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Pacific seamount volcanoism in space and time

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SUMMARY
Seamounts constitute some of the most direct evidence about intraplate volcanism. As such, when seamounts formed and into which tectonic setting they erupted (i.e. on-ridge or off-ridge) are a useful reflection of how the properties of the lithosphere interact with magma generation in the fluid mantle beneath. Proportionately few seamounts are radiometrically dated however, and these tend to be recently active.

In order to more representative sample and better understand Pacific seamount volcanoism this paper estimates the eruption ages \( t_{\text{volc}} \) of 2706 volcanoes via automated estimates of lithospheric strength. Lithospheric strength \( (GTR_{\text{rel}}) \) is deduced from the ratio of gravity to topography above the summits of volcanoes, and is shown to correlate with seafloor age at the time of volcanic loading \( (\Delta t) \) at 61 sites where radiometric constraints upon \( \Delta t \) exist. A trend of \( GTR_{\text{rel}} = 0.09825 \sqrt{\Delta t} + 0.20251 \) fits data for these 61, and with seafloor age \( (t_{\text{sf}}) \) known, can date the 2706 volcanoes; \( t_{\text{volc}} = t_{\text{sf}} - \Delta t \).

Widespread recurrences of volcanism proximal to older features (e.g. the Cook-Austral alignment in French Polynesia) suggest that the lithosphere exerts a significant element of control upon the location of volcanism, and that magmatic throughput leaves the lithosphere more susceptible to the passage of future melts. Observations also prompt speculation that: the Tavara seamounts share morphological characteristics and isostatic compensation state with the Musicians, and probably formed similarly; the Easter Island chain may be a modern analogy to the Cross-Lines; a Musicians – South Hawaiian seamounts alignment may be deflecting the Hawaiian hotspot trace.

Key words: flexural isostasy, lithosphere, pacific, seamount dating, volcanism.

1 INTRODUCTION
Beneath the world’s oceans, the seafloor is littered with relatively small-scale (up to ~250 km across) rises that are subcircular in plan view. By analogy with volcanic oceanic islands, and by direct sampling of a few hundred (i.e. <1 per cent; Wessel 1997), these submarine mountains or ‘seamounts’ are thought to be volcanic. They constitute some of the most direct evidence about geodynamic processes in and below the lithosphere that control intraplate volcanism.

Dating is a critical part of this evidence. For instance, the ‘hotspot’ hypothesis (Wilson 1963b; Morgan 1971), where linear volcanic chains on a moving tectonic plate may be traced back to a single approximately stationary region of high magma productivity, seeks to explain edifices of monotonically increasing age to the NW along the Hawaiian-Emperor seamount chain.

Up to 50 000 seamounts taller than 1 km are estimated to exist in the Pacific (Menard 1964; Wessel & Lyons 1997), 12 000 of which have been counted on maps (Batiza 1982). Unfortunately, only ~250 are radiometrically dated after direct sampling (Fig. 1), and these samples favour recent hotspot volcanism (39 per cent of ages are <10 Ma, 3.8 times greater than in any other 10 Ma interval), so the distribution of seamount volcanism in space and time is not representatively sampled.

An alternative, albeit indirect, technique to constrain seamount ages uses an elastic plate model of regional isostatic compensation to reconcile the relationship between gravity and bathymetry. The plate rigidity required (commonly quantified as plate thickness, \( T_p \)) systematically increases with lithospheric age at the time of loading, \( \Delta t \) (Watts 1978; Watts et al. 1980; Calmant et al. 1990; Watts 2001). Since lithospheric age in the proximity of a seamount, \( t_{\text{sf}} \), is relatively well dated by magnetic lineations in the ridge-produced oceanic crust, rigidity can be related to \( \Delta t \) and edifice age, \( t_{\text{volc}} \); \( t_{\text{volc}} = t_{\text{sf}} - \Delta t \).

Watts et al. (1980) constrained the tectonic setting of volcanic emplacement for ~130 seamounts to be ‘on-ridge’ (\( \Delta t < 8 \) Ma) or ‘off-ridge’ (\( \Delta t > 35 \) Ma). Craig & Sandwell (1988) estimated rigid- ity from nine Seasat altimeter profiles. Calmant et al. (1990) first proposed quantitatively inverting the relationship between rigidity and \( \Delta t \), fitting \( T_p = 2.7 \pm 0.15\sqrt{\Delta t} \) to 36 data, but only dating the
2 METHOD

An elastic plate model of regional (i.e. flexural) isostatic compensation (e.g. Vening Meinesz 1941; Gunn 1943; Watts & Daly 1981, Fig. 2) well reconciles the relationship between gravity and bathymetry over seamounts (e.g. Watts 1978; Freedman & Parsons 1986). With increasing plate thickness the ratio of gravity to seamount height (also called topography) above the centre of a seamount rises. G is gravity, T is topography, and the ratio the GTR. As well as lithospheric strength, however, a seamount’s GTR is affected by its height, shape and depth, which must be independently determined from bathymetry data. So, in order to estimate self-consistently volcano ages from GTRs following five steps are necessary:

(i) Find T for seamounts.
(ii) Measure G for as many of these as possible.
(iii) Use GTR values to measure lithospheric strength.
(iv) At a subset of sites where radiometric estimates of Δt exist, calibrate strength against Δt.
(v) Use the calibrated relationship between strength and Δt to date the seamounts.

These are detailed in order, then the method is discussed.

2.1 Isolating seamounts’ bathymetry

To avoid the size distortions present in regularised bathymetric grids (e.g. ETOPO-5) (McNutt 1998), the dimensions of seamounts are determined directly from 21.7 × 10^3 km of 3098 echo-sounder profiles shown in Fig. 1. Using scale-independent morphological criteria, seamounts up to 200 km across are isolated from larger-scale trends such as hotspot swells using MiMic, an algorithm already proven across the Pacific (Hillier & Watts 2004, 2005a). The resulting traverses across seamounts are approximated as flat-topped trapezia (e.g. Jordan et al. 1983; Smith 1988; Smith & Jordan 1988; Smith & Cann 1992; Bridges 1997; Hammond 1997; Rappaport et al. 1997; Wessel 2001) optimized using a simplex-like method (Appendix A). Only trapezia well fit in shape and cross-sectional area are retained (ΔArea < 10 per cent). Traverses that do not pass directly (i.e. edge1-summit-edge2 acute angle of <150°, or an along-track distance of >110 per cent of a direct path) across seamounts are also discounted at this point. Finally, where multiple traverses of a seamount exist, the tallest trapezium is presumed to best estimate the relief of a radially symmetric seamount. 121 882 heights (>100 m) of possible seamounts are determined, 3874 of the 121 882 height determinations thought to represent 2706 of the 121 882 height determinations thought to represent seamounts are located coincident with one of Wessel & Lyons’ (1997) 7968 gravity anomalies in the area 60°S–60°N and 130°E–70°W; namely, the centre of the gravity anomaly is within the bathymetric footprint of the seamount. These gravity anomalies were carefully isolated from larger-scale spatial trends in the gravity field.

Figure 2. Thin elastic plate model as applied to an oceanic seamount, adapted from (Watts 2001). A volcano (dark grey) loads a strong lithosphere and causing a distributed depression (light grey) including a flexural moat and a bulge. The buoyancy response is created when less dense material (crust) is pushed downwards into more dense material (mantle). The effective thickness of the mechanically strong lithosphere, $T_e$, is the thickness of a standard plate (with properties of each author’s choosing, see Appendix B) that has a rigidity causing it to best fit the observed deformation. $\rho$ is density, vertical arrows represent forces.

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such that they most likely represented seamounts. Any disagreement in location may result from either incomplete coverage bathymetric, incomplete gravity coverage (~10 km track spacing), or migration during the gravity anomaly location procedure of Wessel & Lyons (1997). Uncertainty also exists where seamounts are in close proximity and their flanks merge substantially. Here interpretation of the two data sets may differ and, when more than one gravity anomaly falls within a bathymetric footprint, the seamount is discarded and does not contribute to the total of 2706. These GTR observations are now ready to be interpreted in terms of a flexural model, and then related to $\Delta t$ values.

### 2.3 Estimating lithospheric strength

In theory, for each seamount, an observed GTR value ($GTR_{os}$) falls somewhere on a linear scale between a minimum and a maximum theoretical GTR value ($GTR_{min}$ & $GTR_{max}$) representing very strong and very weak lithosphere, respectively. The position of $GTR_{os}$ on the scale relative to these extreme strength values relates to the strength of the lithosphere supporting the seamount. This position may be dubbed $GTR_{rel}$, quantified (see eq. 1), and is a measure of lithospheric strength in the way that $T_e$ is:

$$GTR_{rel} = \frac{GTR_{os} - GTR_{min}}{GTR_{max} - GTR_{min}}.$$  

Practically, for each seamount, a plate with rigidity equivalent to a $T_e$ of 40 km is used to approximate $GTR_{max}$, and equivalent to a $T_e$ of 0.01 km used for $GTR_{min}$. An FFT solution of the flexural equation that allows for an arbitrarily shaped load and infill density to differ from load density ($p_{188}$ of Watts 2001) is computed within a 660 × 660 km area using physical constants in Table 1.

In an additional step measurement errors, estimated as ±250 m (1 standard deviation, $\sigma$) for topography (Carter 1980; Smith 1993) and ±5 mGal (1 $\sigma$) for gravity (Neuman et al. 1993; Marks 1996; Smith & Sandwell 1997), are accounted for by expanding the theoretical GTR range by an amount $\pm \epsilon$, where $\epsilon = [(500 \ m/T) + (10 \ mGal/G)] \times GTR_{os}$; an arbitrary but simple function. 500 m and 10 mGal represent 2 $\sigma$ errors in topography and gravity, respectively, and T and G are still observed topography and gravity over the centre of the seamount. This amendment is implemented by replacing $GTR_{min}$ and $GTR_{max}$ in eq. (1) with ($GTR_{min} - \epsilon$) and ($GTR_{max} + \epsilon$) (see eq. 2). Using the amended form of $GTR_{rel}$ stabilises it as a strength estimate, and is used from now on.

$$GTR_{rel} = \frac{GTR_{os} - GTR_{min} + \epsilon}{GTR_{max} - GTR_{min} + 2\epsilon}.$$  

It is also possible to estimate the most suitable $T_e$ for a seamount from GTR observations if, for example, one compares $GTR_{os}$ to theoretical values of GTR computed for a spectrum of possible $T_e$ values. Unlike for $GTR_{rel}$ where the effect of random measurement noise remains approximately symmetrically distributed, however, $T_e$ values are skewed to low values. $T_e$ values, therefore, correlate substantially less strongly with radiometric estimates of $\Delta t$, and best-fitting trend lines are biased. Using GTRs, therefore, $T_e$ is a less useful measure of strength than $GTR_{rel}$ when dating seamounts, and $GTR_{rel}$ is used from here on.

### 2.4 Calibrating strength against $\Delta t$

In order to date seamounts using geophysical estimates of lithospheric strength, it must be possible to convert strength into $\Delta t$. This can be done self-consistently within a study by taking a subset of strength data where $\Delta t$ is independently ‘known’ and calibrating a relationship between the two.

Fig. 3 plots strength of the lithosphere at the time of volcanic loading ($GTR_{rel}$) against lithospheric age at the time of loading ($\Delta t$) estimated from radiometric dating of seamounts (Koppers et al. 2003; Clouard & Bonneville 2005) and seafloor age the seafloor age of Müller & Roest (1997); $t_{volc} = t_{sf} - \Delta t$. Grey shading is an envelope around the data (black dots) at 61 locations (Fig. 1). Thin lines are trends fitted by the ordinary least squares method, and thick line (equation given) accounts for uncertainty in both variables (Marks & Sandwell 1991). Dashed line fits large seamounts, basal radius >20 km, shown as bigger dots.

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Table 1. Values of physical parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
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<tr>
<td>Gravity</td>
<td>9.81 m/s²</td>
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<tr>
<td>Young’s Modulus</td>
<td>$10^{11}$ N m</td>
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<tr>
<td>Poisson’s Ratio</td>
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<tr>
<td>Mantle density</td>
<td>3330 kg/m³</td>
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<tr>
<td>Crustal density</td>
<td>2800 kg/m³</td>
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<tr>
<td>Infill density</td>
<td>2650 kg/m³</td>
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<tr>
<td>Load density</td>
<td>2800 kg/m³</td>
</tr>
<tr>
<td>Water density</td>
<td>1030 kg/m³</td>
</tr>
<tr>
<td>Crustal thickness</td>
<td>7 km</td>
</tr>
</tbody>
</table>

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trends that consider errors in 1 variable (thin lines). Local studies to determine the ages of seamount chains also suggest that the correlation is real and correct (see Section 3.1).

2.5 Estimating seamount ages

The linear trend fitted to the $GTR_{rel} - \sqrt{\Delta t}$ data where radiometric ages exist (61 data) establishes a relationship of $GTR_{rel} = 0.09825 \sqrt{\Delta t} + 0.20251$, that is, a relationship between lithospheric strength and seamount age, presuming $t_{volc}$ is known: $t_{volc} = t_{rel} - \Delta t$. The trend, shown as the solid thick line on Fig. 3, assumes error in both variables (Marks & Sandwell 1991). It is this trend that is used to estimate an age for each of the 2706 Pacific seamounts shown in Fig. 4. Large seamounts (basal radius > 20 km) have an almost identical trend of $GTR_{rel} = 0.1036 \sqrt{\Delta t} + 0.1486$, shown as the thick dashed line on Fig. 3.

These calibrations of $GTR_{rel}$ against $\sqrt{\Delta t}$ use a subset of the 2706 geophysical strength estimates. As such the 61 strength estimates used in the calibration benefit from containing exactly the same systematic biases in the flexural model (e.g. assumed load density ($\rho_{load}$), Poisson's ratio ($\nu$), or Young's modulus ($E$), see Appendix B) as the rest of the 2706 geophysical strength estimates. Thus, systematic biases expected to have little effect on the 'flexural' age estimates. For instance, due to some systematic bias $GTR_{rel}$ values could be twice those produced using a model that truly matched reality. The calibration line would then have a gradient twice as steep, but also the 2706 $GTR_{rel}$ values would be twice what they should. A comparison of these $GTR_{rel}$ values to this steepened calibration line in order to deduce flexural ages would thus be unaffected by the systematic bias.

Only one trend is used to describe the calibration data set, on Fig. 3, as there is insufficient evidence to justify a distinct difference between the North Pacific and South Pacific or weak lithosphere in the Superswell region. Perhaps this study's resolution is insufficient. Alternatively, the 'abnormal' seamount chains in French Polynesia (Cook-Austral and Pitcairn) have multiple volcanic episodes.
(McNut 1998) and the GTR method considering only values over the summit may be less affected by this overprinting.

In Fig. 3, 1r about the mean $GTR_{\text{rel}}$ is, on average, 0.130. This translates into $\pm 1.32\sqrt{\Delta t}$ or $\sim 1$–11 Ma for a young (4 Ma) seamount and 114–178 Ma for an old (144 Ma) seamount. So uncertainties for individual seamounts are substantial (up to 10s of Ma) but map view plots (Fig. 4) demonstrate patterns for groups of seamounts to be reliable, a validation that is argued for in Section 3.1.

Before this, a justification of complexities in the method for estimating strength, an assessment of the causes of the scatter in the calibration plot, and further computational details are given in the following section.

### 2.6 Discussion of preferred method and further computational details

The method in Section 2.3 contains some complexities, which are most easily justified by considering methods without them.

The most simplistic way to estimate an observed GTR would be to extract G from the GTR model (Fig. 2) probably required to accurately model reality may also cause scatter, but are difficult to assess. For example, Watts & Ribe (1984) document the effect of variable infill density (allows blugses to be modelled accurately), laterally variable $T_e$, and a plate fractured under the load. Inelastic yielding, stress relaxation, a thick plate and underplating may also be required and are discussed in Watts (2001). It is also noted that the interpretation of effective plate thickness $T_e$ with respect to crustal thickness and indicators of material properties causing the strength (seismogenic thickness, mantle xenoliths, tomography) are currently debated (e.g. Maggi et al. 2000; Watts & Burov 2003; McKenzie et al. 2005). However, since we do not interpret further than requiring strength estimates to correlate with radiometric ages ($\sqrt{\Delta t}$), the simple model is perfectly valid one to date seamounts with.

### 3 RESULTS

After quality control 121 882 seamounts (3874 > 1.5 km tall) are extracted from echo-sounder data, of which 2706 (~1/3) match one of 7968 gravity anomalies (Wessel & Lyons 1997). For these, a measure of lithospheric strength has been calculated, and presuming the seamounts to be volcanoes they are associated with a Delta (i.e. tectonic setting) based on a self-consistent calibration using a subset of locations where radiometric dates exist.

Fig. 4 shows the distribution of on-ridge (▲) and off-ridge (●) Pacific seamount volcanism ($n_{\text{tot}} = 2706$) during four time intervals, which are plotted on current-day maps. From flexurally estimated values of $\Delta t$, seamount age, $t_{\text{volc}}$, is calculated from seafloor ages (Müller & Roest 1997), $t_{\text{fl}}$, using $t_{\text{volc}} = t_{\text{fl}} - \Delta t$. The division between eruption into ‘on-ridge’, $\Delta t < 20$ Ma, and ‘off-ridge’, $\Delta t > 45$ Ma, tectonic settings, that is, onto young and old lithosphere, is chosen to emphasise patterns documented from localised studies, but is not dissimilar to the speculative values of 2–8 Ma and >35 Ma used by Watts et al. (1980). Given doubt that may be expressed about the strength of correlation in Fig. 3, a validation of the geophysically estimated volcanic ages against other age estimates is important.

#### 3.1 Validation

A basic level validation is obtained from the general distribution of the estimated ages in Fig. 4. First, almost all data plot west of the spreading centre (i.e. on crust that has been created). This is expected of a ‘correct’ $GTR_{\text{rel}} - \sqrt{\Delta t}$ relationship, but in no way required by the conversion of strength estimates into ages. Secondly, basin-wide west-to-east, off-ridge to on-ridge progressions in the time intervals are consistent with older seafloor in the western Pacific. Finally, the geophysically predicted ages coincide spatially with radiometric estimates (+), which supports the predictions as the calibration does not require seamounts to plot in the right place at the right time. Note that spatial coincidence does not exist for arbitrary relationships, such as $\Delta t = 20$ Ma. Any en masse determination of age or tectonic setting should pass tests such as these.
More specifically, geophysical estimates of $t_{\text{vol}}$ are in accord with literature determinations of tectonic setting (either by $T_c$, radiometric dates, isotopes, magnetics, geochemistry, subsidence analysis). These agreements are summarised below; Hillier (2005) gives more detail, and Clouard & Bonneville (2005) collate radiometric ages.

(i) Oceanic plateaus (on-ridge): Shatsky Rise, 145–135 Ma (Den et al. 1969; Marks & Sandwell 1991; Sager & Han 1993; Nakaniishi et al. 1999; Sager et al. 1999; Verzhbitskii & Merkin 2000); Hess Rise, $\sim$100 Ma (Windom et al. 1981; Vailler et al. 1983; Marks & Sandwell 1991); Manihiki Plateau, 124–118 Ma (Hussong et al. 1979; Mahoney & Spencer 1991; Tarduno et al. 1991; Coffin & Eldholm 1993; Ito & Clift 1998), with late-stage reactivation at $\sim$70 Ma (Bercovici & Mahoney 1994; Ito & Clift 1998); Tumatuatu Plateau, $\sim$40 Ma (Talandier & Okal 1987; Ito et al. 1995; McNeaut 1998; Patriat et al. 2002).

(ii) Volcanic groups forming at divergent plate boundaries (on-ridge): Musicians, $\sim$95–75 Ma (Schwank & Lazarewicz 1982; Dixon et al. 1983; Freedman & Parsons 1986; Kopp et al. 2003) and the geophysical ages record a previously observed (Schwank & Lazarewicz 1982) north to south volcanic progression through time; Cobb hotspot trace, 20–0 Ma (e.g. Desonie & Duncan 1990); Foundation summits 20–0 Ma (Mammerickx 1992; Devey et al. 1997; O’Connor et al. 1998; Maia et al. 2001; Maia & Hamed 2002); Guadalupe chain (e.g. Koppers et al. 2001). Puka-Puka Ridge was also formed on young lithosphere (Sandwell et al. 1995).

(iii) Intraplate volcanic groups (off-ridge): Hawaiian chain, 0–65 Ma (Wilson 1963a,b; Moore & Clague 1992); Louisville Ridge, 0–80 Ma (Cazeneve & Dominh 1984; Cheng et al. 1987; Hawkins et al. 1987; Watts et al. 1988; Lyons et al. 2000); Marquesas, 0–40 Ma (Filmer et al. 1993; Caress et al. 1995; McNeaut 1998); Society 0–40 Ma (Filmer et al. 1993; Clouard & Bonneille 2001); Marshall-Gilbert summits, few radiometric dates $\sim$80 Ma (Davis et al. 1989), 60–70 Ma (Koppers & Staudigel 2005), $\sim$50 Ma (Kulp 1963) also bio-stratigraphic dates (Lincon et al. 1993).


This good agreement between flexurally estimated volcanic ages and dating from other sources gives strong support to the predictions. It is probably worth noting that plotting flexural ages in four time periods is a tougher test of these predictions than using a single plot. The Hawaiian-Emperor chain, 0–65 Ma (e.g. Wilson 1963a,b; Moore & Clague 1992), for example, are off-ridge along their length (except the northern tip) as the results of Watts et al. (1980), and show as this if Figs 4(a)–(d) are condensed onto a single plot (Data supplied in supplementary material). However, some Hawaiian volcanoes plot in Fig. 4(c) and some Emperor Chain summits in Fig. 4(d). It is uncertain what causes this but the Hawaiian Swell, flexural bulge from trenches to the north, or the Emperor chain passing between the Hess and Shatsky rises may be involved. As a counter-case, volcanism along the Louisville chain progresses as expected.

3.2 New observations and implications

Radiometric dates are few and dominantly from recent hotspot volcanism, so a key advantage of “flexural ages” is that they are more likely to constitute a representative distribution of seamount volcanism. It is, therefore, possible to better comment upon the distribution of Pacific volcanism across space and time.

First-order patterns presented in Fig. 4 are a strip of on-ridge volcanism from Easter Island through the Tuamoto Plateau to the Line Islands, and off-ridge volcanism most dominant in the NW Pacific.

On-ridge volcanism tends to occur in spatially diffuse groups such as the Wake seamounts and Mid-Pacific Mountains in Fig. 4(a) and south of Guadalupe on Fig. 4(d), but also shows many elongated trends (e.g. Wentworth summits, Foundation summits) although these are not as strictly linear as the off-ridge trends. Off-ridge volcanism is dominantly in lineations (e.g. Hawaii-Emperor chain, Louisville Ridge), but also occurs in a diffuse grouping upon the oldest seafloor in the NW Pacific. This demonstrates that a division between diffuse on-ridge volcanism and lined off-ridge volcanism as suggested by Watts et al. (1980) is not necessarily adhered to.

From Fig. 4 it can be observed that multiple ages of volcanism commonly appear to occupy the same area of seafloor, for example, around the Line Islands. Alternatively phrased, geologically defined regions may mix of both on-ridge and off-ridge tectonic settings. This implies prolonged periods of activity, or quiescence then reactivation. Minimal off-ridge activity in places such as Hess Rise and the Musicians imply that the mixture is not just an artefact due to scatter caused by the method. Where two or three generations of volcanism occur in close proximity, $T_c$ estimation is difficult (McNeaut 1998), but just using GTR values over seamounts’ summits could act to reduce the influence of neighbouring edifices.

A region where prolonged activity (~160–75 Ma) seems to have occurred is in the old NW Pacific around the Wake summits. This area is most clearly shown by the cluster of off-ridge activity in Fig. 4(b). Similar is true, though less obviously, for the Mid-Pacific Mountains. Interestingly, volcanism initiated near a spreading centre, but continued in the same region well after the ridge is distant, implies a lithospheric (as opposed to asthenospheric) component to the mechanism that controls the location of volcanism.

French Polynesia also exhibits enduring volcanic activity, but along lineaments. On-ridge activity starts at the NW end of the Cook-Austral alignment in (b), moves to near the present day Pitcairn hot spot in (c), and continues ESE along the Foundation chain in (d). Later, off-ridge volcanism starts in (c), again at the NW end of the alignment, and occurs along the alignment in (d) presumably related to the current day hot spot. A somewhat similar relationship (later hot spot on pre-existing volcanic chain) appears to exist between the Easter Island chain and Pitcairn hot spot. One explanation for these patterns could be that the first on-ridge episode facilitated or influenced the later off-ridge one. Perhaps fresh volcanism occurs as on-ridge chains move west on the Pacific plate and encounter anomalous asthenosphere thought to exist there (see SOPITA, superplume (Larson 1991), and Darwin Rise (Menard 1964)). A possible mechanism for the facilitation, proposed at Mid-Ocean Ridges (Niu & Batiza 1993; Aharanov et al. 1995; Kelemen et al. 1997), is that melt–matrix interaction creates low-permeability dunite conduits. Alternatively, the stress field imposed by existing edifices, through cracking, could govern where volcanoes erupt (e.g. Ten Brink 1991).

More specific original observations from Fig. 4 include the following:
(i) In (a) and (b), the Cross Lines are poorly dated, but flexural ages indicate that they formed in the Cretaceous near, and at a high angle to, a spreading ridge. The Easter Island and Foundation chains may be modern analogues of this near-ridge formation. Flexural ages also indicate less prevalent later off-ridge activity in the Cross Lines.

(ii) In (b), flexural ages supplement sparse observations of the Wentworth seamounts (Watts et al. 1980; Pringle & Dalrymple 1993) and provide the first age indication for the ‘Liliuokalani Ridge’ (Pringle & Dalrymple 1993) immediately to the east. Both appear on-ridge, so are probably related to the formation of the Hess Rise.

(iii) In (b), the Musicians and the Southern Hawaiian seamounts (a.k.a Geologist Seamounts) (Sager & Pringle 1987) start to form contemporaneously near the ridge crest from ~100 Ma, the gap between the two probably explained by ridge geometry still visible in the 80 Ma isochron. Volcanism along this ‘soft’ Musicians-Geologists alignment, then continues as the ridge moves away. Tentatively, note that the recent trace of the Hawaiian hotspot (last several Ma) appears to divert to the south along this alignment, perhaps explained if lithosphere previously subject to volcanism is more easily permeable to subsequent melt. In caution, note that seafloor here has few magnetic reversals (Cretaceous quiet zone), which contributes to uncertainties (e.g. Kopp et al. 2003).

(iv) In (c) volcanism 200–300 km SW of the current Society hot spot is on-ridge. The volcanism is called the Tavara seamounts (formerly the Savannah seamounts (Bonneville et al. 1997; Sichoix et al. 1998; Hillier & Watts 2004)), and the Va’a Tau Piti Ridges. Clouard & Bonneville (2003) find ‘on-ridge’ $T_e$ values of 8 and 3 km, respectively. If 2 K/Ar dredge haul ages of 43 and 36 Ma are trusted (see Baksi 2005; Clouard & Bonneville 2005), however, they indicate an off-ridge tectonic setting. A scenario where early large-volume volcanism is covered with a later veneer of off-ridge volcanism would reconcile these observations. The volcanism SW of Society shares several features with the Musicians; on-ridge tectonic setting, proximity to a transform fault, predicted morphology (Smith & Sandwell 1997), and very low $T_e$ ridges approximately parallel to fracture zones (i.e. were approximately perpendicular to the spreading ridge). So, mechanisms of origins may be similar.

(v) In (d), on-ridge activity clustered south of Guadalupe coincides with a broad region of shallow seafloor and slow topographic velocities in the depth range 0–125 km (Hillier & Watts 2004). If the observations have a common origin, they may indicate enhanced mantle melting leading to regionally thick crust, or alternatively imply a thermally thin lithosphere. Whatever mechanism is selected, however, this is a good indication that the oceanic lithosphere does not form homogeneously along a ridge or through time. Thus, any planform of large-scale bathymetric anomalies may substantially relate to processes creating the plate at the ridge and not to processes such as mantle convection.

4 DISCUSSION

It has become something of an orthodoxy to associate weak lithosphere, measured by low $T_e$ values, with an ‘on-ridge’ tectonic setting; namely volcanic loads applied to young seafloor (low $\Delta t$) in the proximity of an oceanic spreading centre. The primary evidence for this relationship between $T_e$ and $\Delta t$ comes from compilations such as that of Watts (2001) where $\Delta t$ is independently estimated, generally using radiometric dates. High $T_e$ is similarly linked to ‘off-ridge’. So, in flexural studies the terms ‘on-ridge’ and ‘off-ridge’ implicitly assign formation dates to oceanic features, albeit approximated ones. The steps made in creating this link are; determining strength, calibrating strengths against independently obtained age data, and then associating strengths with ages.

This work automates the above steps and self-consistently combines them into a single paper. This harmonises the computational method and physical properties used, but automation probably introduces some scatter. However, a link between lithospheric strength and $\Delta t$, and thus volcano age, has been established. Here, instead of just quoting tectonic setting, strengths are converted into ages. With estimated uncertainties in the order of 10 Ma (see Section 2.5) this may be pushing the data, but conversion into ages is favoured as it offers a direct, numerical, and perhaps stark, comparison to chronological constraints from other sources. The flexural ages estimated here appear to withstand such a comparison in the validation, and the data is given as a supplementary table for future comparisons.

Converting to ages also permits an examination of the rate of volcanic activity through time and, unlike previous studies, a division into tectonic settings is possible giving insights into magmatic processes.

4.1 Rate through time

Fig. 5(a) is a histogram of how the volume of seamount volcanism distributes itself across oceanic seafloor of different ages. There are two main peaks of activity centred upon seafloor of 160–150 and 120–110 Ma, respectively spanning 170–140 Ma and 130–100 Ma (convex curves). Most obviously in these peaks, on-ridge through to off-ridge volcanism occurs on the same age of seafloor (bold vertical arrows). This indicates volcanism over a long duration through the same lithosphere. There is also an increase in volume from 60 Ma to 0 Ma seafloor.

Fig. 5(b) is a histogram of the volume of seamount volcanism through time. The two peaks of activity above are detectable (convex curves and arrows), but overall pattern now has a dominant peak between 80 and 120 Ma. Volcanism before this is quite high, afterwards dips around 60 Ma, and has a broad subsidiary peak centred around 30 Ma. Volcanism is dominantly (>1/2) on-ridge. Off-ridge volcanism is a quasi-constant between 30 and 120 Ma, but most voluminous between 30 and 50 Ma, is reduced below 30 Ma, and by definition not present before 130 Ma.

The most comparable previous curves to Figs 5(a) and 5(b) are those of Smith (1990) and of Wessel & Lyons (1997), shown on Fig. 5(c). Smith’s curve of volcanism upon seafloor of a given age and Fig. 5(a) both have the peak at ~120 Ma, but Smith does not find the one at 160–150 Ma, or the increase from 60 Ma. The increase is much reduced when the normalized for seafloor area (km$^2$/km$^2$), so is partly due to the increasing area of young seafloor, but relatively easy altimetric detection of seamounts on young shallow seafloor is probably also a factor.

If total volcanism through time is considered Fig. 5(b) looks rather similar to that deduced for his ‘pseudo-ages’ by Wessel & Lyons (1997), despite the cautions about this data stated in the introduction. The broad eruptive peak spans 120–80 Ma. It starts with the Ontong-Java Plateau (Eldholm & Coffin 2000) and coincides with a period of high crustal production at mid-ocean ridges. Large Igneous Province (LIP) volcanism (Fig. 4d) (Eldholm & Coffin 2000) and seamount volcanism do not seem to affect long-term strontium isotopes, oxygen isotopes or sea-level (Figs 5e–g), which it is reasonable to assume are controlled by mid-ocean ridge spreading (Jones & Jenkins 2001). It appears however that seafloor spreading, perhaps assisted by seamount volcanism, sets the conditions between 120
and 80 Ma for the environmentally stressful conditions to create ‘black shales’ (Jenkins 1980), and allow LIP volcanism at \( \sim 120 \) and \( \sim 90 \) Ma to cause ocean-wide anoxic events as well as short term variations in the strontium isotope curve (Jones & Jenkins 2001; Jenkins 2003). The volcanism at 120 Ma also appears to have caused extinction and a marine biological crisis (Coffin & Eldholm 1994; Erba 1994; Sinton & Duncan 1997; Kerr 1998; Larson & Erba 1999).

Whilst seamount volcanism is much less voluminous than LIP volcanism and crustal production at the spreading ridges, the geometry of seamounts may amplify their effects. Being close to the ocean’s surface, rising plumes can more easily impact upon shallow waters that are an important in some models (e.g. Larson & Erba 1999). Similarly, recent numerical modelling of chemical transport also indicates that seamount formation may impact disproportionately and ‘contribute to globally significant hydrothermal fluxes’ (Harris et al. 2004).

Instead of the total volcanic volumes, it is the division into tectonic settings that provides evidence regarding the mechanism behind seamount volcanism. Both Figs 5(a) and 5(b) show 2 peaks of activity (convex curves and arrows) indicate volcanism over a long duration through the same lithosphere, in these cases around the Wake Seamounts and Mid-Pacific Mountains (1st peak) and the Line Islands (2nd peak). This reiterates the observations of persisting volcanism through lithosphere as it moves a considerable distance over the asthenosphere in Fig. 4, which again imply a strong lithospheric component to the mechanism controlling the location of volcanism. As for the chains in French Polynesia, a possible mechanism for this proposed at Mid-Ocean Ridges (Niu & Batiza 1993; Aharon et al. 1995; Kelemen et al. 1997) is that melt–matrix interaction creates low-permeability dunite conduits. Alternatively, the stress field imposed by existing edifices, through cracking, could govern where volcanoes erupt (e.g. Ten Brink 1991).

5 CONCLUSIONS

It is demonstrated that automated, self-consistent, basin-wide estimation of lithospheric strength (GTR\(_{rel}\)) at the time of volcanic loading (\( \Delta t\)) is possible in the Pacific using the ratio of gravity to topography above the summits of volcanoes. These strength estimates are shown to correlate with directly sampled, radiometric, ages. At the accuracy of this study, only one strength–\( \Delta t\) relationship is required for the Pacific; \( GTR_{rel} = 0.09825 \sqrt{\Delta t} + 0.20251 \). Using this, it is for the first time possible to assign eruption ages en masse, specifically to 2706 volcanoes. Validation proves these ‘flexural ages’ are at least accurate enough to determine the tectonic setting of volcanic emplacement (i.e. on-ridge and off-ridge), make maps of seamount volcanism through time, and construct histograms of activity.

The flexural ages show that: A division between diffuse on-ridge volcanism and lineated off-ridge volcanism as suggested by Watts et al. (1980) is not necessarily adhered to; Variation in the spatial density of on-ridge volcanos on 0–80 Ma seafloor supports the idea that oceanic lithosphere does not form homogeneously along a ridge or through time; The Cross Lines are on-ridge and \( \sim 120–110 \) Ma; The Wentworth Seamounts and Liliuokalani Ridge are probably connected to the formation of the Hess Rise.

Perhaps the most significant observation, however, is of widespread recurrences of volcanism proximal to older features, namely a mix of tectonic setting within geological regions. Prominently, this occurs in regions such as the oldest seafloor in the NW...
Pacific, and within volcanic chains such as Cook-Austral and Pitcairn in French Polynesia. The former is seemingly persisting volcanism through the same lithosphere even as it moves substantially over the asthenosphere, and the later reactivation of chains as they encounter more fertile asthenosphere. What both appear to signify, however, is that the lithospheric exerts a significant element of control upon the location of volcanism, and that magmatic throughput leaves the lithosphere more susceptible to the passage of future melts.

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APPENDIX A: APPROXIMATING SEAMOUNTS AS TRAPEZIA

Fig. A1 illustrates a robust, scale-independent, automated method to approximate seamounts as trapezia using one of the bathymetric highs on cruise v3312 as an example.

To ensure scale invariance, the data are plotted in normalised distance, $x_{\text{frac}}$, and height, $z_{\text{frac}}$, which range from 0 to 1. $x_{\text{frac}} = \frac{x - x_{\text{start}}}{(x_{\text{finish}} - x_{\text{start}})}$, and $z_{\text{frac}} = \frac{z - z_{\text{min}}}{(z_{\text{max}} - z_{\text{min}})}$, where $z_{\text{min}}$ and $z_{\text{max}}$ are the shallowest and deepest extremities of the seamount, whilst $x_{\text{start}}$ and $x_{\text{finish}}$ are the left and right hand limits, respectively.

The trapezium is described by the $x$ and $z$ coordinates of its 4 corners ($x_{1...4}$, $z_{1...4}$), numbered on Fig. A1(a). To efficiently invert for these 8 parameters a method is adopted where each of the corners of the moves sequentially (i.e. 1, 2, 3, 4, 1, 2 ... etc) toward a trapezium that better fits the data. In a further simplex-like (Dantzig 1963) optimisation, each movement selects the best location from 8 directions, taking a step that is double, half, or the same size as last time (Fig. A1b) such that the movement possibilities for the next iteration upon that corner are increased in size, decreased, or remain the same, respectively. To simplify and stabilise the inversion the top is kept flat and the corners are required to retain their relative position (e.g. corner 2, upper-left).

Data is resampled at equal intervals along the line between points so the visual shape is fitted unaffected by seafloor slope and data density.

Total misfit, $\Delta_{\text{tot}}$, is an equally-weighted sum of misfits, $\Delta_{\text{side}}$, for the 4 sides of the trapezium. Each $\Delta_{\text{side}} = \sum_{i=1}^{n} |z_{\text{measured}} - z_{\text{line}}|$, where $n$ is the number of data points ($n$ kept $\geq 1$). Misfits for the left side are horizontal, calculated for points with $z_{\text{frac}} < z_{2}$ to the left of the summit. The right side mirrors this using $z_{\text{frac}} < z_{1}$. Misfit for the top is vertical and calculated for the remainder of the points. Basal misfit is vertical and between the bottom and regional depth.

Initial locations of the corners used are (0, 0) (0, 0.75) (1, 0.75) (1, 0), and initial step size is 0.1. Iteration is stopped after 20 cycles, when all points are not moving (i.e. step to use is $< 0.001$), or $\Delta_{\text{tot}} < 0.01$. Commonly, 8–12 iterations are required, and the 340 seamounts of v3312 (Fig. A2) are approximated in 45 seconds by an ~1 GHz processor, about 0.13 s per fit.

APPENDIX B: VALUE OF PHYSICAL PROPERTIES

From drilling (Hyndman et al. 1979) seamount density, $\rho_{\text{load}}$, is commonly stated as 2800 kg m$^{-3}$ (e.g. Freedman & Parsons 1986; Craig & Sandwell 1988; Schubert & Sandwell 1989; Wessel & Lyons 1997), comparable to average oceanic basement (2860 ± 30 kg m$^{-3}$) (Carlson & Herrick 1990). However, 2950 kg m$^{-3} > \rho_{\text{load}} \geq 2100$ kg m$^{-3}$, probably dependent upon the amount of extrusives present and the extent of mass wasting (Carlson & Herrick 1990; Hammer et al. 1994; Minshull & Charvis 2001; Turcotte & Schubert 2002). Poisson’s ratio ($\nu$) is 0.25–0.5 (e.g. Calmant et al. 1990; Ribe & Watts 1982; Freedman & Parsons 1986; Craig & Sandwell 1988), and Young’s modulus ($E$) is $6.5 \times 10^{10}$ N m$^{-3}$ (e.g. Sandwell 1984; Craig & Sandwell 1988) to $10^{12}$ N m$^{-3}$ (Calmant et al. 1990).
Figure A2. Isolation (by MiMIC) and approximation (as trapezia) of seamounts for bathymetry data collected on cruise v3312. Expanded sections are marked by dashed boxes and labelled with letters a–d. In these, well-fitted seamounts are dark-grey, poorly fitted highs light-grey. Feint line is bathymetry, bold line is the ‘regional’ depth (i.e. that which passes underneath seamounts) estimated by MiMIC, and for comparison intermediate thickness solid, dotted and dashed lines are regional depths estimated by spatial median, mean and mode, respectively (400 km wide, filtered of GMT Wessel & Smith (1998)). MiMIC is applied exactly as first pass in Hillier & Watts (2004, 2005a). Star indicates seamount in Fig A1.