distribution with a major concentration along the eastern, southern and western flanks of the HP (Fig. 9b). Just a few magmatic extrusion centers were identified in the south-central, elevated part of the plateau. We therefore conclude that the magmatic activity moved from the center to the flanks of the HP.

A multitude of magmatic extrusion centers with a total number of more than 510 extrusion centers is visible for the top of basalt (R7) and the end of volcaniclastic deposition (R6) (Fig. 9a and b). A previous single channel seismic reflection study detected a field of 100 cones on the north-eastern edge of the HP (Coulbourn and Hill, 1991). We suggest that for the extrusion centers, different source mechanisms have to be taken into account, which is supported by petrological data (Beiersdorf et al., 1995; Timm et al., 2011). We therefore distinguish between magmatism associated with the emplacement of the HP itself and later tectonic induced magmatism. We infer that extrusion centers within reflection R7 are mainly emplacement expressions whereas extrusion centers of reflection R6 are primarily induced by tectonics.

At the transform faulted Manihiki Scarp (Viso et al., 2005), multiple ridge-type seamounts have been identified (Fig. 5). The intra-basalt reflection IB1 cannot be traced further eastward than the edge of the Manihiki Scarp. Instead, it gets cut off abruptly by volcanism of reflection R7 and R6. As no subsidence of IB1 across the Manihiki Scarp has occured it suggests that reflection IB1 continued beyond the Manihiki Scarp before the emplacement of R7 and R6. This result supports the hypothesis of Larson (2002) who suggested that a missing piece of the Manihiki Plateau was located at this eastern margin during the emplacement period and was drifted away to the east and subducted under the Farallon Plate.

Along the south-western margin normal fault structures are developed within R7.They show that rifting processes have occurred ~120 Ma ago. They could be related to crustal stretching associated with the break-up of the Hikurangi Plateau (Davy et al., 2008) at the Osbourn Trough (Worthington et al., 2006) ~115 Ma ago (Mortimer et al., 2006). The smooth appearance beyond the south-western margin supports a stretching breakup mechanism instead of a transform faulted margin as ridge-type seamounts are absent.

Across the Suvarov trough, which marks the western margin of the HP a graben structure can be observed, growing towards the north (Fig. 6). The inlcined reflections show that this structure has a maximum age of less than 65 Ma, and has therefore developed after the late stage volcanic active period. A strong tectonic activity is also displayed by the fracture zones and faults, which occur in particular in the western-southern part of the HP (Fig. 4c) and penetrate Units 1, 2 and 3, and to some extent Unit 4.
Post-emplacment tectonic and volcanic activity
The identified intrusion centers have been mapped in Fig. 9 c. In the north-east a few intrusions are visible, which show an age > 45 Ma. Along the Manihiki Scarp no intrusion centers are visible. The multitude of intrusion centers is concentrated on the western and southern flanks of the HP with ages ranging from early Eocene to recent, e.g. CDP 7600 in Profile AWI-20120201 (Fig. 7). As seen for example in Profile AWI-20120010 the south-western part of the HP shows numerous recent tectonic structures (Fig. 4c). We suggest that the identified intrusions at the western and southern flanks of the HP represent a reactivation of the Manihiki Plateau.

Conclusions
We have investigated the volcanic emplacement history of the High Plateau (Winterer et al., 1974), part of the Manihiki Plateau using new high resolution seismic reflection data gathered during cruise So224 (Uenzelmann-Neben, 2012). We extended the existing seismostratigraphic model (Ai et al., 2008; The Shipboard Scientific Party, 1976) by incorporating new reflections R5a, R6a and IB1. Reflection R5a corresponds to the onset of sedimentation after the third volcanic period (~65 Ma). Reflection R6a represents the onset of the alkali late stage volcanic period (~85 Ma) (Rea and Vallier, 1983; The Shipboard Scientific Party, 1976). The intra-basalt reflection IB1 likely represents an earlier volcanic flow period prior to R7 (>120 Ma). The intra-basalt reflection sequence is elevated in the southern central part of the HP and could represent a shield volcano and/or a guyot structure. We conclude that it represents the onset of the first volcanic emplacement phase within oceanic crust prior to the main emplacement phase. Age estimations are difficult due to the lack of borehole data but suggest an age of ~132 Ma. We were able to correlate the two volcanic periods identified on LIPs in the west Pacific Ocean (Rea and Vallier, 1983) to reflections R7 and R6 and could trace them throughout the whole HP. The end of the major initial emplacement (Unit 4) is marked by the top of basalt (reflection R7) and characterized by a multitude of extrusion centers. We interpret those as a result from initial emplacement volcanism in the central part of the HP (R7) and volcanism induced by tectonics at the eastern, western and southern flanks. For the late stage volcanic period, ending ~65 Ma, the spatial distribution of the identified extrusion centers shows a relocation from the center of the HP towards its eastern, southern and western flanks. We therefore conclude that two different source mechanisms have to be considered; sources either related to the initial emplacement of the HP and tectonic induced volcanism at the margins. This is well in agreement with petrological data taken along the Suvarov and Danger Island Troughs (Timm et al., 2011).
At the eastern margin, the abrupt termination of the intra-basalt reflection IB1 across the Manihiki Scarp support the hypothesis that a missing piece of the Manihiki Plateau rifted to the east (Larson et al., 2002) before the formation of reflection R7. The south-western margin is characterized by a normal fault structure and a smooth appearance of reflections R7 and R6. This appearance supports the hypothesis of a stretched and rifted margin, which is associated with the break-up of the Hikurangi Plateau at the Osbourn Trough (Davy et al., 2008; Hoernle et al., 2010; Worthington et al., 2006). The graben structure along the Suvorov Trough and fault structures give evidence that the HP has been tectonically active after the alkali late stage volcanic period (<65 Ma). Analyses of intrusion structures provide a spatial distribution of recent magmatism accumulating in the western and southern part showing that this part of the HP was recently active.

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References

Ai, H.-A., Stock, J., Clayton, R., Luyendyk, B., 2008. Vertical tectonics of the High Plateau region, Manihiki Plateau, Western Pacific, from seismic stratigraphy. Mar Geophys Res 29, 13-26.
Beiersdorf, H., Bach, W., Duncan, R., Erzinger, J., Weiss, W., 1995. New Evidence for the Production of Em-Type Ocean Island Basalts and Large Volumes of Volcaniclastites during the Early History of the Manihiki Plateau. Marine Geology 122, 181-205.
Bryan, S.E., Ernst, R.E., 2008. Revised definition of Large Igneous Provinces (LIPs). Earth-Science Reviews 86, 175-202.
Chandler, M.T., Wessel, P., Taylor, B., Seton, M., Kim, S.-S., Hyeong, K., 2012. Reconstructing Ontong Java Nui: Implications for Pacific absolute plate motion, hotspot drift and true polar wander. Earth and Planetary Science Letters 331/332, 140-151.
Coffin, M.F., Eldholm, O., 1994. Large Igneous Provinces - Crustal Structure, Dimensions, and External Consequences. Rev Geophys 32, 1-36.
Coulbourn, W.T., Hill, P.J., 1991. A field of volcanoes on the Manihiki Plateau: Mud or lava? Marine Geology 98, 367-388.
Davy, B., Hoernle, K., Werner, R., 2008. Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history. Geochemistry Geophysics Geosystems 9.
Downey, N.J., Stock, J.M., Clayton, R.W., Cande, S.C., 2007. History of the Cretaceous Osbourn spreading center. J Geophys Res-Sol Ea 112.
Frey, F.A., Coffin, M.F., Wallace, P.J., Weis, D., Zhao, X., Wise, S.W., Wahnert, V., Teagle, D.A.H., Saccocia, P.J., Reusch, D.N., Pringle, M.S., Nicolaysen, K.E., Neal, C.R., Muller, R.D., Moore, C.L., Mahoney, J.J., Keszthelyi, L., Inokuchi, H., Duncan, R.A., Delius, H., Damuth, J.E., Damasceno, D., Coxall, H.K., Borre, M.K., Boehm, F., Barling, J., Arndt, N.T., Antretter, M., 2000. Origin and evolution of a submarine large igneous province: the Kerguelen Plateau and Broken Ridge, southern Indian Ocean. Earth Planet Sc Lett 176, 73-89.
Ganino, C., Arndt, N.T., 2009. Climate changes caused by degassing of sediments during the emplacement of large igneous provinces. Geology 37, 323-326.
Gohl, K., Uenzelmann-Neben, G., 2001. The crustal role of the Agulhas Plateau, southwest Indian Ocean: evidence from seismic profiling. Geophysical Journal International 144, 632-646.
Heezen, B.C., Glass, B., Menard, H.W., 1965. The Manihiki Plateau.
Hochmuth, K., Gohl, K., Uenzelmann-Neben, G., Werner, R., 2014. The diverse crustal structure and magmatic evolution of the Manihiki Plateau, Central Pacific. Alfred-Wegener-Institute Helmholtz-Zentrum für Polar- und Meeresforschung.
Hoernle, K., Hauff, F., van den Bogaard, P., Werner, R., Mortimer, N., Geldmacher, J., Garbe-Schonberg, D., Davy, B., 2010. Age and geochemistry of volcanic rocks from the Hikurangi and Manihiki oceanic Plateaus. Geochim Cosmochim Ac 74, 7196-7219.
Hoernle, K., Hauff, F., Werner, R., 2004. New insights into the origin and evolution of the Hikurangi oceanic plateau. EOS 85, 401-410.
Ingle, S., Coffin, M.F., 2004. Impact origin for the greater Ontong Java Plateau? Earth Planet Sc Lett 218, 123-134.
Inoue, H., Coffin, M.F., Nakamura, Y., Mochizuki, K., Kroenke, L.W., 2008. Intrabasement reflections of the Ontong Java Plateau: Implications for plateau construction. Geochem Geophy Geosy 9.

Jackson, E.D., Bargar, K.E., Fabbi, B.P., Chris, H., 1976. Petrology of the basaltic rocks drilled on Leg 33 of the Deep Sea Drilling Project, in: Kaneps, A.G. (Ed.), Initial Rep Deep Sea. US Governmental Printing Office, Washington, DC, pp. 571-630.
Jenkyns, H.C., 1976. Sediments and sedimentary histroy of the Manihiki Plateau, South Pacific Ocean, in: Kaneps, A.G. (Ed.), Initial reports of the Deep Sea Drilling Project. US Governmental Printing Office, Washington, D. C., pp. 873-890.
Keszthelyi, L., Self, S., 1998. Some physical requirements for the emplacement of long basaltic lava flows. J Geophys Res-Sol Ea 103, 27447-27464.
Klarner, S., Klarner, O., 2012. Identification of Paleo-Volcanic Rocks on Seismic Data, in: Stoppa, P.F. (Ed.), Updates in Volcanology - A Comprehensive Approach to Volcanological Problems. InTech.
Lanphere, M.A., Dalrymple, G.B., 1976. K-AR Ages of basalts from DSDP Leg 33: Sites 315 (Line Island) and 317 (Manihiki Plateau), in: Kaneps, A.G. (Ed.), Initial Rep Deep Sea. US Governmental Printing Office, Washington, D. C., pp. 649-653.
Larson, R.L., 1997. Superplumes and ridge interactions between Ontong Java and Manihiki plateaus and the Nova-Canton trough. Geology 25, 779-782.
Larson, R.L., Erba, E., 1999. Onset of the mid-Cretaceous greenhouse in the Barremian-Aptian: Igneous events and the biological, sedimentary, and geochemical responses. Paleoceanography 14, 663-678.
Larson, R.L., Pockalny, R.A., Viso, R.F., Erba, E., Abrams, L.J., Luyendyk, B.P., Stock, J.M., Clayton, R.W., 2002. Mid-Cretaceous tectonic evolution of the Tongareva triple junction in the southwestern Pacific Basin. Geology 30, 67-70.
Mahoney, J.J., Storey, M., Duncan, R.A., Spencer, K.J., Pringle, M., 1993. Geochemistry and Age of the Ontong Java Plateau, The Mesozoic Pacific: Geology, Tectonics, and Volcanism. American Geophysical Union, pp. 233-261.
Mortimer, N., Hoernle, K., Hauff, F., Palin, J.M., Dunlap, W.J., Werner, R., Faure, K., 2006. New constraints on the age and evolution of the Wishbone Ridge, southwest Pacific Cretaceous microplates, and Zealandia-West Antarctica breakup. Geology 34, 185-188.
Nakanishi, M., Sager, W.W., Klaus, A., 1999. Magnetic lineations within Shatsky Rise, northwest Pacific Ocean: Implications for hot spot-triple junction interaction and oceanic plateau formation. J Geophys Res-Sol Ea 104, 7539-7556.
Rea, D.K., Vallier, T.L., 1983. Two Cretaceous volcanic episodes in the western Pacific Ocean. Geol Soc Am Bull 94, 1430-1437.
Rotstein, Y., Schlich, R., Munschy, M., Coffin, M.F., 1992. Structure and tectonic history of the Southern Kerguelen Plateau (Indian Ocean) deduced from seismic reflection data. Tectonics 11, 1332-1347.
Sager, W.W., Zhang, J., Korenaga, J., Sano, T., Koppers, A.A.P., Widdowson, M., Mahoney, J.J., 2013. An immense shield volcano within the Shatsky Rise oceanic plateau, northwest Pacific Ocean. Nature Geosci advance online publication.
Schaming, M., Rotstein, Y., 1990. Basement reflectors in the Kerguelen Plateau, south Indian Ocean: Indications for the structure and early history of the plateau. Geol Soc Am Bull 102, 580-592.
Taylor, B., 2006. The single largest oceanic plateau: Ontong Java-Manihiki-Hikurangi. Earth Planet Sc Lett 241, 372-380.
The Shipboard Scientific Party, 1976. Site 317, in: Kaneps, A.G. (Ed.), Initial Rep Deep Sea. US Governmental Printing Office, Washington, D.C., pp. 161-300.
Timm, C., Hoernle, K., Werner, R., Hauff, F., den Bogaard, P.v., Michael, P., Coffin, M.F., Koppers, A., 2011. Age and geochemistry of the oceanic Manihiki Plateau, SW Pacific: New evidence for a plume origin. Earth and Planetary Science Letters 304, 135-146.
Uenzelmann-Neben, G., 2012. The Expedition of the Research Vessel "Sonne" to the Manihiki Plateau in 2012 (SO 224), in: Uenzelmann-Neben, G. (Ed.), Reports on Polar and Marine Research. Alfred-Wegener-Institut für Polar- und Meeresforschung in der Helmholtz Gemeinschaft, Bremerhaven.

Uenzelmann-Neben, G., Gohl, K., Ehrhardt, A., Seargent, M., 1999. Agulhas Plateau, SW Indian Ocean: New evidence for excessive volcanism. Geophysical Research Letters 26, 1941-1944.
Viso, R.F., Larson, R.L., Pockalny, R.A., 2005. Tectonic evolution of the Pacific–Phoenix–Farallon triple junction in the South Pacific Ocean. Earth Planet Sc Lett 233, 179-194.
Werner, R., Hauff, F., 2007. FS Sonne Fahrtbericht / Cruise Report So193 Manihiki, Temporal, Spatial, and Tectonic Evolution of Oceanic Plateaus. Leibniz-Institut für Meereswissenschaften, Kiel.
Winterer, E.L., Lonsdale, P.F., Matthews, J.L., Rosendah.Br, 1974. Structure and Acoustic Stratigraphy of Manihiki Plateau. Deep-Sea Res 21, 793-&.
Worthington, T.J., Hekinian, R., Stoffers, P., Kuhn, T., Hauff, F., 2006. Osbourn Trough: Structure, geochemistry and implications of a mid-Cretaceous paleospreading ridge in the South Pacific. Earth Planet Sc Lett 245, 685-701.

PIETSCH, R., UENZELMANN-NEBEN, G. (SUBMITTED): HOTSPOT TRACKS DISCOVERED AT THE MANIHIKI PLATEAU, A LARGE IGNEOUS PROVINCE IN THE CENTRAL PACIFIC. EARTH AND PLANETARY SCIENCE LETTERS-ANLAGE 5

Hotspot tracks discovered at the Manihiki Plateau, a Large Igneous Province in the central Pacific

R. Pietsch*, G. Uenzelmann-Neben

Alfred Wegener Institute Helmholtz Center for Polar and Marine Research, Postfach 120161, 27515 Bremerhaven, Germany

*Corresponding author: Ricarda Pietsch, Email: ricarda.pietsch@awi.de, Tel.: +49 (0)471 4831 1206

Abstract

Seismic reflection profiles of Kiwi 12 and So224 across the Manihiki Plateau, a Large Igneous Province (LIP) in the central Pacific, are investigated to solve the question of origin of Tertiary magmatic and Quaternary tectonic reactivations, which have occurred after the initial LIP formation. For the first time, identified acoustic diapirs provide evidence for induced hotspot volcanism exclusively in the southern High Plateau (HP), which are probably related to the Society Islands and Tahiti hotspots (22-14 Ma). The penetration of the anomalously thick LIP crust can be explained by a fractured and stretched crust associated to the break-up of the Hikurangi Plateau. Additionally, some indications for volcanism of the Pitcairn hotspot are found in the northern High Plateau. A Quaternary tectonic reactivation of the High Plateau occurred as a northwest-southeast striking graben structure, which appears to form an extension of the Suvarov Trough, for which we found evidence that it was formed after the end of the initial LIP formation volcanism (< 65 Ma).

1. Introduction

Intra-plate, age progressive, volcanic hotspot chains have been in the focus of research due to their importance for understanding the Earth's dynamic system, e.g. mantel convection and source and behavior of time dependent intra-plate volcanism. They have been used as an important tool, for example as a reference frame in absolute plate motion models (e.g. Wessel and Kroenke, 2008), which are used in plate reconstruction models (e.g. Chandler et al., 2012; Seton et al., 2012). In general, they are believed to display surface images of hot rising mantle plumes originating from a boundary layer (Morgan, 1972), although other explanation models exist including shear-stress driven upwelling within the lithosphere (e.g. Ballmer et al., 2013). A self-consistent model is still in progress as some open questions, for example relative movements of hotspots in time, are still lacking conclusive answers. To gain a better understanding, it is important to investigate areas, where the mapped hotspots are missing and have been interpolated so far.

Figure 1

Especially in the central Pacific Ocean, the occurrence of crossing hotspot tracks of similar age and absence of continuous seamount chains make an easy and strait forward explanation and modeling difficult. The Pitcairn Hotspot, Society Island Hotspot, Tahiti Hotspot in the central Pacific - to state a few recently active hotspots (Clouard and Bonneville, 2001) - age towards the west-northwest and are limited by subduction zones in the northern and western Pacific (Fig.1). In the central Pacific the seamount chains are not traceable. What causes the absence? Is the morphology in this area different to the surrounding seafloor?

In this region, the Manihiki Plateau, a Large Igneous Province, is located, which has an anomalous thick oceanic crust up to 20 km (Hochmuth et al. 2015, in review.). Those circumstances raise the following questions, 1) can hotspot volcanism be identified on the Manihiki Plateau, 2) can later stage hotspot volcanism break through anomalously thick oceanic crust, 3) can volcanic overprinting be distinguished from early formation volcanism and 4) how does hotspot volcanism influence tectonic activity.

Within this study we have used seismic reflection data gathered during cruise So224 (Pietsch and Uenzelmann-Neben, 2015; Uenzelmann-Neben, 2012a) and cruise Kiwi12 (Ai et al., 2008).

2. Geological background

The Manihiki Plateau is a Large Igneous Province located in the central Pacific (Fig.1). It is supposed to be the centerpiece of Ontong-Java-Nui, a Super-LIP (Chandler et al., 2012; Taylor, 2006a, Hochmuth et al. 2015, in review). During its main formation in the early Cretaceous (~ 125-110 Ma), massive eruptions of basaltic lava occurred (Hoernle et al., 2010; Lanphere and Dalrymple, 1976; Larson and Erba, 1999; Mahoney et al., 1993; The Shipboard Scientific Party, 1976; Timm et al., 2011), which resulted in an anomalously thick oceanic crust (Hussong et al., 1979) up to 20 km (Hochmuth et al. 2015, in review). Secondary volcanism (~90- 65 Ma) consisted of alkalic volcaniclastics and is related to the formation of seamounts at the margin of the Manihiki Plateau (Pietsch et al. 2015, in prep.).

Tectonic processes played an important role for the development of the different morphologies of the three sub-provinces, the Western Plateaus, the North Plateau and the High Plateau (Winterer et al., 1974b). In the west, the Danger Islands Troughs separate the Western Plateaus from the High Plateau. It is a series of pull-apart structure, which were formed when the Ontong-Java Plateau broke-off the Western Plateaus around 118 Ma (Hochmuth et. al 2015, in prep). In the north, the Pacific-Farallon-Phoenix triple junction was displaced to the newly formed Manihiki Plateau and originated at the Tongareva triple junction at 119 Ma with spreading rates average 18-20 cm/yr until 84 Ma (Larson et al., 2002). The strike-slip (transform) fault developed from north to south along the Manihiki Scarp, the eastern margin of the High Plateau. It displays the break-up of two fragments, which rifted along the east on the Farallon and Phoenix Plates. In agreement, recent seismic reflection data (Pietsch and Uenzelmann-Neben, 2015) hold evidence that the High Plateau was extended to the east > 125 Ma and the onset of the break-off occurred between the ending of the main emplacement period and the beginning of the secondary volcanic period (~116-110 Ma). Additionally, seismic reflection data show that volcanism along the ridge structure lasted until 65 Ma.

The morphology of the southern and south-western part of the High Plateau is determined by stretched and rifted crustal structures associated with the break-up of the Hikurangi Plateau (Ai et al., 2008; Pietsch and Uenzelmann-Neben, 2015, Hochmuth et al. 2015, in prep.). Normal faults are exhibited, which show a south-westward orientation. The spreading center developed at the Osbourn Trough ~118-86 Ma and ceased when the Hikurangi Plateau collided with a southward-dipping subduction system developed along Gondwana Margin (eastern New Zealand) (Davy et al., 2008; Worthington et al., 2006). The Nassau Plateau, a terrace-like surface at the south-western margin of the High Plateau bounded by the Nassau Island, is separated from the High Plateau by the Suvarov Trough. The presented seismic reflection data hint towards a formation unrelated to the break-up of the Hikurangi Plateau. Therefore, this tectonic feature is discussed in detail in section 0.

3. Data acquisition and seismic stratigraphy

The data basis for this study are single channel seismic reflection data (SCS) of cruise Kiwi12 (Ai et al., 2008; Stock et al., 1998) and multichannel high resolution data (MCS) gathered during cruise So224 (Uenzelmann-Neben, 2012a), which cover the High Plateau (Fig. 1b) and the surrounding deep sea.

Kiwi12 SCS data were gathered using two 210 cubic inch air guns, firing with a 10 s interval, recording length of 7s and an average trace interval spacing of 45 m (Ai et al., 2008). So224 MCS data were collected using a cluster of four GI-guns (volume 45 in^3 + 105 in^3), which were fired in a depth of 2 m with a shot interval of 10 s (~25 m). For profile AWI-20120201, a cluster of eight G-Guns (total volume 4160 in^3) in a depth of 10 m was used with a shot interval of 60 s. Data were recorded with a 3 km long, 240-channel streamer (Sercel SEAL™) at a sampling rate of 1 ms. A frequency range up to 250 Hz allowed for a maximum vertical resolution of about ~3 m at the ocean bottom (Pietsch and Uenzelmann-Neben, 2015; Uenzelmann-Neben, 2012a).

Processing steps for the MCS data, included Common Depth point sorting (CDP), analyses of velocity-time-profiles every 50th CDP, Normal-Move-Out correction, stacking and time migration.

For the MCS data, a correlation of seismic stratigraphy to borehole data of DSDP Leg 33 Site 317 (The Shipboard Scientific Party, 1976) was carried out using a synthetic seismogram and summarized by (Pietsch and Uenzelmann-Neben, 2015) (supplementary material). As the Kiwi12 profiles never crossed the DSDP Leg 33 Site 317 location, the SCS data have been correlated by comparing the seismic stratigraphy to crossing lines of CATO3 (Ai et al., 2008) and crossing lines of So224 (Uenzelmann-Neben, 2012a).

The seismic stratigraphy has been correlated to lithology and age of borehole DSDP Leg 33 Site 317 (The Shipboard Scientific Party, 1976) and summarized in (Pietsch and Uenzelmann-Neben, 2015) (supplementary material). Seven reflections R1-R7 have been identified. The sedimentary column has been divided into four Units. Sediments of Units 1 and 2 comprise Quaternary calcareous ooze to Late Eocene chalk, bounded by unconformities. Units 3 and 4 were formed by basaltic lavas and volcaniclastic deposition during the initial formation of the Manihiki Plateau.

4. Results

4.1 Identification of Tertiary volcanism

Volcanic structures can be identified in the seismic reflection data using their characteristic facies, such as geometry e.g. divergent vs. subparallel reflections and amplitude characteristics, e.g. alteration of strong and weak amplitudes (Klarner and Klarner, 2012; Magee et al., 2013; Pietsch and Uenzelmann-Neben, 2015; Planke et al., 2005; Planke et al., 2000; Uenzelmann-Neben et al., 1999). Volcanic structures associated with the emplacement period (~125-116 Ma), pre-main (>125 Ma) and secondary volcanism (95-65 Ma), have been discussed in (Pietsch and Uenzelmann-Neben, 2015, in prep.). In the following, we focus on volcanic structures seen as acoustic diapirs within the sedimentary column, which occurred after the initial emplacement period.

4.1.1 Volcanic structures in the north-east

The northern part of the High Plateau is characterized by a terrace-like surface with an exhibition of distinct volcanic islands like the Rakahanga Island and Manihiki Island (Fig. 2c). Profile Kiwi1211 prolongates in south-southeast to north-northwest direction, north of Rakahanga Island (Fig. 2 b and c). Along the seismic profile, the morphology of the strike-slip faulted Manihiki Scarp changes to a smooth topographic margin of the High Plateau (Fig. 2a). As seen between CDP 2600-5000 of Profile Kiwi1211, two strong reflection units Unit 3 and Unit 1-2 can be distinguished, bounded by the acoustic basement (reflection R7), reflection R6 and the seafloor respectively. Unit 3 has previously identified to consist of volcaniclastic material, associated with the secondary volcanism of the LIP formation (95-65 Ma) (Beiersdorf et al., 1995a; Pietsch and Uenzelmann-Neben, 2015). From southeast to northwest, a buried depression (CDP 2600-3200) is located adjacent to the Manihiki scarp, which is bounded by a tectonically faulted structure in the northeast (CDP 3100-3500). A basement depression (CDP 3500-5000) has been filled and major volcanic structures can be identified within reflection R6 (CDP 4500-4700). Erosional unconformities occur, leading to a lack of sediments.

Figure 2

Some intrusion and/or tectonic features (CDP 2800, 3900 and 4100) with diameters between ~1.5-4.3 km can be identified, which penetrate the remaining sedimentary column. The most prominent one (CDP 2750-2850) is a magmatic structure, which shows a clear deformation of the upper sediments. Dating of these features is difficult, as erosion has occurred and a correlation to previously defined reflections remains with uncertainties. Reflection R6 however, marks the maximum age.

4.1.2 Volcanic structures in the south

Distinct acoustic diapirs could be identified in the southern part of the High Plateau, which are described in the following from West to East. The westernmost ones are located along profiles AWI-20120201/-028/-023 and Kiwi1204 (Fig. 3). Along Profile AWI-20120201 the most prominent feature (CDP 6950-7100 in Fig. 3) penetrates the entire sedimentary column and exhibits a bathymetric high of about 300 m (Fig. 3). With a diameter of about 5 km, the extrusion is characterized by a sudden decrease of amplitude in the MCS data. Reflections above R5 are bent upward. This specific seismic facie shows that the strata was already deposited when the intrusion occurred and therefore gives it its maximum age. Erosional unconformities occur within Unit 2. Reflection R3 marks the lower boundary of an onlap structure. This seismic facie develops after the intrusion was emplaced. Therefore it gives the minimum age of the intrusion. Using the age correlated seismic stratigraphy of the MCS data, the defined reflections R3-R5 reveal an age of 10-45 Ma for the intrusion. Vertical fault structures are visible south-west of the magmatic feature. They show a vertical displacement up to 60 ms TWT.

Figure 3

Furthermore, another acoustic diapir (CDP 6950-7100 in Fig. 3) and a fault system (CDP 6550-6700), which is discussed in more detail in section **Error! Reference source not found.**, can be identified along Profile AWI-20120201. This magmatic feature has penetrated and disturbed Units 3 and 4 (R6-R7). Reflections R7-R5 are bent upwards. Reflection R5 is undisturbed. This either suggests an older age for the intrusion or the intrusion got stuck below. The second possibility is supported by the absence of an onlap structure. Therefore, only a maximum age of ~45 Ma (R5) can be given. At its south-western margin a vertical fault is developed (R7-R5).

Along the north-south striking profile AWI-20120023, three intrusion centers can be identified. The most prominent one is displayed in CDP 1500-1650 (Fig. 3) and crosses the base of at a bathymetric high with elevations of 250 m (Fig. 3). The seismic facies show an amplitude disturbance within Units 2-4 (Reflections R5-R7) in the center and upward bending reflections at the edge. Onlap structures can be identified for the upper seismic sequence (Reflection R2 in Fig. 3b). Erosional unconformities occur within Unit 2 (Reflection R4). Age estimation is given by the disturbed reflection R5, which reveals a maximum age of ~45 Ma and reflection R3, which states a maximum age of ~10 Ma.

Figure 4

In the south-east of the HP, intrusion centers could be identified on Profiles AWI-20120013/-016 and Kiwi1206, printed in in (Ai et al., 2008) (Fig. 4). Profile AWI-20120013 is located at the south-eastern margin of the HP. An intrusion is visible between CDP 3300-3500 (~5km), which exhibit the same characteristic as the above described intrusions (Fig. 4). In the central zone the amplitudes of Units 2-4 (Reflection R5-R7) are disturbed and bent upwards at the edge of zone. The deposition of reflections R3 and R4 appears to be complete with smaller disturbances, whereas reflections R1 and R2 are completely undisturbed. An erosional unconformity (R2-R3) can be identified ~2.5 km northwest of the intrusion.

4.2 Cenozoic tectonic structures

4.2.1 Suvarov Trough

The appearance of the south-western margin of the High Plateau has been described as a terrace-like surface with smooth topography and a few minor relief features (Heezen et al., 1966). It is separated from the High Plateau by the Suvarov Trough, a ~30 km wide graben structure.

Figure 5

From north to south, profiles AWI-20120001, Kiwi1201, Kiwi1202 and AWI-20120006 cross the Suvarov Trough (Fig. 5). Profile AWI-20120001 covers the eastern margin of the Suvarov Trough, which is characterized by steep flanks (Fig. 5a). Reflections R7 and R6, associated with the formation of the Large Igneous Province (Pietsch and Uenzelmann-Neben, 2015), can clearly been traced into the Suvarov Trough. They show a local elevation high with a dipping reflection towards the margin. Erosion unconformities can be seen within the sedimentary column above the volcanic sequence. Further on the High Plateau (CDP 3550-3600 in Fig. 5), another series of vertical faults can be seen in Profile AWI-20120001. Erosional unconformities occur between reflection R6 and R5a with a divergent filling of the depression afterwards.

Profile Kiwi1201, partly presented by (Ai et al., 2008) crosses the entire Suvarov Trough almost perpendicular (supplementary material). The eastern (CDP 10800-11270) and western (CDP 9500-10000) fragments show the same morphology – the two phases of volcanism (reflections R7 and R6) and the same reflection patterns for the sedimentary column above. The trough shows fragmentation to the west (CDP 10100-10400) and a deep graben structure to the east (CDP 10450-1600) as well as erosional unconformities.

The northeast-southwest trending profile Kiwi1202 shows a similar morphology as the previous profile Kiwi1201. Between CDP 45-600 the southwestern margin of the High Plateau. Onlap structures are developed on top of the reflection associated with the volcaniclastic deposition (R6), showing strong erosion and absence of reflectors R3-R4. A deep normal fault structure marks the onset of the graben (CDP 650), which is fragmented and serpentinizated towards the southwest (CDP 900-1400). The sedimentary deposits cannot easy be linked to the previously defined reflectors. Beyond the Suvarov Trough (CDP 1350-2000) the morphology of the High Plateau continues. In this segment, reflections R7 and R6 can be traced but most of the sedimentary column is eroded.

In the southern-most profile AWI-20120006 (Fig. 5b) the typical morphology of the High Plateau (CDP 1200-4000) shows onlap structures towards the margin within the sedimentary column. Within the deepest part of the graben structure (CDP 800-1100) reflections R7 and R6 dip towards the High Plateau. Wedging of the inclined reflections with overlying horizontal reflections show that this fault structure has been emplaced after reflection R6 was deposited. Therefore, the graben has a maximum age of less than 65 Ma (Fig. 5b) and developed after the late stage volcanic emplacement period.

4.2.2 Quaternary graben structures

In general, the sedimentary facies on the High Plateau is parallel layered and with small tectonic disturbances and an overall smooth topography. However, in spatial distinct locations, graben structures can be identified, which penetrate the entire sedimentary column, e.g. Profile AWI-20120010 and AWI-201200201 (Fig. 3d) and Kiwi1205.

A major example is displayed in Profile AWI-20120201 CDP 6550-6700 (Fig. 3), where a series of graben and horst structures is developed. Vertical displacements up to 60 ms occur.

Figure 6

Another significant example can be seen along Profile AWI-20120010 (Fig. 6a), which crosses the High Plateau in a northeast-southwest direction. The graben structure (CDP 1100-1500) is located in the south-central part of the High Plateau. It consists of a series of normal faults, which penetrate the entire sedimentary and column (reflections R1-R7).

In the south-west of the High Plateau a notable example has first been published in (Ai et al., 2008) (Fig. 6b). An old volcano, first been interpreted to form a " deeply buried basement high" with the sedimentary column "draped over previously disturbed basement high" (Ai et al., 2008) displays the base of the structure. It was emplaced in two periods, their ends marked by reflections R7 and R6, respectively. The main emplacement period (R7) occurred around ~125-116 Ma, the secondary volcanic phase (R6) ended around ~65 Ma (Pietsch and Uenzelmann-Neben, 2015; The Shipboard Scientific Party, 1976). Afterwards the caldera of the volcano collapsed and formed the disturbed depression around 45 Ma (reflection R5), followed by pelagic sedimentation, which formed a parallel layered sedimentary column. A series of normal fault structures crosses the entire sedimentary column (reflections R1-R7) and forms a displacement, which is strongest in the center of the caldera. The maximum formation age of the graben structure of < 1.8 Ma is given by reflection R1.

5. Discussion

5.1 Can hotspot tracks be identified on the Manihiki Plateau?

5.1.1 North-eastern part of the High Plateau

In the pole rotation model of (Wessel and Kroenke, 2008) it is calculated that the Pitcairn Hotspot crossed the north-eastern part of the High Plateau between 57-42 Ma. In order to investigate if hotspot volcanism has occurred in the north-eastern High Plateau, bathymetry and seismic reflection profiles have been analyzed (Fig.2).

The bathymetry shows a terrace like morphology descending from the High Plateau with an occurrence of distinct volcanic features like the Rakahanga or the Manihiki Atolls. At least two unnamed volcanoes can be identified within a 100 km distance range of the hotspot trail. Dredge samples of Mr. Eddie and surrounding area (e.g. Beiersdorf et al., 1995a) revealed an age up to the late Paleocene for the limestone interbedded with volcaniclastic material. Overall, they concluded those features were formed by volcaniclastic material associated with the secondary volcanism of the initial LIP formation, which ended ~65 Ma. This is supported by the Kiwi12 seismic reflection data, where two phases of volcanism associated with the initial LIP emplacement can clearly be identified and built the topographic base.

Only a few, small late stage tectonic and magmatic features could be identified in profile Kiwi11 (Fig. 2a). The most prominent one (diameter ~1.7 km) is an acoustic diapir located in the filled depression adjacent to the Manihiki Scarp. Age estimation remains difficult as multiple flow-units and erosional unconformities appear. They have modified the seismic reflection pattern so that no direct correlation of the previously defined reflections R1-R6 is possible. Nevertheless, the two volcanic phases associated with LIP formation can clearly be seen and give a maximum age of ~65 Ma for the acoustic diapir.

Overall, the bathymetric highs and tectonic and magmatic features seen in the seismic reflection data hint to an active period after ~65 Ma, which could be related to hotspot volcanism. Convincing constrains remain enigmatic due to the lack of high resolution multi-channel seismic reflection data and age determination.

5.1.2 Southern part of the High Plateau

In the southern parts of the High Plateau volcanic intrusion/extrusion centers could be identified based on seismic reflection characteristics; i) detection of an acoustic diapir, ii) amplitude reduction and disturbance of Units 3 and 4, iii) upward bending of reflections R6, R7 and partly R5. To prove whether the acoustic diapirs are related to hotspot volcanism of

Society/Tahiti hotspots or not, we will discuss their spatial distribution and age estimation in the following.

Age estimation was carried out using characteristic seismic patterns, to state upward bending reflections for the lower units. They were deposited before the intrusion took place and therefore give a maximum age. The minimum age was determined using onlap structure for extrusion centers (e.g. AWI-20120201 in Fig. 3) and changes from disturbed to undisturbed reflections within the upper units for intrusion structures. Erosional unconformities limited the age constrains, especially for the extrusion features. For an additional minimum age support we compared the upper reflection boundaries of adjacent tectonic features to the formation age of the acoustic diapirs. Summarized, the range of formation ages varies between 45 Ma and 5.3 Ma (R5-R2). A minor age progression is visible, as the south-western diapirs seem to be slightly older (>25 Ma, reflections R4-R5) than the south-eastern ones (<10 Ma), where the disturbance of reflection R3 is general visible.

In the High Plateau, 19 acoustic diapirs have been identified exclusively in the southern part. Their spatial distribution is shown in Figure 7. They trend is parallel to the interpolation path of (Wessel and Kroenke, 2008) and shifted 50-100 km to the south. The location displacement is well within the uncertainty error of 100 km for the global interpolation of the Pacific hotspot tracks. Additionally, the seismic reflection studies of (Pietsch and Uenzelmann-Neben, 2015) has shown that the southern High Plateau is characterized by a disturbed and stretched crust related to the break-up of the Hikurangi Plateau. Hence, has a weaker crust, which is likely a preferential condition for magma to rise through the crust. Summarizing, these characteristics fit the postulated criteria of hotspot volcanism from the Society Islands and Tahiti Hotspots. Therefore, we conclude that the identified diapirs are Tertiary intrusions/extrusion, which display hotspot volcanism of probably the Society Islands and Tahiti hotspots.

5.2 *Tectonic features*

5.2.1 The Suvarov Through

The Suvarov Trough is a northwest-southeast striking graben structure, which separates the central High Plateau from a terrace-like margin, the Nassau Plateau (Fig. 5). It has been suggested that the Nassau Plateau was deformed by gravity tectonics of Late Cretaceous or Early Cenozoic age, and that the Suvarov Trough is a break-away structure (Stock et al. 1998). An analysis and interpretation of the seismic reflection data (section 0) has shown a multi-segmented, complex fault system with an increase displacement from ~0.5 s TWT in the south to ~1.5 s TWT towards the Danger Islands troughs. Seamounts with a rough topography are absent. We conclude that the Suvarov Trough was formed due to stretching in southwest direction.

Age estimation could be given (e.g. profile AWI-20120006) by tracking the seismic units into the graben and identifying the boundary reflections of wedge structures. It revealed that the graben formed after the end of secondary volcanism associated with the initial LIP formation < 65 Ma (Pietsch and Uenzelmann-Neben, 2015). More precise age estimations remains difficult, as erosional unconformities have occurred.

What caused the extension of the Suvarov Trough remains enigmatic, as previously proposed models, e.g. induced stress field due to the break-up of the Super-LIP fragments fail to explain the Trough because of the timing issue, as rifting ceased ~85 Ma (Billen and Stock, 2000a; Worthington et al., 2006, Hochmuth et al., 2015, in prep.). Additionally, lithospheric stress fields induced from (i) hotspots or (ii) subductions zones seem rather unlikely. During the early Cenozoic, the Pitcairn, MacDonald and Austral-Cook hotspots are the nearest ones, which could have triggered the event. The Pitcairn hotspot trail crossed the Manihiki Plateau around 400 km northwest of the trough between 57-42 Ma. Between ~60-34 Ma, the McDonald and Austral-Cook hotspots passed the High Plateau in a distance of 400 km to the south (Fig. 1). Studies of hotspot trails in the Pacific, e.g. the Tarava seamount trail (Clouard et al., 2003) have shown that hotspot volcanism is a localized phenomenon. Some hotspot alternative models (e.g. Ballmer et al., 2013; Hieronymus and Bercovici, 2000) have suggested that the volcanism is triggered from horizontal stress components between East-Pacific-Rise and the Tonga-Kermadec subduction zone, but no vice-versa influence has been stated. Therefore, we infer no distant influence of hotspot volcanism on the lithospheric stress field.

During the middle Eocene (~50-45 Ma) a reorganization of the subduction zones occurred in the central Pacific (Gurnis et al., 2004; Neall and Trewick, 2008), which probably involved a short-lived north-east subduction zone (Eissen et al., 1998; Kroenke, 1984) and finally resulted in the westward subduction at the Tonga-Kermadec trench around 45 Ma (Bloomer et al., 1995; Gurnis et al., 2004). This event certainly induced a stress field into the lithosphere. Nevertheless, we presume a negligible influence on the Manihiki Plateau, as the plate reconstruction model of (Seton et al., 2012) has calculated a distance of more than 5000 km to the Tonga-Kermadec trench at 45 Ma.

Here, we present an alternative formation model related to north-south striking fracture zones cutting through the Nova Canton trough, which is an extinct spreading center formed during the Cretaceous Normal Superchron (Larson et al., 2002). It shows that the central Pacific had been tectonically active until the late Cretaceous/early Cenozoic. This is supported by studies of two hotspot trails, the Gilbert Ridge and Tokelau seamounts northeast of the Manihiki Plateau (Koppers and Staudigel, 2005; Taylor, 2006a). They have revealed direction bends of the seamount trails at 67 Ma and 57 Ma and proposed an explanation of “jerk-like” plate extension, which reactivated a preconditioned lithosphere. Additionally, in the Danger Islands Troughs, en enchelon system separating the *High* Plateau and the Western Plateaus (Winterer et al., 1974b), which developed ~118 Ma, MCS data, (Pietsch and Uenzelmann-Neben, 2015) have shown that the Danger Islands Troughs had been tectonically active until the early Cenozoic. The Suvarov Trough forms a side branch of the Danger Islands Troughs and a stronger vertical displacement has been revealed towards their connection in the north. Therefore, we conclude that the formation of the Suvarov Trough displace a local phenomenon associated to a stress field induced from fracture zones north of the Manihiki Plateau.

5.2.2 Quaternay graben structure

Across the southern High Plateau, distinct and localized fault systems have been identified with vertical displacements of up to ~0.5 s TWT, which form a south-westward striking graben structure (Fig. 6, 7). We interpret those series of faults to have formed after the accumulation of reflection R1, as these vertical faults penetrate the entire sedimentary column. Therefore, they are of Quaternary age (< 1.8 Ma) and display a tectonic reactivation. Looking at the spatial distribution (Fig. 7), the Quaternary graben structures can be clustered into groups that (i) appear to be a prolongation of the Suvarov Trough and (ii) to be located close to the identified south-eastern diapir locations. We can directly conclude that due to the induced stress fields, both regions built weakness zones of the LIP crust, which got deformed by horizontal stress fields within the Pacific lithospheric plate (Clouard and Gerbault, 2008).

5.2.3 Is tectonic reactivation induced from hotspot volcanism?

We have shown in the previous discussion sections that the identified acoustic diapirs are images of Tertiary hotspot volcanism. Adjacent to a few of these intrusion centers e.g. CDP 6950-7100 in profile AWI-20120201 (Fig. 3), vertical tectonic faults occur, which have been interpreted to be induced magmatism. Most of the intrusion centers do not show adjacent induced tectonism. However, in the south-eastern part of the High Plateau, a series of faults occur in profile Kiwi07 (supplementary material), which penetrate the entire sedimentary column and are closed to the identified acoustic diapirs. A tectonic reactivation within this region is not astonishing as its base is formed by a Cretaceous volcano with an overprint of the Tertiary hotspot volcanism, which formed an extremely weak zone. We therefore conclude that the tectonic reactivation in the south-eastern part of the High Plateau was induced stress fields from hotspot volcanism.

In contrast, the major graben structures, e.g. CDP 1100-1500 of profile AWI-20120010 (Fig. 6) and CDP 6550-6700 (Fig. 3) in the south-western High Plateau are of Quaternary age. Comparing their age and spatial distribution to the late stage magmatism, a clear discrepancy can be seen. The trend of the Tertiary magmatism is from West to East, whereas the graben structure appears to be a prolongation from the Suvarov Trough from Northwest to Southeast. We therefore conclude that the graben structure is more closely related to a reactivation and extension of the existent NNW-SSE striking Suvarov Trough and is not directly linked to hotspot volcanism of the Society Islands or Tahiti hotspots.

6. Conclusion

In order to solve the question about the absence of hotspot tracks in the central Pacific, we have investigated the Manihiki Plateau, a Large Igenous Province, using single- and multi-channel seismic reflection data of Kiwi12 (Ai et al., 2008) and So224 (Uenzelmann-Neben, 2012a). We focused on the High Plateau, where the study of (Wessel and Kroenke, 2008) interpolated a crossing of three hotspot tracks; the Pitcairn Hotspot track (57-42 Ma) from north to south-east, the Society Islands (22-15 Ma) and Tahiti Hotspot (20-14 Ma) from West to East in the southern High Plateau. Acoustic diapirs could be identified and correlated to age of DSDP Leg 33 Site 317 (Pietsch and Uenzelmann-Neben, 2015; The Shipboard Scientific Party, 1976). Those Tertiary intrusion and extrusion centers are localized in two distinct regions, the northern High Plateau and the southern High Plateau. In the central High Plateau acoustic diapirs are absent.

The northern High Plateau is characterized by an appearance of several volcanoes, like the Rakahang or Manihiki Atolls, seen in the bathymetry, which are located along the supposed Pitcairn hotspot track (Wessel and Kroenke, 2008). In the seismic reflection data tectonic and magmatic features could be identified, which penetrate or disturb Unit 3, the volcanic base associated with the initial LIP formation. For the most prominent acoustic diapir an age of at least < 65 Ma could be identified. Generally, an age determination was difficult due to erosional unconformities and a lack of dated dredge samples. Summarizing, there is some evidence of hotspot volcanism, which needs to be studies in detail to understand the interaction of volcanism associated to the LIP formation and the later hotspot volcanism of the Pitcairn hotspot.

In the southern region, a total number of 19 acoustic diapirs could be identified. They show i) an extrusive or intrusive seismic characteristic ii) an exclusive occurrence south and parallel of the interpolated hotspot track of (Wessel and Kroenke, 2008), ii) grouping in the South-west and South-east and iii) average ages of 10-23 Ma with a small rejuvenation towards the East. As these characteristics fit to the assumed criteria of the supposed hotspot volcanism, we conclude that the acoustic diapirs are the images of hotspot volcanism within a Large Igneous Province. Additionally, the exclusively occurrence in the southern High Plateau can be explained by fractured and stretched LIP crust, associated to the break-up of the Hikurangi Plateau (Hochmuth et al., 2015, in prep., Pietsch and Uenzelmann-Neben, 2015), which eases the way for a magmatic ascend.

The Suvarov Trough, a graben structure, shows an increased vertical displacement towards the Danger Islands Troughs. It has formed after the initial LIP formation (< 65 Ma) by stretching of the LIP crust in southwestern direction. The graben extends to the south-east across the High Plateau, where it exhibits a Quaternary graben structure. We proposed a formation model that included a local stress field related plate extension and reactivation of a preconditioned lithosphere due to north-south striking fracture zones crossing the Nova Canton Trough north of the Manihiki Plateau.

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References

Ai, H.-A., Stock, J., Clayton, R., Luyendyk, B., 2008. Vertical tectonics of the High Plateau region, Manihiki Plateau, Western Pacific, from seismic stratigraphy. Marine Geophysical Researches 29, 13-26.

Ballmer, M.D., Conrad, C.P., Smith, E.I., Harmon, N., 2013. Non-hotspot volcano chains produced by migration of shear-driven upwelling toward the East Pacific Rise. Geology 41, 479-482.

Beiersdorf, H., Bach, W., Duncan, R., Erzinger, J., Weiss, W., 1995a. New Evidence for the Production of Em-Type Ocean Island Basalts and Large Volumes of Volcaniclastites during the Early History of the Manihiki Plateau. Marine Geology 122, 181-205.

Beiersdorf, H., Bach, W., Duncan, R., Erzinger, J., Weiss, W., 1995b. New evidence for the production of EM-type ocean island basalts and large volumes of volcaniclastites during the early history of the Manihiki Plateau. Marine Geology 122, 181-205.

Billen, M.I., Stock, J., 2000a. Morphology and origin of the Osbourn Trough. Journal of Geophysical Research: Solid Earth 105, 13481-13489.

Billen, M.I., Stock, J., 2000b. Morphology and origin of the Osbourn Trough. J. Geophys. Res. 105, 13481-13489.

Bloomer, S.H., Taylor, B., MacLeod, C.J., Stern, R.J., Fryer, P., Hawkins, J.W., Johnson, L., 1995. Early arc volcanism and the ophiolite problem: a perspective from drilling in the western Pacific. Active margins and marginal basins of the western Pacific, 1-30.

Chandler, M.T., Wessel, P., Taylor, B., Seton, M., Kim, S.-S., Hyeong, K., 2012. Reconstructing Ontong Java Nui: Implications for Pacific absolute plate motion, hotspot drift and true polar wander. Earth and Planetary Science Letters 331/332, 140-151.

Clayton, R.W., 1998. KIWI 12. CalTech, p. 12.

Clouard, V., Bonneville, A., 2001. How many Pacific hotspots are fed by deep-mantle plumes? Geology 29, 695-698.

Clouard, V., Bonneville, A., Gillot, P.Y., 2003. The Tarava Seamounts: a newly characterized hotspot chain on the South Pacific Superswell. Earth and Planetary Science Letters 207, 117-130.

Clouard, V., Gerbault, M., 2008. Break-up spots: Could the Pacific open as a consequence of plate kinematics? Earth and Planetary Science Letters 265, 195-208.

Coffin, M.F., Eldholm, O., 1993. Scratching the surface: Estimating the dimensions of large igneous provinces. Geology 21, 515-518.

Coffin, M.F., Werner, R., Hauff, F., Hoernle, K., Party, S.S., 2007. Extensional and Transtensional Tectonics of the Manihiki Plateau, Western Equatorial Pacific Ocean, AGU Fall Meeting, San Francisco, USA.

Davy, B., Hoernle, K., Werner, R., 2008. Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history. Geochemistry Geophysics Geosystems 9.

Downey, N.J., Stock, J.M., Clayton, R.W., Cande, S.C., 2007. History of the Cretaceous Osbourn spreading center. J. Geophys. Res. 112, B04102.

Eissen, J.-P., Crawford, A.J., Cotten, J., Meffre, S., Bellon, H., Delaune, M., 1998. Geochemistry and tectonic significance of basalts in the Poya Terrane, New Caledonia. Tectonophysics 284, 203-219.

Gurnis, M., Hall, C., Lavier, L., 2004. Evolving force balance during incipient subduction. Geochemistry, Geophysics, Geosystems 5.

Hauff, F., Hoernle, K., Werner, R., van den Bogaard, P., Timm, C., Klügel, K., 2008. Geochemistry and age data from the Manihiki Plateau (R/V SO193): multiple volcanic events from EM1-FOZO-HIMU mantle sources within a single oceanic LIP, AGU Fall Meeting, San Francisco, California, USA.

Heezen, B.C., Glass, B., Menard, H.W., 1966. The Manihiki Plateau. Deep Sea Research and Oceanographic Abstracts 13, 445-458.

Hieronymus, C.F., Bercovici, D., 2000. Non-hotspot formation of volcanic chains: control of tectonic and flexural stresses on magma transport. Earth and Planetary Science Letters 181, 539-554.

Hoernle, K., Hauff, F., van den Bogaard, P., Werner, R., Mortimer, N., Geldmacher, J., Garbe-Schonberg, D., Davy, B., 2010. Age and geochemistry of volcanic rocks from the Hikurangi and Manihiki oceanic Plateaus. Geochimica Et Cosmochimica Acta 74, 7196-7219.

Hoernle, K., Hauff, F., Werner, R., Mortimer, N., 2004. New Insights into the Origin and Evolution of the Hikurangi Oceanic Plateau (Southwest Pacific) from Multi-beam Mapping and Sampling. EOS Transactions 41, 401-408.

Hoernle, K., Hauff, F., Werner, R., van den Bogaard, P., Timm, C., Coffin, M.F., Mortimer, N., Davy, B., 2008. A similar multi-stage geochemical evolution for the Manihiki, Hikurangi and Ontong Java Plateaus?, AGU Fall Meeting, San Francisco, California, USA.

Hussong, D.M., Wipperman, L.K., Kroenke, L.W., 1979. Crustal Structure of the Ontong Java and Manihiki Oceanic Plateaus. Journal of Geophysical Research 84, 6003-6010.

Ingle, S., Coffin, M.F., 2004. Impact origin for the greater Ontong Java Plateau? Earth Planet Sci Lett 218, 123-134.

Ingle, S., Mahoney, J.J., Sato, H., Coffin, M.F., Kimura, J.-I., Hirano, N., Nakanishi, M., 2007. Depleted mantle wedge and sediment fingerprint in unusual basalts from the Manihiki Plateau, central Pacific Ocean. Geology 35, 595-598.

Klarner, S., Klarner, O., 2012. Identification of Paleo-Volcanic Rocks on Seismic Data, in: Stoppa, P.F. (Ed.), Updates in Volcanology - A Comprehensive Approach to Volcanological Problems. InTech.

Koppers, A.A.P., Staudigel, H., 2005. Asynchronous bends in Pacific seamount trails: A case for extensional volcanism? Science 307, 904-907.

Kroenke, L.W., 1984. Cenozoic tectonic development of the Southwest Pacific / by Loren W. Kroenke ; with a contribution by Peter Rodda. CCOP/SOPAC, [Suva, Fiji].

Lanphere, M.A., Dalrymple, G.B., 1976. K-AR Ages of basalts from DSDP Leg 33: Sites 315 (Line Island) and 317 (Manihiki Plateau), in: Kaneps, A.G. (Ed.), Initial Reports of the Deep Sea Drilling Project. US Governmental Printing Office, Washington, D. C., pp. 649-653.

Larson, R.L., Erba, E., 1999. Onset of the mid-Cretaceous greenhouse in the Barremian-Aptian: Igneous events and the biological, sedimentary, and geochemical responses. Paleoceanography 14, 663-678.

Larson, R.L., Pockalny, R.A., Viso, R.F., Erba, E., Abrams, L.J., Luyendyk, B.P., Stock, J.M., Clayton, R.W., 2002. Mid-Cretaceous tectonic evolution of the Tongareva triple junction in the southwestern Pacific Basin. Geology 30, 67-70.

Magee, C., Hunt-Stewart, E., Jackson, C.A.L., 2013. Volcano growth mechanisms and the role of sub-volcanic intrusions: Insights from 2D seismic reflection data. Earth and Planetary Science Letters 373, 41-53.

Mahoney, J.J., Storey, M., Duncan, R.A., Spencer, K.J., Pringle, M., 1993. Geochemistry and Age of the Ontong Java Plateau, The Mesozoic Pacific: Geology, Tectonics, and Volcanism. American Geophysical Union, pp. 233-261.

Morgan, W.J., 1972. Deep Mantle Convection Plumes and Plate Motions. American Association of Petroleum Geologists Bulletin 56, 203-&.

Neall, V.E., Trewick, S.A., 2008. The age and origin of the Pacific islands: a geological overview. Philosophical Transactions of the Royal Society B-Biological Sciences 363, 3293-3308.

Pietsch, R., Uenzelmann-Neben, G., 2015. The Manihiki Plateau - a multistage volcanic emplacement history. Earth and Planetary Science Letters.

Planke, S., Rasmussen, T., Rey, S.S., Myklebust, R., 2005. Seismic characteristics and distribution of volcanic intrusions and hydrothermal vent complexes in the Vøring and Møre basins. Geological Society, London, Petroleum Geology Conference series 6, 833-844.

Planke, S., Symonds, P.A., Alvestad, E., Skogseid, J., 2000. Seismic volcanostratigraphy of large-volume basaltic extrusive complexes on rifted margins. Journal of Geophysical Research-Solid Earth 105, 19335-19351.

Seton, M., Müller, R.D., Zahirovic, S., Gaina, C., Torsvik, T., Shephard, G., Talsma, A., Gurnis, M., Turner, M., Maus, S., Chandler, M., 2012. Global continental and ocean basin reconstructions since 200 Ma. Earth-Science Reviews 113, 212-270.

Shipboard_Scientific_Party, 1976. Site 317, in: Schlanger, S.O., Winterer, E.L. (Eds.), Initial Reports. Deep Sea Drilling Project, Washington, D.C., pp. 161-300.

Stock, J.M., B.P., L., R.W., C., 1998. Tectonics and structure of the Manihiki Plateau, western Pacific Ocean. . Eos Trans Am Geophys Union 79:F870.

Taylor, B., 2006a. The single largest oceanic plateau: Ontong Java-Manihiki-Hikurangi. Earth and Planetary Science Letters 241, 372-380.

Taylor, B., 2006b. The single largest oceanic plateau: Ontong Java–Manihiki–Hikurangi. Earth and Planetary Science Letters 241, 372-380.

The Shipboard Scientific Party, 1976. Site 317, in: Kaneps, A.G. (Ed.), Initial Reports of the Deep Sea Drilling Project. US Governmental Printing Office, Washington, D.C., pp. 161-300.

Timm, C., Hoernle, K., Werner, R., Hauff, F., den Bogaard, P.v., Michael, P., Coffin, M.F., Koppers, A., 2011. Age and geochemistry of the oceanic Manihiki Plateau, SW Pacific: New evidence for a plume origin. Earth and Planetary Science Letters 304, 135-146.

Uenzelmann-Neben, G., 2012a. The Expedition of the Research Vessel "Sonne" to the Manihiki Plateau in 2012 (SO 224), in: Uenzelmann-Neben, G. (Ed.), Reports on Polar and Marine Research. Alfred-Wegener-Institut für Polar- und Meeresforschung in der Helmholtz Gemeinschaft, Bremerhaven.

Uenzelmann-Neben, G., 2012b. The expedition of the research vessel "Sonne" to the Manihiki Plateau in 2012 (So 224). Berichte zur Polar- und Meeresforschung = Reports on polar and marine research 656, 70.

Uenzelmann-Neben, G., Gohl, K., Ehrhardt, A., Seargent, M., 1999. Agulhas Plateau, SW Indian Ocean: New evidence for excessive volcanism. Geophysical research letters 26, 1941-1944.

Werner, R., Hauff, F., 2007. FS Sonne Cruise Report SO 193: MANIHIKI Temporal, Spatial, and Tectonic Evolution of Oceanic Plateaus, IFM-GEOMAR Report. IFM-Geomar, Kiel, p. 201.

Werner, R., Hauff, F., Hoernle, K., Coffin, M.F., Party, S.S., 2007. R/V Sonne Cruise SO193: New Insights Into the Geodynamic History of the Manihiki Plateau, SW-Pacific, AGU Fall Meeting, San Francisco.

Wessel, P., Kroenke, L.W., 2008. Pacific absolute plate motion since 145 Ma: An assessment of the fixed hot spot hypothesis. Journal of Geophysical Research-Solid Earth 113.

Winterer, E., Lonsdale, P.L., Matthews, J.L., Rosendahl, B.R., 1974a. Structure and acoustic stratigraphy of the Manihiki Plateau. Deep-Sea research I 21, 793-814.

Winterer, E.L., Lonsdale, P.F., Matthews, J.L., Rosendah.Br, 1974b. Structure and Acoustic Stratigraphy of Manihiki Plateau. Deep-Sea Research 21, 793-&.

Worthington, T.J., Hekinian, R., Stoffers, P., Kuhn, T., Hauff, F., 2006. Osbourn Trough: Structure, geochemistry and implications of a mid-Cretaceous paleospreading ridge in the South Pacific. Earth and Planetary Science Letters 245, 685-701.

Figure captions

Fig. 1 (a) Hotspot tracks in the Pacific Ocean on top of GEBCO bathymetry with most recent hotspot centers (white stars). Red box marks investigation area. (b) Enlargement of the investigation area, where the Manihiki Plateau is located. Interpolated hotspot tracks of WK08-A (Wessel and Kroenke, 2008) are shown in dashed yellow line. Pitcairn, Society Island and Tahiti hotspot crossed the High Plateau between 57-42 Ma and 22/20-15/14 Ma, respectively. Star marks location of DSDP Leg 33 Site 317 borehole (The Shipboard Scientific Party, 1976). Seismic reflection profiles of Kiwi12 and So224 are marked in orange and red line, respectively.

Fig. 2 (a) Seismic section of profile Kiwi1211 in the northern part of the High Plateau and (b) bathymetry.

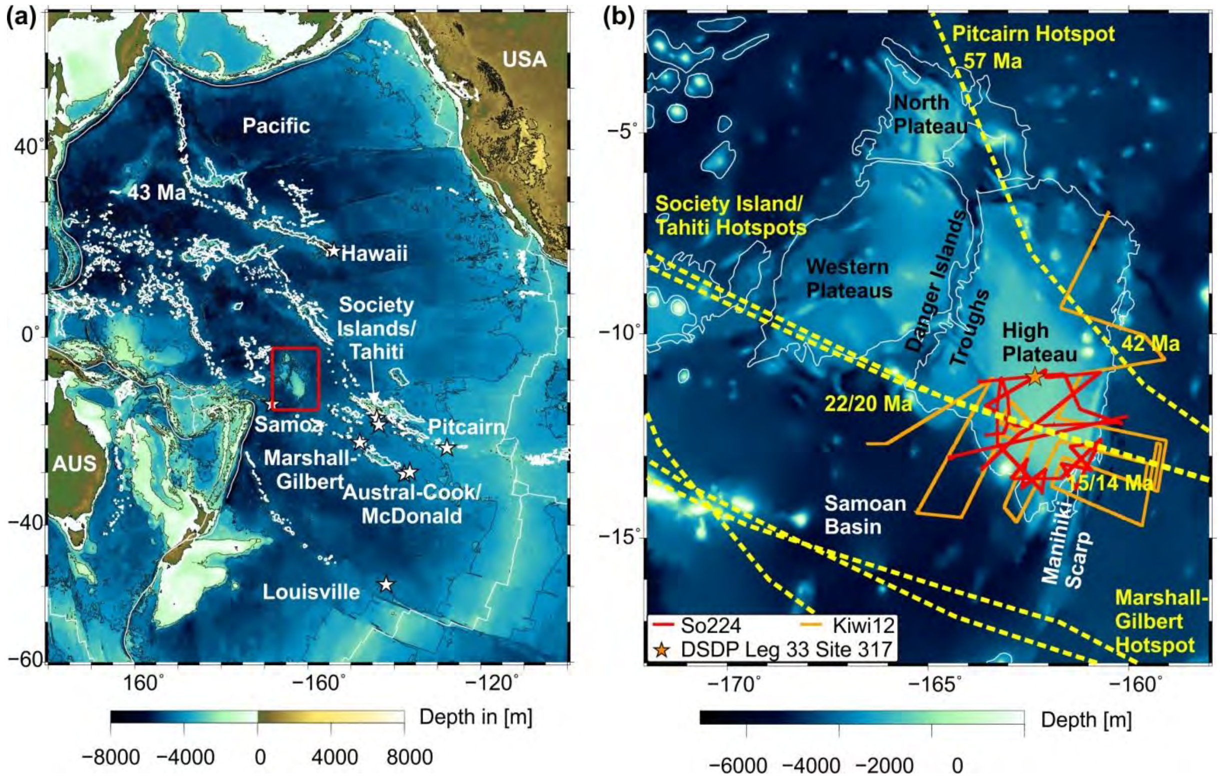
Fig. 3 Swath bathymetry and seismic profiles of AWI-20120201/-028/-023 and Kiwi1204 in the south-western part of the High Plateau.

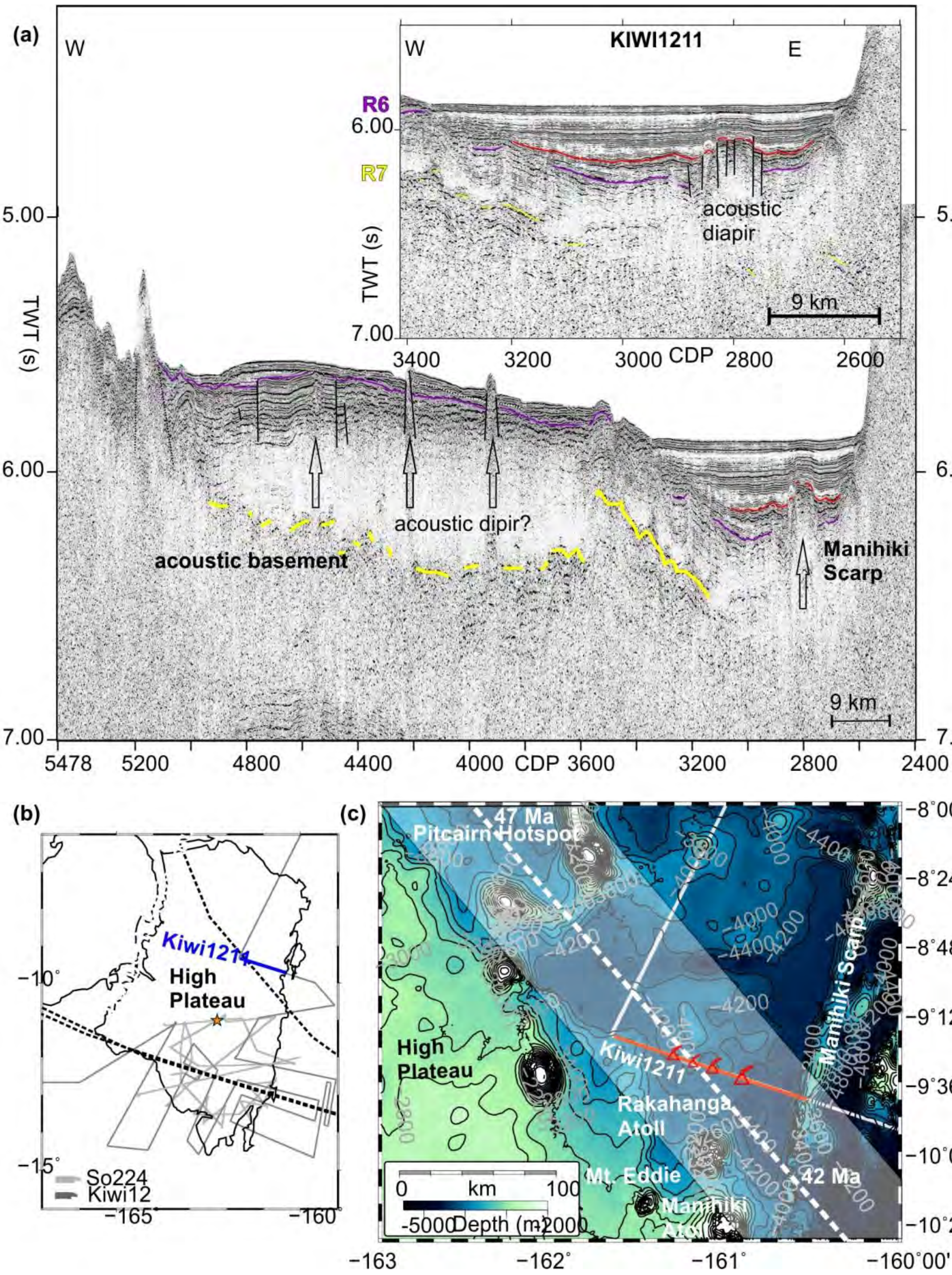
Fig. 4 Swath bathymetry and seismic profiles of AWI-20120013/-016 and Kiwi1206 in the south-eastern part of the High Plateau.

Fig. 5 Bathymetry (a) and (b-d) MCS profiles AWI-20120006 and Kiwi02 across the Suvarov Trough.

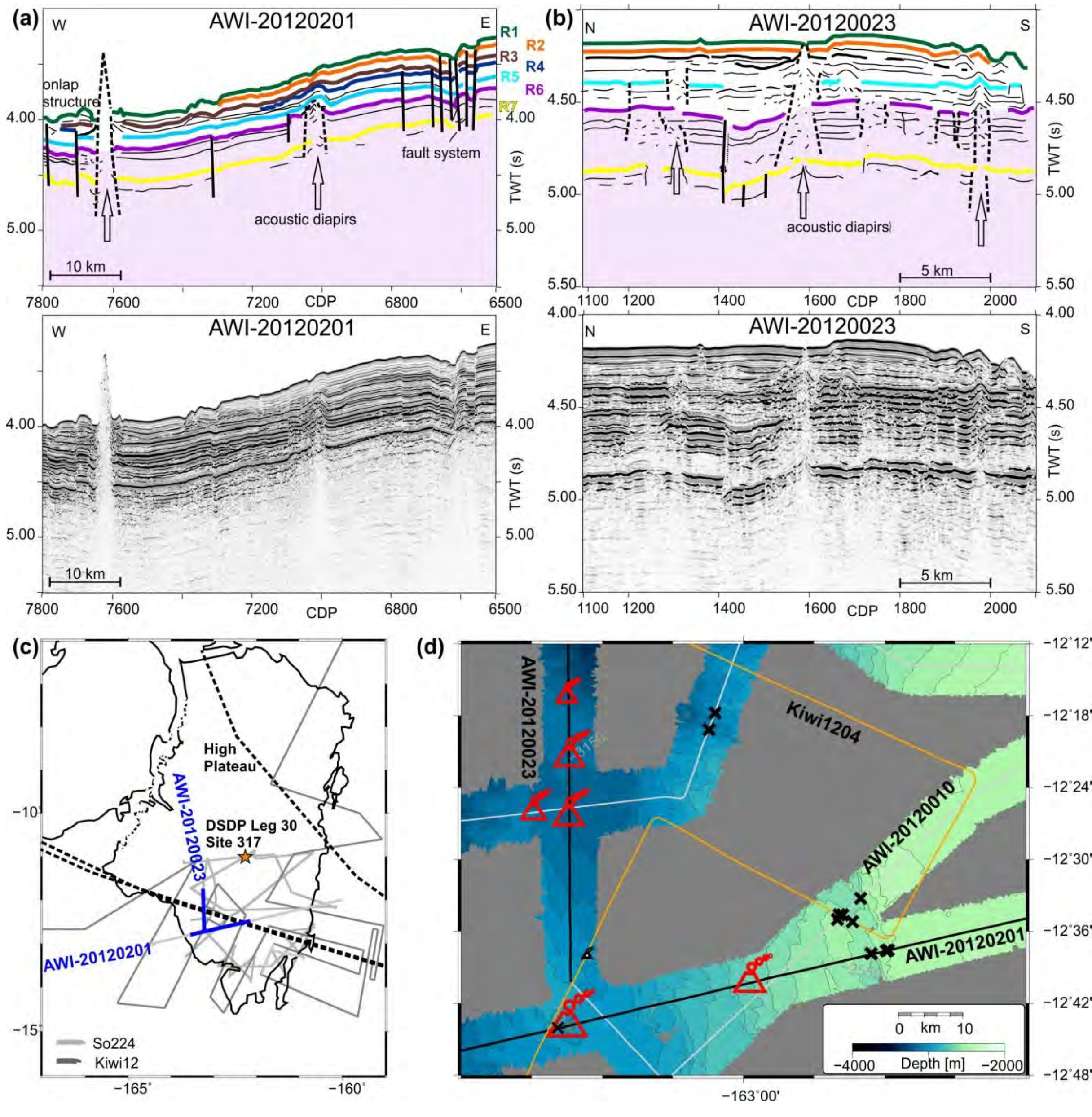
Fig. 10 Seismic profile AWI-2012010 across the fault system in the southern High Plateau. Bathymetry is shown in Fig. 3.

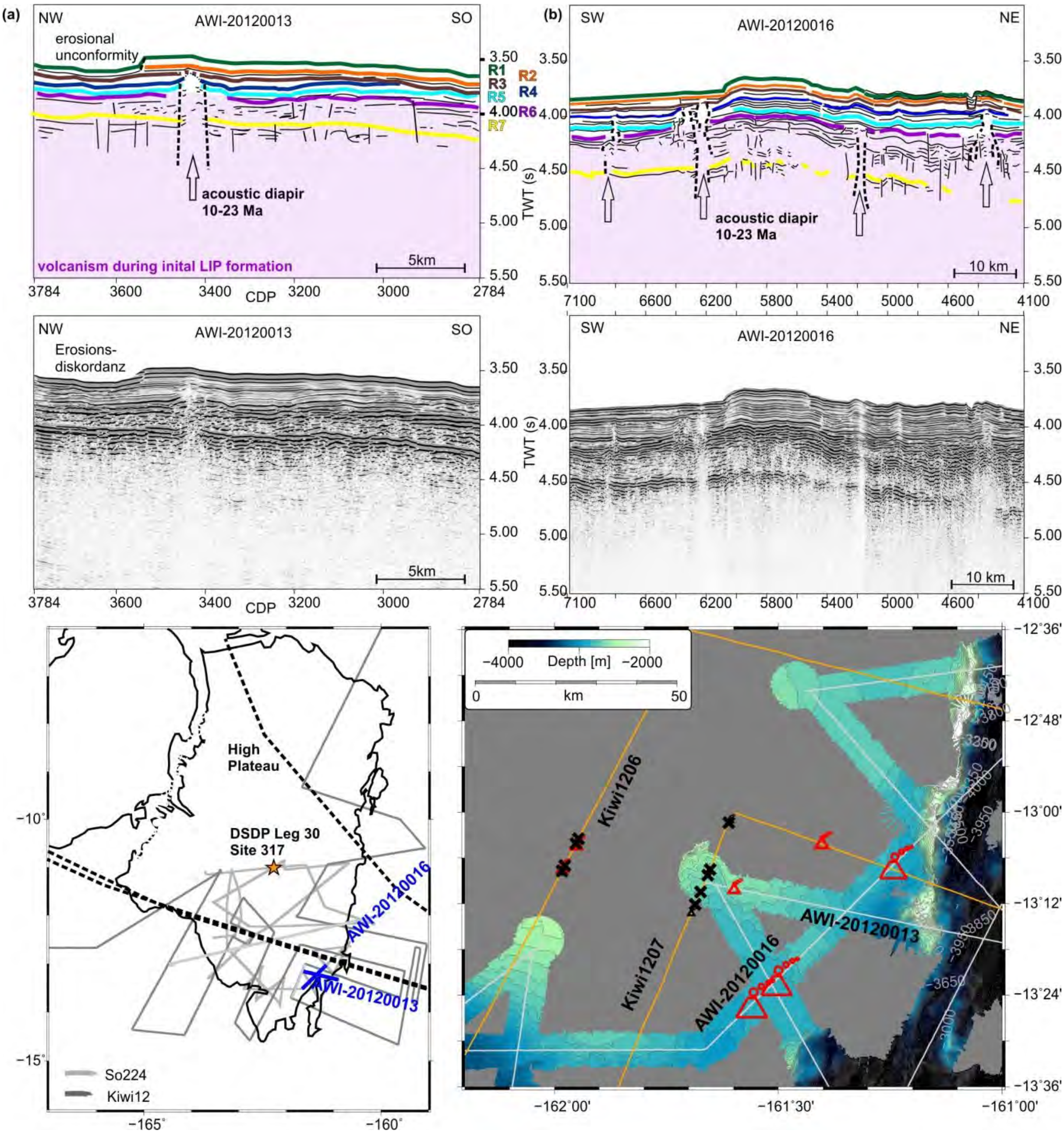
Fig. 7 (a) Spatial distribution of identified intrusion centers across the High Plateau, subplateau of the Manihiki Plateau. Interpolated hotspot tracks of absolute plate motion model WK08-A (Wessel and Kroenke, 2008) with 100 km span are shown in yellow. (b) Spatial distribution of Quaternary graben structure on top of bathymetry.

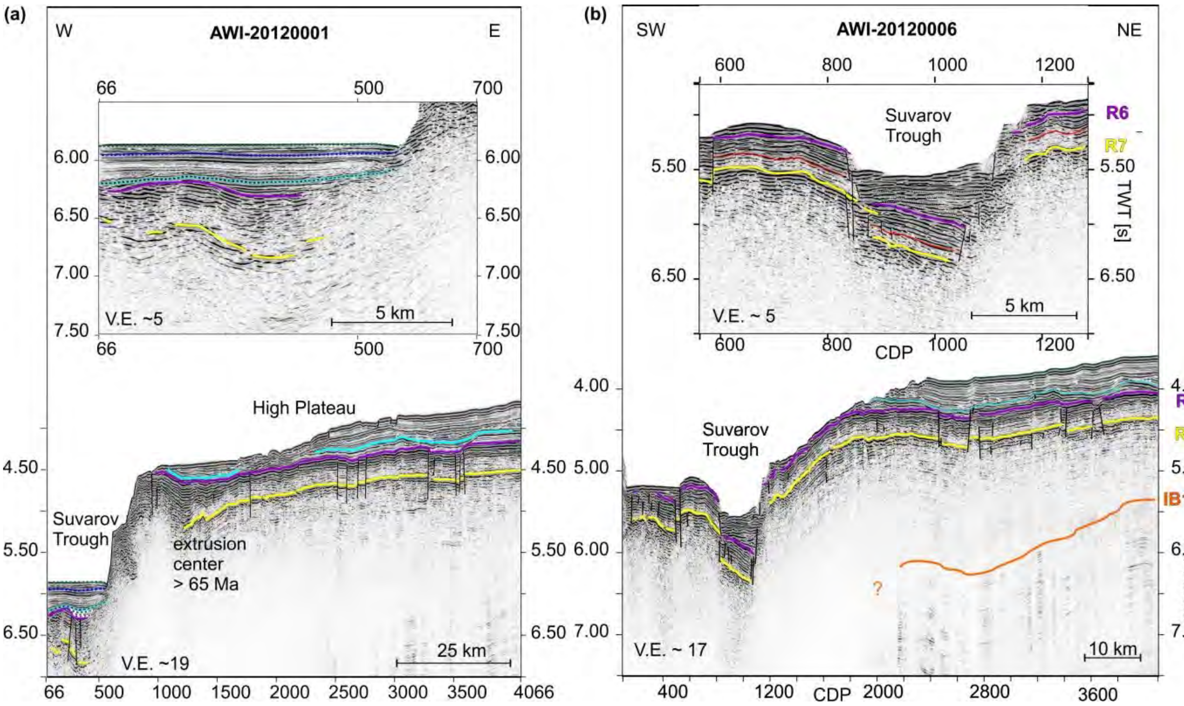

(a)
USA
Pacific
40°
43 Ma
Hawaii
Society
Islands/
Tahiti
0°
Samoa
Pitcairn
AUS
Marshall-
Gilbert
Austral-Cook/
McDonald
−40
Louisville
−60
160°
−160°
−120°
Depth in [m]
−8000 −4000 0 4000 8000
(b)
Pitcairn Hotspot
57 Ma
North
Plateau
−5°
Society Island/
Tahiti Hotspots
Western
Plateaus
Danger Islands
Troughs
High
Plateau
−10°
42 Ma
22/20 Ma
15/14 Ma
Samoan
Basin
−15°
Manihiki
Scarp
Marshall-
Gilbert
Hotspot
So224
Kiwi12
DSDP Leg 33 Site 317
−170°
−165°
−160°
Depth [m]
−6000 −4000 −2000 0

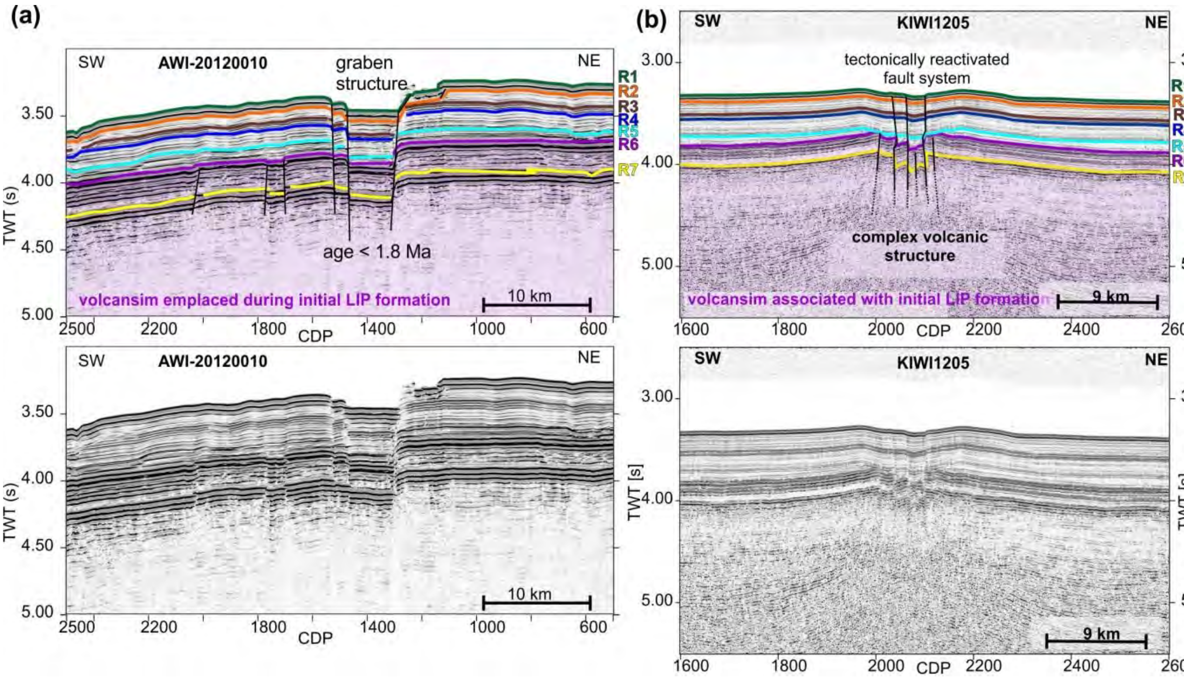


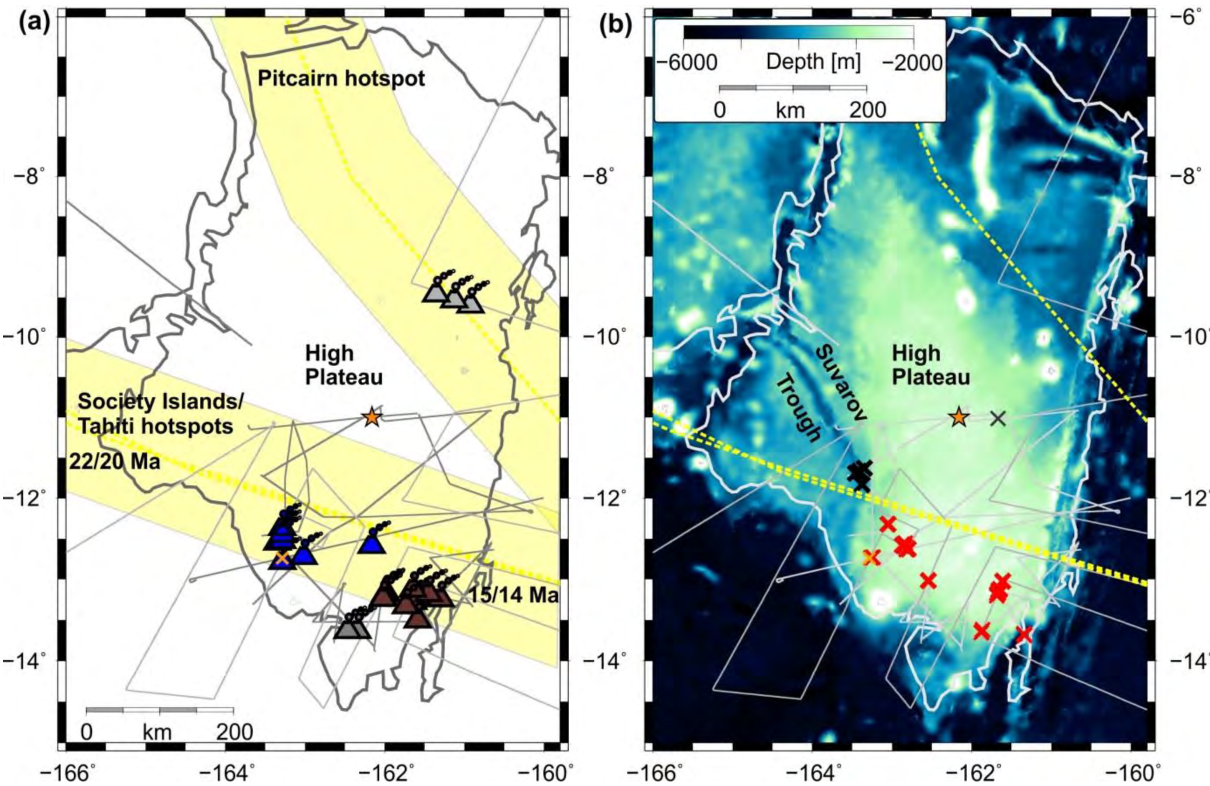
(a)
W
KIWI1211
W
E
R6
R7
acoustic diapir
TWT (s)
9 km
CDP
acoustic dipir?
acoustic basement
Manihiki Scarp
(b)
Kiwi1211
High Plateau
So224
Kiwi12
(c)
47 Ma
Pitcairn Hotspot
High Plateau
Kiwi1211
Rakahanga Atoll
Manihiki Scarp
Mt. Eddie
Manihiki Atoll
42 Ma
km
Depth (m)

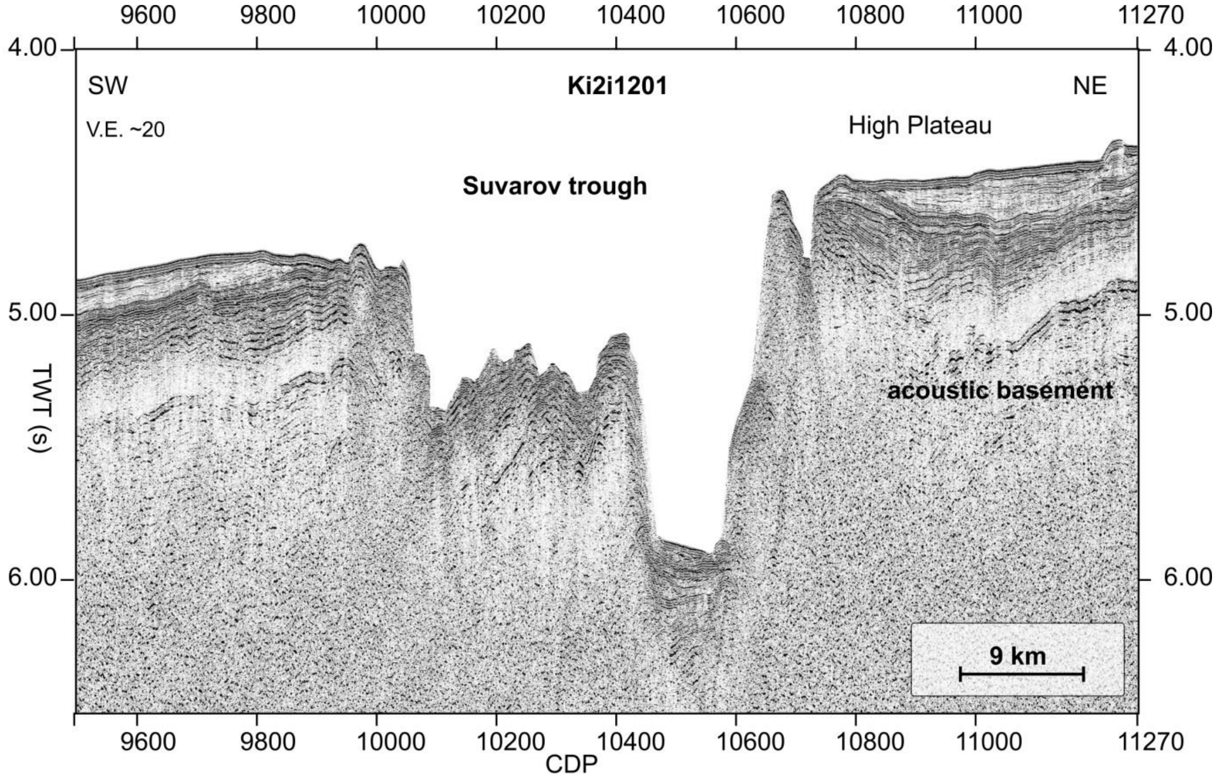

(a)
W
AWI-20120201
E
R1
R2
R3
R4
R5
R6
R7
onlap structure
fault system
acoustic diapirs
10 km
TWT (s)
CDP
(b)
N
AWI-20120023
S
acoustic diapirs
5 km
(c)
High Plateau
AWI-20120023
DSDP Leg 30 Site 317
AWI-20120201
So224
Kiwi12
(d)
AWI-20120023
Kiwi1204
AWI-20120010
AWI-20120201
0 km 10
Depth [m]
-4000
-2000
-163°00'

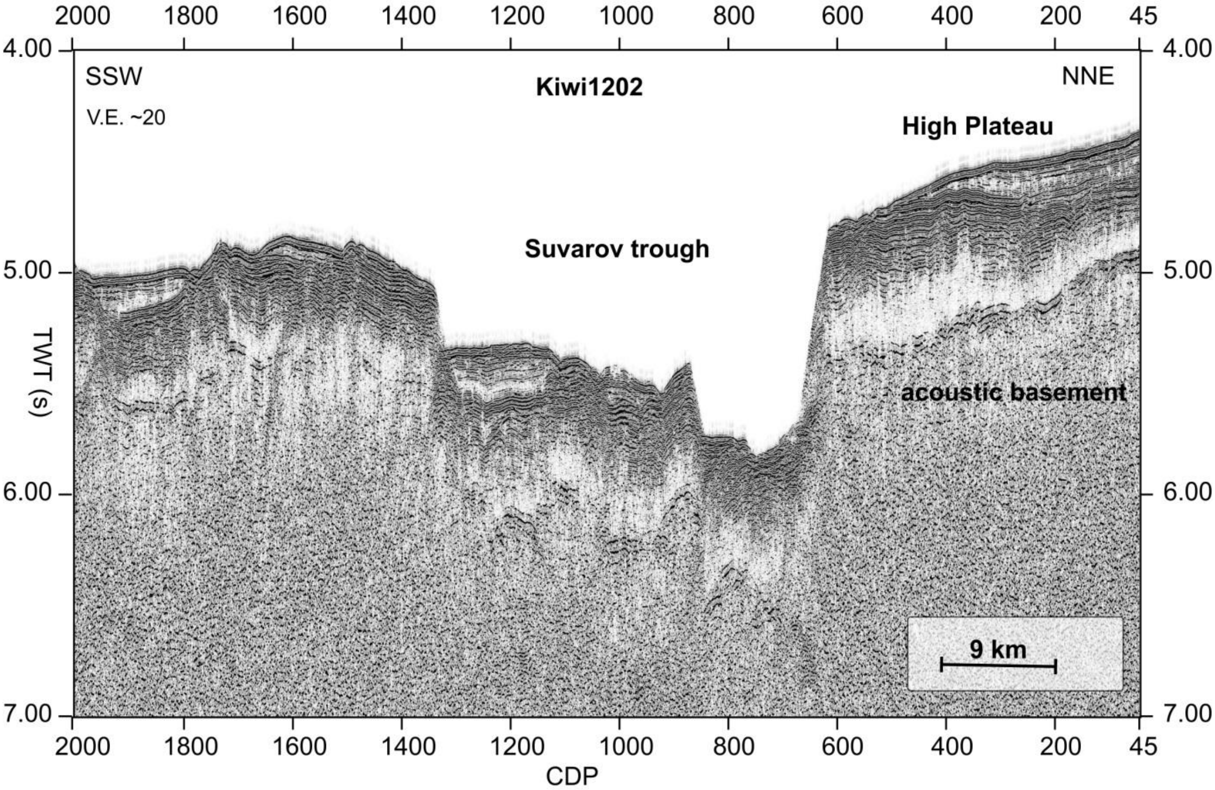

(a)
NW
AWI-20120013
SO
erosional unconformity
R1
R2
R3
R4
R5
R6
R7
acoustic diapir 10-23 Ma
volcanism during inital LIP formation
5km
CDP
(b)
SW
AWI-20120016
NE
TWT (s)
acoustic diapir 10-23 Ma
10 km
Erosions-diskordanz
High Plateau
DSDP Leg 30 Site 317
AWI-20120016
AWI-20120013
So224
Kiwi12
Depth [m]
km
Kiwi1206
Kiwi1207


(a)
W
AWI-20120001
E
V.E. ~5
5 km
High Plateau
Suvarov Trough
extrusion center > 65 Ma
V.E. ~19
25 km
(b)
SW
AWI-20120006
NE
Suvarov Trough
R6
R7
TWT [s]
V.E. ~ 5
5 km
CDP
Suvarov Trough
IB1
?
TWT (s)
V.E. ~ 17
10 km
CDP


(a)
SW
AWI-20120010
NE
graben structure
R1
R2
R3
R4
R5
R6
R7
TWT (s)
age < 1.8 Ma
volcansim emplaced during initial LIP formation
10 km
CDP
SW
AWI-20120010
NE
TWT (s)
10 km
CDP
(b)
SW
KIWI1205
NE
tectonically reactivated fault system
R1
R2
R3
R4
R5
R6
R7
complex volcanic structure
volcansim associated with initial LIP formation
9 km
CDP
SW
KIWI1205
NE
TWT [s]
TWT [s]
9 km
CDP


(a)
Pitcairn hotspot
High Plateau
Society Islands/ Tahiti hotspots
22/20 Ma
15/14 Ma
0 km 200
−8°
−10°
−12°
−14°
−166°
−164°
−162°
−160°
(b)
−6000 Depth [m] −2000
0 km 200
Suvarov Trough
High Plateau
−6°
−8°
−10°
−12°
−14°
−166°
−164°
−162°
−160°


9600
9800
10000
10200
10400
10600
10800
11000
11270
4.00
5.00
6.00
SW
Ki2i1201
NE
V.E. ~20
High Plateau
Suvarov trough
acoustic basement
TWT (s)
9 km
CDP


2000
1800
1600
1400
1200
1000
800
600
400
200
45
4.00
5.00
6.00
7.00
SSW
Kiwi1202
NNE
V.E. ~20
High Plateau
Suvarov trough
acoustic basement
TWT (s)
9 km
CDP

BESCHEINIGUNGEN DES BSH UND PANGAEA - ANLAGE 6

BUNDESAMT FÜR SEESCHIFFFAHRT UND HYDROGRAPHIE

BSH · Neptunallee 5 · 18057 Rostock

Dr. Gabriele Uenzelmann-Neben
Alfred-Wegener-Institut für Polar-
und Meeresforschung
Postfach 12 01 61
27515 Bremerhaven

Dienstsitz Rostock

Datum
10.12.2012
Durchwahl
+ 49 (0) 381 4563 - 604
Aktenzeichen
(bitte bei Antwort angeben)
3127 / 2012 N1130

Bathymetriedaten der Reise SO-224
- Eingangsbestätigung

Sehr geehrte Frau Dr. Uenzelmann-Neben,

die bathymetrischen Daten der Reise SO-224 und die dazugehörende Metadatendokumentation waren am 4.12.2012 bei uns eingegangen.

Vielen Dank.

Mit freundlichen Grüßen,

Volkmar Leimer

Neptunallee 5
18057 Rostock
Tel. + 49 (0) 381 4563 – 5
Fax. + 49 (0) 381 4563 – 948
posteingang@bsh.de
www.bsh.de

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Data Description

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Citation: **Uenzelmann-Neben, Gabriele (2014):** Profile of sediment echo sounding during SONNE cruise SO224 with links to ParaSound data files. doi:10.1594/PANGAEA.829069

Related to: **Uenzelmann-Neben, Gabriele (2012):** The Expedition of the Research Vessel Sonne to the Manihiki Plateau in 2012 (SO 224). *Berichte zur Polar- und Meeresforschung = Reports on Polar and Marine Research*, **656**, 70 pp, doi:10.2312/BzPM_0656_2012

Project(s): **Paleoenvironmental Reconstructions from Marine Sediments @ AWI** (AWI_Paleo)

Coverage: *Median Latitude:* -11.711516 * *Median Longitude:* -162.749479 * *South-bound Latitude:* -13.882130 * *West-bound Longitude:* -168.018690 * *North-bound Latitude:* -7.209980 * *East-bound Longitude:* -159.960050

Date/Time Start: 2012-10-15T03:00:00 * *Date/Time End:* 2012-11-12T19:50:00

Event(s): **SO224-track** * *Latitude Start:* -18.128600 * *Longitude Start:* 178.423700 * *Latitude End:* -18.128600 * *Longitude End:* 178.423700 * *Date/Time Start:* 2012-10-09T00:00:00 * *Date/Time End:* 2012-11-18T00:00:00 * *Location:* Manihiki Plateau, Pacific Ocean * *Campaign:* SO224 (MANIHIKI II) * *Basis:* Sonne * *Device:* Underway cruise track measurements (CT) * *Comment:* Suva - Suva

Comment: File format description see: Data Acquisition of ATLAS PARASTORE, Atlas Hydrographic (2007), Operator Manual ATLAS PARASTORE, Doc. Id.: ED 6006 G 212:/Version: 4.0/Edition: 05/2007, hdl:10013/epic.35589.d001. ASD Online data file: http://hs.pangaea.de/para/so224/ASD_Online/ASD_Online.tar; Navigation track data file: http://hs.pangaea.de/para/so224/DSHIP/DSHIP.tar; Auxilary data file: http://hs.pangaea.de/para/so224/PSAUX/PSAUX.tar

Parameter(s):

#	Name	Short Name	Unit	Principal Investigator	Method	Comment
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2	LATITUDE	Latitude				Geocode
3	LONGITUDE	Longitude				Geocode
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5	Uniform resource locator/link to raw data file	URL raw		Uenzelmann-Neben, Gabriele		PHF_PS3 files in tar-archive
6	Uniform resource locator/link to raw data file	URL raw		Uenzelmann-Neben, Gabriele		PHF_SGY files in tar-archive
7	Uniform resource locator/link to raw data file	URL raw		Uenzelmann-Neben, Gabriele		SLF_ASD files in tar-archive
8	Uniform resource locator/link to raw data file	URL raw		Uenzelmann-Neben, Gabriele		SLF_PS3 files in tar-archive
9	Uniform resource locator/link to raw data file	URL raw		Uenzelmann-Neben, Gabriele		SLF_SGY files in tar-archive

Size: 20970 data points

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