



Dominant influence of volcanic loading on vertical motions of the Hawaiian Islands



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ABSTRACT

Rates of island vertical motion above intra-plate hotspots record the processes that arise from interactions between lithospheric plates and mantle plumes. To assess the contribution of mantle and lithospheric processes to surface motion above the Hawaiian hotspot, we compare simple models of lithospheric deformation to vertical motion rates measured from dated paleoshorelines and tide gauge records in the Hawaiian Islands. Our analysis shows that observed uplift and subsidence rates mainly record the flexural response of the lithosphere to volcanic loads. The effective elastic plate thickness that best fits the spatial distribution of subsidence and uplift rates is ~ 40 km, consistent with previous estimates based on total vertical deflection.

Because volcanic loading dominates the vertical motion signal, Hawaiian Islands appear to follow a predictable trajectory of vertical motion when they reside within one flexural half-wavelength of the active volcanic center. Islands initially subside at rapid and decreasing rates in the first ~ 1 Myr following their construction, uplift relatively slowly ~ 1 – 2.5 Myr following their construction, and eventually subside again, but at slow rates, within ~ 5 Myr of their construction. This observed pattern of uplift and subsidence is consistent with the pattern of vertical motion predicted to result from volcanic loading at the Hawaiian hotspot.

Lithospheric migration over the long-wavelength topographic swell associated with the Hawaiian hotspot has a comparatively minor influence on island uplift and subsidence. Its contribution to island vertical motion is not readily observed in our data, with the possible exception of some uplift observed in the past ~ 500 kyr on O'ahu that might correspond to non-steady state behavior of the Hawaiian plume.

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1. Introduction

Bathymetric swells localized over mantle plumes provide a source of uplift and subsidence for intra-plate volcanic ocean islands. Islands may uplift or subside with the seafloor as swell topography evolves (Ramalho et al., 2010a, 2010b) or as plate motion carries them over existing swells (Zhong and Watts, 2002). Rates of island vertical motion may therefore aid in constraining upper mantle dynamics in oceanic hotspot settings. However, rates may also be influenced by conductive cooling of the oceanic lithosphere (Detrick and Crough, 1978), intrusive magmatism (Ramalho et al., 2010a; Madeira et al., 2010), and the flexural response of the lithosphere to volcanic loads (Moore, 1970).

The relative contributions of these processes to island vertical motion are uncertain. Although volcanic loading of the lithosphere is suspected to play a significant role in driving island subsidence in the Hawaiian Islands (Moore, 1970), the primary cause of uplift is less certain (Zhong and Watts, 2002). Studies of uplift and subsidence in the Hawaiian Islands are typically based on total vertical deflection of the lithosphere (e.g. Watts and ten Brink, 1989). Measurements of intervening uplift and subsidence rates, although subject to uncertainty, can provide an independent observation of lithospheric deformation. Vertical motion rates averaged over different timescales can also help to distinguish between asynchronous deformation processes. Nevertheless, few studies have compared observed rates of vertical motion to theoretical models (Wessel and Keating, 1994; Watts and Zhong, 2000; Zhong and Watts, 2002; Ramalho et al., 2010a, 2010b).

Because of their sustained, dominantly extrusive magmatism (Robinson and Eakins, 2006; Flinders et al., 2013), steady rate of underlying plate motion (Morgan and Morgan, 2007), and abun-

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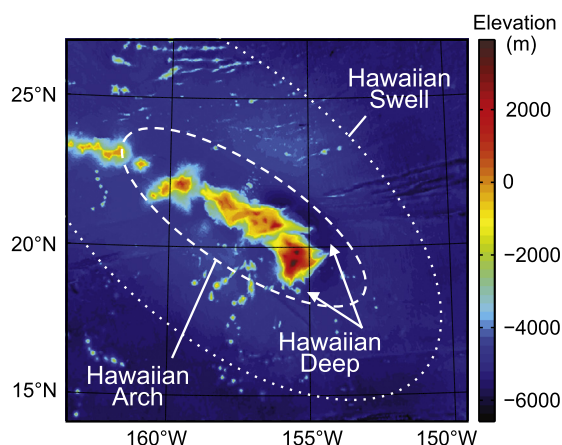


Fig. 1. Bathymetry of Hawai'i (Smith and Sandwell, 1997) with dotted line outlining the Hawaiian Swell, dashed line following the Hawaiian Arch, and arrows pointing to the Hawaiian Deep.

dance of dated paleoshoreline markers (Jones, 1993b; McMurtry et al., 2010), the Hawaiian Islands provide a unique opportunity to study coupled lithosphere and mantle dynamics. Here we examine the contributions of volcanic loading and mantle buoyancy to observed rates of Hawaiian island vertical motion. Specifically, we compare uplift and subsidence rates from tide gauge records and dated paleoshorelines to rates of vertical motion predicted for two models of the Hawaiian Swell and a range of effective elastic plate thicknesses. These comparisons provide constraints on the flexural rigidity of the lithosphere, the transient evolution of swell topography, and the contribution of mantle and lithosphere processes to uplift and subsidence in the Hawaiian Islands.

2. The Hawaiian Islands

The Hawaiian Islands lie at the southeast end of the Hawaiian-Emperor Seamount Chain. Formed by the passage of the Pacific Plate over the Hawaiian plume, a region of hot upwelling mantle, the chain of shield volcanoes stretches over 6000 km across the North Pacific Ocean and records ~80 Myr of hotspot volcanism (Clague and Dalrymple, 1987). The eight main Hawaiian Islands, formed in the past ~6 Myr, and their surrounding islets are the only subaerial exposures of the volcanoes.

As the seafloor passes over the Hawaiian hotspot, the zone of focused melting associated with the Hawaiian plume, it is uplifted and subjected to the massive additions of magma and volcanism forming the Hawaiian Islands. The "Hawaiian Swell" (Fig. 1) extends ~1200 km across the volcanic ridge (Detrick and Crough, 1978). Regionally, the seafloor is more than a kilometer shallower than typical 80 Myr to 100 Myr-old seafloor. Heat, pressure from viscous flow in the asthenosphere, intrusive magmatism, or some combination of these forces is thought to support this bathymetric high (Wilson, 1963; Detrick and Crough, 1978; Li et al., 2004). Within the larger swell, the islands sit in a ~280-km wide depression flanked by a broad ring of shallow bathymetry. The "Hawaiian Deep" and "Hawaiian Arch" refer to the flexural moat and peripheral bulge of the lithosphere as it is flexurally deformed by the load of the volcanoes (Dietz and Menard, 1953).

The individual Hawaiian volcanoes are typically constructed during a ~500 kyr period of shield-stage volcanism (Moore and Clague, 1992) that is initially volumetrically large ($\geq 0.1 \text{ km}^3 \text{ yr}^{-1}$) but that declines progressively with time (Lipman, 1995). Shield building typically follows a ~100 kyr stage of early alkalic volcanism. The volcanoes generally emerge above sea level 300 ± 100 kyr

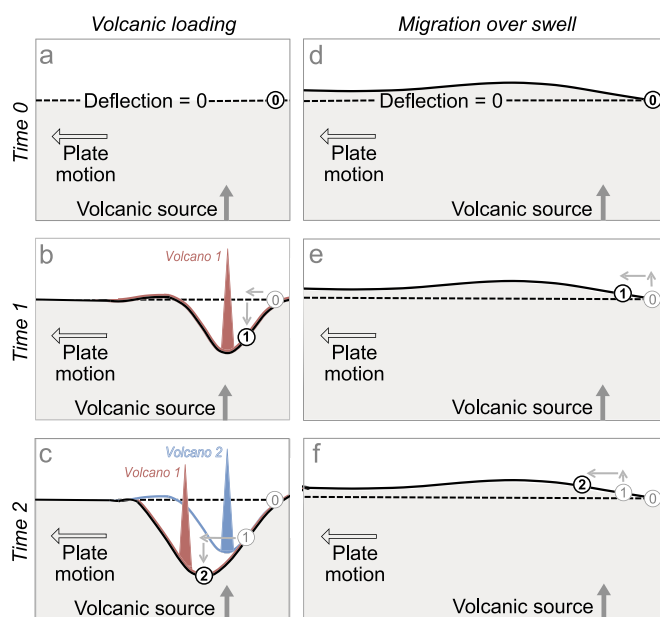


Fig. 2. Island vertical motion due to volcanic loading and plate migration over swell topography. The initially undeformed lithosphere (a) flexes under the weight of a new volcano, Volcano 1, resulting in subsidence beneath and adjacent to the volcano and uplift at the flexural outer bulge (b). Plate motion carries Volcano 1 and the deflected lithosphere away from the volcanic source, and the pattern of flexural subsidence and uplift is repeated with construction of a new volcano, Volcano 2 (c). Plate motion simultaneously carries surface features over swell topography, which remains fixed in relation to the volcanic center. A surface feature consequently uplifts and subsides as it travels over gentle swell slopes (d, e, f).

into this ~600 kyr period and attain their maximum subaerial exposure near the end of the shield building phase (Moore and Clague, 1992). The growing volcanoes apply a load to the underlying lithosphere and cause flexural subsidence (Fig. 2b). At the end of shield building, regional subsidence related to flexural loading supersedes local volcanic growth and Hawaiian volcanoes begin to submerge.

Northwesterly movement of the Pacific Plate translates both the volcanoes and the deflected lithosphere away from the hotspot at a velocity of $92 \pm 3 \text{ mm yr}^{-1}$ (Morgan and Morgan, 2007). Construction of new volcanoes "upstream" also flexes the lithosphere, and older "downstream" volcanoes subside if they are within the flexural moat of the new volcano (Fig. 2c) or uplift if they reside on its flexural outer bulge. This pattern is repeated with the construction of each new volcano, so a Hawaiian volcano is ultimately deflected by a series of younger volcanoes in the chain. Plate motion also translates the volcanoes over the swell topography, and a volcano may uplift and subside slowly as it migrates over gentle swell slopes (Fig. 2d, e, f).

Thus the rates of hotspot island vertical motion are set by mantle processes and by the response of the lithosphere to volcano emplacement. In this paper, we compare rates of vertical motion measured from tide gauge records and dated paleoshorelines in the Hawaiian Islands to simple models of volcanic loading and migration of the lithosphere over swell topography.

3. Observed rates of island vertical motion

3.1. Vertical motion rates from dated paleoshorelines

We compiled 459 published age determinations for 384 samples collected from three types of paleoshoreline indicators in the Hawaiian Islands: (1) transitions between submarine and subaerial

ally erupted basalts, (2) drowned coral deposits, and (3) uplifted coral deposits (Appendix A).

3.1.1. Paleoshoreline indicators and dating techniques

Intercalated ash and marine sediments and changes in lava morphology, sulfur content, vesiculation, and vitrification mark transitions between submarine and subaerially erupted basalts (Moore and Thomas, 1988; Moore and Clague, 1992; Sharp and Renne, 2005). The present elevations of these transitions indicate the change in relative sea level since the basalt crystallized. Potassium–argon and $^{40}\text{Ar}/^{39}\text{Ar}$ ratios can be used to determine the time since crystallization.

Slope breaks corresponding to transitions between slowly cooled subaerial lava and lava rapidly chilled and fragmented by seawater are also evident in the bathymetry of each Hawaiian volcano (Moore and Clague, 1992). These features (e.g., transitions from slopes of $\sim 4^\circ$ to slopes of $\sim 13^\circ$ 500 m below present sea level on Kilauea and Mauna Loa) indicate the relative sea level at the end of each volcano's shield building phase.

Coral also preserves a record of relative sea level. Coral can develop extensive reefs during periods of relatively stable sea level. However, when sea level rises and vertical reef accretion cannot keep pace, coral reefs drown (Toomey et al., 2013). When sea level falls, coral reefs can be subaerially exposed or subjected to water depths too shallow for their survival. Reservoirs of radioactive isotopes that living corals take up from their environment, including ^{14}C , ^{234}U and ^{87}Sr , decay without replenishment after coral algae cease to photosynthesize or coral growth terminates, thus providing a record of the time since corals drowned or died from exposure (Libby and Johnson, 1955; Ku, 1976; McArthur, 2010). Electron spin resonance (ESR) frequencies can also be used to determine the length of time that coral deposits have been irradiated by doses of natural radioactivity in their immediate environment (Grün, 1989).

3.1.2. Vertical motion rate calculations

For each of the 252 paleoshoreline ages $\tau_{\text{obs}} \geq 100$ kyr (214 samples), we calculated an average rate of island vertical motion, $\dot{z}_{\text{obs}}$:

$$\dot{z}_{\text{obs}} = \frac{z_{\text{sample}} + z_{\text{crit}} - z_{\text{psl}}}{\tau_{\text{obs}}} \quad (1)$$

where z_{sample} is the sample elevation, z_{crit} is the critical water depth for coral survival, and z_{psl} is the elevation of sea level relative to present sea level at time τ_{obs} (Fig. 3). For drowned coral samples dredged or collected as loose talus, we set z_{sample} equal to the mean elevation of the nearest upslope terrace mapped in the bathymetry (Ludwig et al., 1991) because their sampling elevation does not necessarily correspond to a past sea level.

The majority of reef-building coral in the Pacific Ocean, including most species of *Pocillopora* and *Porites*, grow in water depths ranging from <1 m to ~ 27 m, though some grow at depths >60 m (Wells, 1954). We assumed a critical water depth for coral survival z_{crit} of 0.5 ± 0.25 m for emerged corals, 30 ± 15 m for drowned corals, and 0 m for noncoralline samples (e.g. Muhs and Szabo, 1994).

For sample ages $\tau_{\text{obs}} \leq 361$ kyr ($n = 161$), we linearly interpolated z_{psl} , the elevation of sea level at time τ_{obs} before present, from the linear transformation of a $\delta^{18}\text{O}$ record of planktonic foraminifera adjusted for temperature and local hydrologic changes based on Mg/Ca ratios (Lea et al., 2002). For samples predating this paleo sea level reconstruction ($\tau_{\text{obs}} > 361$ kyr, $n = 92$), we interpolated z_{psl} from direct linear transformation of spliced $\delta^{18}\text{O}$ planktonic foraminifera records (Miller et al., 2005).

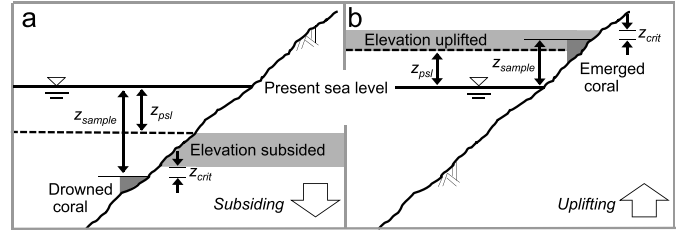


Fig. 3. Schematic illustration of the method for inferring (a) island subsidence from the elevation of drowned coral on a subsiding volcanic edifice and (b) island uplift from the elevation of emerged coral on an uplifting volcanic edifice. Solid horizontal line marks present-day sea level and dashed horizontal line marks sea level z_{psl} at the time of coral death. Elevations z_{sample} and z_{psl} are defined to be negative below present sea level and height z_{crit} is positive.

We calculated vertical motion rates corresponding to the bathymetric slope breaks on the Big Island of Hawai'i and the slope break on Haleākala using ages for these features based on their stratigraphic relation to radiometrically dated coral terraces (Moore and Clague, 1992). For slope breaks on the older islands, we used K-Ar ages of shield stage flows (Clague and Dalrymple, 1987).

3.1.3. Vertical motion rate uncertainty calculations

We assigned each vertical motion rate $\dot{z}_{\text{obs}}$ an approximate 1σ uncertainty $\delta\dot{z}_{\text{obs}}$ equal to the sum of the fractional uncertainties on its paleoshoreline elevation and paleoshoreline age:

$$\delta\dot{z}_{\text{obs}} = |\dot{z}_{\text{obs}}| \cdot \left(\frac{\delta z_{\text{crit}} + \delta z_{\text{psl}}}{|z_{\text{sample}} + z_{\text{crit}} - z_{\text{psl}}|} + \frac{\delta\tau_{\text{obs}}}{|\tau_{\text{obs}}|} \right) \quad (2)$$

where δz_{crit} is the assumed uncertainty on z_{crit} , δz_{psl} is the standard deviation of paleo sea level elevations interpolated at 100 evenly spaced increments within the 2σ error bounds of the paleoshoreline age τ_{obs} , and $\delta\tau_{\text{obs}}$ is the 1σ analytic and measurement uncertainty on the sample age. We used the sum of the fractional uncertainties rather than their quadratic sum because the uncertainties on the paleoshoreline sea level elevation δz_{psl} and paleoshoreline age $\delta\tau_{\text{obs}}$ are correlated (age uncertainty widens the range of possible paleo-sea levels, which makes paleoshoreline elevation less certain). This provides an upper-bound estimate of $\delta\dot{z}_{\text{obs}}$. The formulation of δz_{psl} assumes that uncertainty in the sample age τ_{obs} is the dominant source of uncertainty in the inferred paleo sea level z_{psl} relative to the uncertainty associated with the sea level curve itself. We neglected uncertainty on the sample elevation z_{sample} .

3.2. Vertical motion rates from tide gauge records

We calculated vertical motion rates from sea level measurements recorded through 2011 at five tide gauge stations in the Hawaiian Islands (Appendix B). We estimated the rate of relative sea level change at each station from the slope of a least-squares regression line through the timeseries of month-averaged hourly sea level observations. To remove differences from seasonal fluctuations in coastal temperatures, salinities, winds, atmospheric pressures, and ocean currents at each station, we used sea level averages normalized by the mean station value for each month of the year (National Oceanic & Atmospheric Administration, 2009, 2013a, 2013b, 2013c, 2013d, 2013e). We subtracted the globally averaged rate of sea level rise (Church et al., 2004; Church and White, 2006) from the rates of relative sea level change at each station to obtain island vertical motion rates. We interpreted all deviation between the local tide gauge rates and the global average sea level rise rates to correspond to land level

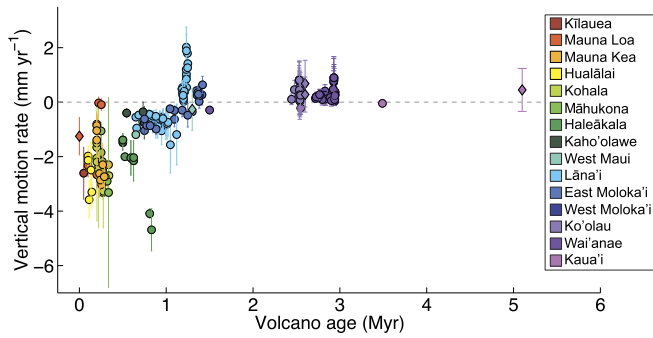


Fig. 4. Rates of island vertical motion $\dot{z}_{\text{obs}}$ ($\pm 2\sigma$ error bars) from tide gauge records (diamond markers) and dated paleoshorelines (circular markers) plotted against volcano age halfway through the time interval over which the rate is averaged. The age on the horizontal axis is calculated as $\tau_v - \frac{1}{2}\tau_{\text{obs}}$, where τ_{obs} is the length of the timespan over which the rate is averaged and τ_v is the age of the volcano on which each rate is observed (Robinson and Eakins, 2006). Four paleoshorelines on Mauna Loa, which predate but lie within uncertainty of Mauna Loa's age τ_v , are plotted at $\frac{1}{2}\tau_{\text{obs}}$.

changes rather than local deviations from eustatic sea level because the Hawaiian Islands are >4000 km from previous extents of the Laurentide ice sheet and therefore subject to minimal (<1 m) glacio-hydro-isostatic deviations from eustatic sea level rise (Dutton and Lambeck, 2012).

3.3. Trends in island vertical motion

Uplift and subsidence rates measured on different volcanoes at comparable times in the lifespan of each volcano suggest that each of the Hawaiian volcanoes has followed a similar vertical motion history over its first ~ 2.5 Myr (Fig. 4). Based on the rate observations, Hawaiian volcanoes appear to (1) subside at rapid and decreasing rates in the first ~ 1 Myr following their construction; (2) uplift beginning ~ 1 – 2.5 Myr following their construction and possibly continuing until ~ 3 Myr following their construction; and (3) subside again within ~ 5 Myr of their construction. Few rate observations constrain the latter stages of island vertical motion.

This observed pattern of rapid but slowing subsidence followed by slower uplift is consistent with the trajectory of vertical motion expected for flexure of the lithosphere under volcanic loading. A volcano initially subsides rapidly under its own weight and then more slowly due to loading of newer neighboring volcanoes. As plate motion carries the volcano away from the center of active volcanism, it eventually reaches the flexural outer bulge of the active volcano where elastic deformation of the lithosphere from far-field loading causes uplift.

The spatial distribution of observed rates averaged over different time intervals also matches expectations for volcanic loading (Fig. 5). Subsidence occurs within and immediately adjacent to the region of active volcanic loading. Regions of uplift remain a similar distance from the volcanic center. For example, Lāna'i and Moloka'i move from the zone of uplift to the zone of subsidence between Figs. 5b and 5c.

4. Modeling island vertical motion rates

To quantify the comparison between observed rates and theoretical predictions, we created simple models of the processes that contribute to vertical motion of the Hawaiian Islands. We calibrated our models using a $1011 \text{ km} \times 939 \text{ km}$ swath of gridded seafloor bathymetry near Hawai'i derived from satellite altimetry and ship depth soundings (Smith and Sandwell, 1997). We first

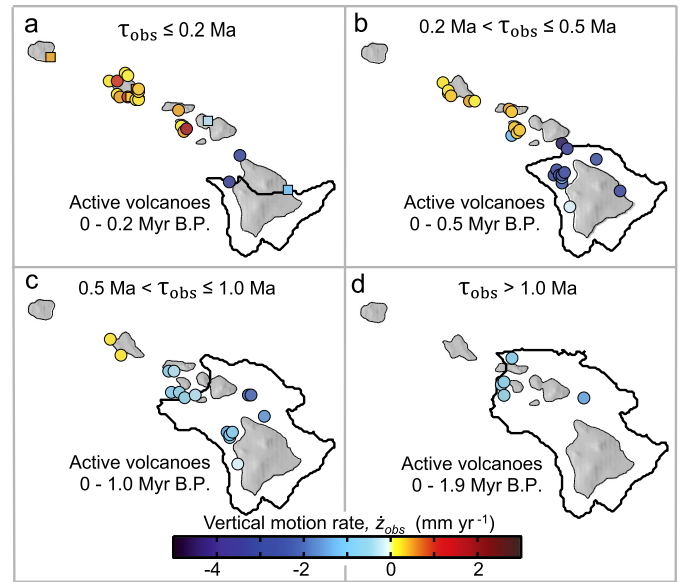


Fig. 5. Geographic distribution of island vertical motion rates $\dot{z}_{\text{obs}}$ that average over different time intervals, plotted atop a shaded relief map of the Hawaiian Islands. Square markers are tide gauge records, which provide modern rates, and circular markers are dated paleoshorelines, which provide rates averaged over a time interval beginning some time τ_{obs} in the past and terminating at the present. Panels correspond to the indicated ranges of τ_{obs} , and black lines bound all volcanoes active between τ_{obs} and the present based on K-Ar ages. Volcanoes are mapped from slope breaks in the bathymetry and subaerial contacts between lithologic units (Robinson and Eakins, 2006). The median value of rates that plot atop one another at the resolution of the map (rates within a ~ 7 km radius of each other) is displayed.

removed depth differences in the bathymetry attributable to long-term lithospheric cooling (Parsons and Sclater, 1977). We then used filters to separate the remaining bathymetry into a regional component associated with the Hawaiian Swell and a residual, more local component associated with the Hawaiian volcanoes and deformed lithosphere.

4.1. Detrending seafloor elevations for age-related depth differences

We calculated the topographic effect of long-term lithospheric cooling (independent of thermal perturbation at the Hawaiian hotspot) as in Parsons and Sclater (1977) using seafloor age estimates linearly interpolated between dated marine magnetic anomalies (Müller et al., 2008). We removed the expected age-related depth differences before filtering the bathymetry. Observed subsidence rates near Hawai'i are much faster than the thermal subsidence rates predicted for the 69 Myr to 103 Myr-old seafloor within the study region, so we neglected the contributions of long-term thermal subsidence to the modeled rates of island vertical motion.

4.2. Decomposition of the bathymetry into regional and residual components

Plume–plate interactions are thought to produce swell topography with wavelengths at least four times the typical diameter of a Hawaiian volcano and three times the width of its flexural moat (e.g. Zhong and Watts, 2002). Exploiting this large difference in scale, we filtered the bathymetry into regional and residual components. We tested two filtering methods, a spectral filter and a median filter. In the spectral domain, we applied a high-pass filter with a cut-off frequency of $1/400 \text{ km}^{-1}$ (e.g. Marks and Sandwell, 1991) and a Gaussian taper with standard deviation $1/3600 \text{ km}^{-1}$ to the discrete Fourier transform of

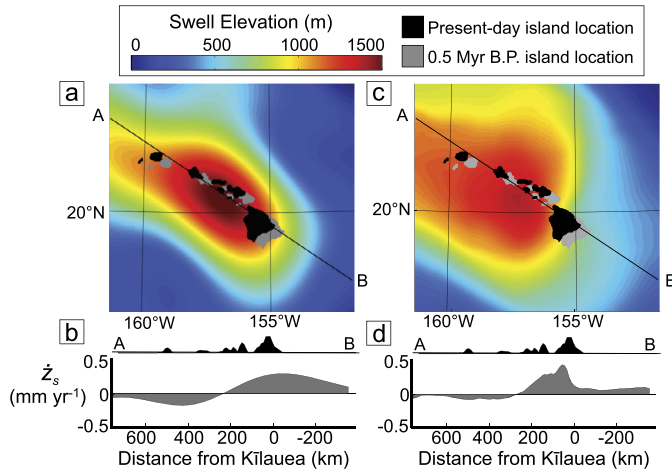


Fig. 6. Estimated Hawaiian swell topography and resulting rates of vertical motion due to island migration over the swell. (a) Swell topography obtained with a spectral filter and (b) corresponding average vertical motion rates due to plate motion over the past 0.5 Myr along the transect A–B. (c) Swell topography extracted with a spatial median filter and (d) corresponding average rates of vertical motion over the past 0.5 Myr.

the unwindowed bathymetry to isolate its high-frequency components (Fig. 6a) (Perron et al., 2008). In the spatial domain, we applied a median filter, which returns the median value of all pixels within a moving window (Fig. 6c). We used a $480 \text{ km} \times 480 \text{ km}$ window size that maximizes the mean amplitude of the residual topography within the region bounded by its zero contour (Wessel, 1998). In both cases, we assumed that the regional component represents the Hawaiian Swell and the residual component represents the Hawaiian volcanoes and the substratum that they displace. We used a bathymetric grid that extended 240 km beyond the study region in all directions to avoid edge effects.

The planform shapes of the swells obtained by both filtering methods mimic the shape of a zone of low seismic wavespeeds that extends to 400 km depth (Wolfe et al., 2009). Their $\sim 600 \text{ km}$ half wavelengths and $\sim 1500 \text{ m}$ amplitudes are also consistent with previous interpretations of the Hawaiian Swell from unfiltered bathymetric profiles (Detrick and Crough, 1978; Crough, 1978; Zhong and Watts, 2002). The peak amplitudes of our swell estimates occur 200–300 km downstream of Kilauea Volcano, as previously observed (Wessel, 1993a) and predicted by models of plume–plate interactions at the Hawaiian hotspot (Zhong and Watts, 2002).

4.3. Vertical motion rates due to island migration over swell topography

We computed a model rate of vertical motion due to migration over swell topography, $\dot{z}_s$, in the fixed hotspot reference frame using a constant plate velocity of 92 mm yr^{-1} along an azimuth 304° east of north (Morgan and Morgan, 2007). We assumed that swell topography remains constant in time as the lithosphere migrates over it and calculated the net vertical motion of each paleoshoreline over its lifespan τ_{obs} moving across both filtered forms of the Hawaiian Swell. Figs. 6b and 6d illustrate the results of an example calculation along a transect parallel to plate motion over the past 0.5 Myr.

4.4. Vertical motion rates due to volcanic loading

We used the residual short-wavelength component of the bathymetry to compute rates of island vertical motion due to volcanic loading $\dot{z}_v$. Over the timescales of shield building ($\sim 0.5 \text{ Myr}$),

we assumed that the lithosphere behaves as a thin elastic plate overlying an inviscid fluid. We estimated the size and distribution of each volcanic load from the residual bathymetry by calculating the load and deflection simultaneously (Sandwell and Smith, 2001) and assuming that all material between the (computed) deflected surface and the (observed) residual bathymetry corresponds to the volcanic load. This may slightly underestimate the volume of the load where its base is buried beneath sediment within the flexural moat. We assumed that, prior to loading, the residual bathymetry had a uniform depth similar to that outside the flexural moat and arch.

We calculated the flexural deflection of the lithosphere and the corresponding volcanic load at each grid point for 15 effective elastic plate thicknesses T_e between 10 km and 80 km and the residual bathymetry from both filtering methods. Because a weak elastic plate and a more rigid elastic plate deflect at different amplitudes and wavelengths under the same load, each assumed rigidity results in a different load distribution and plate deflection to match the observed residual bathymetry. Assuming $T_e = 40 \text{ km}$, the total load of the volcanoes calculated from the spectrally filtered residual bathymetry is $540.5 \times 10^3 \text{ km}^3$ compared to $579.1 \times 10^3 \text{ km}^3$ calculated from the median filtered residual bathymetry.

We assigned the load computed at each point to one of the 17 Hawaiian volcanoes based on volcano boundaries mapped from slope breaks in the bathymetry and subaerial contacts (Robinson and Eakins, 2006). We assumed that the volcanoes do not overlap and that the contacts between the volcanoes are vertically oriented. We also assumed that no significant redistribution has occurred since the volcanoes were emplaced. Thus, for each value of T_e , we could compute the deflection that each volcano produces at each grid point as the sum of the deflections resulting from all of the point loads within the volcano's planform bounds (Fig. 7).

We calculated average vertical motion rates due to volcanic loading $\dot{z}_v$ over intervals extending from the present to times τ_v Myr in the past. We used $\tau_v = [0.2, 0.4, 0.5, 1.0, 1.3, 1.5, 1.9, 2.6, 3.0, 4.9, 5.1] \text{ Myr}$ corresponding to the K–Ar ages of each of the 17 volcanoes (Fig. 7) (Robinson and Eakins, 2006).

We then calculated a total modeled vertical motion rate $\dot{z}_{\text{mod}} = \dot{z}_s + \dot{z}_v$ at the location of each measured vertical motion rate using $\dot{z}_v$ averaged over the nearest timescale τ_v greater than or equal to the observed rate averaging timescale τ_{obs} . We did not calculate a modeled rate corresponding to any of the 15 rate observations from submarine slope breaks because they do not correspond to a specific geographic location.

4.4.1. Flexure calculation assumptions

We assumed a load density of 2800 kg m^{-3} , an upper mantle density of 3300 kg m^{-3} , a Young's modulus of the lithosphere of 10^{11} Pa , and a Poisson's ratio of the lithosphere of 0.25.

We neglected the influence of flexural deformation resulting from post-shield or rejuvenated volcanism. This omission should have only a minor effect on our rate estimates because post-shield and rejuvenated volcanism typically accounts for $< 2\%$ of the total erupted volume of a Hawaiian volcano (Clague and Dalrymple, 1987). We also neglected contributions to surface deflection from volcanic flank collapses, which only elicit an isostatic response in the most extreme cases when a significant volume of material is transported beyond the flexural moat of the volcano (Smith and Wessel, 2000).

In the flexure calculations, we also assumed that the volcanic loads were fully submerged below sea level. We expect this simplification to have only a minor effect on the rates we compute since the volumes of Hawaiian volcanoes are all $< 5\%$ subaerially exposed and more typically $< 1\%$ exposed. Because only the upper-

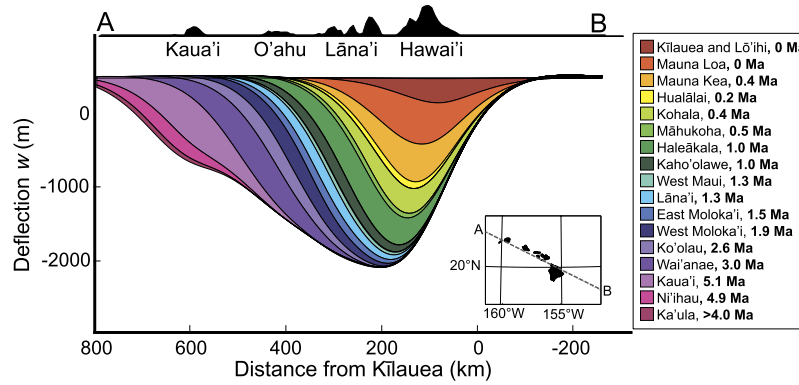


Fig. 7. Flexural response of the lithosphere ($T_e = 40$ km, spectral filtering) to individual volcanic loads. Deflection profiles for each volcano are stacked to fill the total deflection profile along a transect A–B. K–Ar ages of each volcano are given in the legend (Robinson and Eakins, 2006).

most summit of each volcano is exposed above sea level, we also considered isostatic response to glacio-eustatic sea level change to be negligible.

Finally, we modeled the volcanoes as uniform density loads and assumed that their flexural moats did not fill in with sediment after loading. Side-scan sonar (Moore and Clague, 1992) and seismic surveys (ten Brink and Watts, 1985; Rees et al., 1993) have observed sediment deposits up to 2.2 km thick in the moats surrounding the Hawaiian Islands, primarily blocky landslide deposits from volcanic flank collapses. The timings of the large mass wasting events that deposited this sediment are not well constrained, and the extent to which volcanic loads are built upon sediments derived from neighboring volcanoes is unknown (Moore and Clague, 1992). Seismic surveys indicate that sediment within the Hawaiian Deep is thickest near Oʻahu and thins to the south-east near the youngest islands in the chain, where the Deep is significantly underfilled (Rees et al., 1993). These observations suggest that sedimentation occurs progressively. However, without constraints on the timing of sedimentation or the density structure arising from progressive sedimentation on top of progressive volcanic loading, we assumed a uniform density, non-redistributed volcanic load. This simplification may affect the flexures that we calculate on the south side of the Hawaiian Ridge in particular because more submarine landslides have occurred along the south flank of Hawaiʻi than along its northern edge, but this bias is small relative to the observed rates of island vertical motion.

4.5. Model evaluation

We calculated the normalized root mean square error Φ between the modeled rates $\dot{z}_{\text{mod}}$ and the observed rates $\dot{z}_{\text{obs}}$, modified to account for uncertainties in the rate observations $\delta\dot{z}_{\text{obs}}$, which vary considerably for different paleoshoreline dating methods (e.g. mean fractional 2σ uncertainty of 13% for vertical motion rates from paleoshorelines dated by $^{234}\text{U}/^{238}\text{U}$ versus 116% for vertical motion rates from paleoshorelines dated by $^{86}\text{Sr}/^{87}\text{Sr}$):

$$\Phi = \frac{\sqrt{\frac{1}{N} \sum_{n=1}^N \left[\frac{CF_n}{0.5} (\dot{z}_{\text{obs},n} - \dot{z}_{\text{mod},n}) \right]^2}}{\max(\dot{z}_{\text{obs}}) - \min(\dot{z}_{\text{obs}})} \quad (3)$$

where $N = 242$ is the number of rate observations and CF is a correction factor calculated for each observed rate $\dot{z}_{\text{obs}}$ as the probability that $\dot{z}_{\text{obs}} \neq \dot{z}_{\text{mod}}$ based on the area under the normal distribution with mean $\dot{z}_{\text{obs}}$ and standard deviation $\delta\dot{z}_{\text{obs}}$ (Harmel and Smith, 2007).

To assess the significance of the model fits, we compared the model Φ to the Φ obtained when the modeled rates are set to

(a) the mean value of the observed rates ($\dot{z}_{\text{obs}}$) and (b) the mean value of the observed rates within each volcanic loading timescale. These comparisons test whether the observed rates vary in time and space in a manner more consistent with flexure and migration over swell topography than with scatter about a spatially invariant steady rate or about a spatially invariant rate specific to each period of measurement. The latter comparison, in particular, tests the significance of the spatial distribution of modeled rates within each loading timescale rather than the timescale dependence of the rates. Both uplift and subsidence rates show a weak timescale dependence (negative power-law exponents with measurement interval of -0.14 ± 0.04 and -0.04 ± 0.03 , respectively), possibly reflecting unsteadiness in the processes of vertical motion (Sadler, 1993, Appendix C), but this timescale dependence should equally affect the observed and modeled vertical motion rates because they both average over the discrete episodes of volcanic loading and intervening hiatuses leading up to the present day.

5. Time history of island vertical motion from migration over swell topography and volcanic loading

Over hundred thousand to million-year timescales, modeled rates of vertical motion $\dot{z}_{\text{mod}}$ generally match the magnitudes and patterns of observed vertical motion rates $\dot{z}_{\text{obs}}$ (Fig. 8). Modeled subsidence rates averaged over the past 200 kyr are fastest at the Big Island of Hawaiʻi and become progressively slower downstream from Kilauea (in the direction of older volcanism), reaching a minimum subsidence rate or maximum uplift rate (depending on the effective elastic plate thickness T_e) in the central Hawaiian Islands. This minimum subsidence rate or maximum uplift rate is more than ten times slower in absolute magnitude than the maximum subsidence rate. Farther downstream, modeled rates of vertical motion are negligibly small.

Averaged over longer timescales, modeled rates of island vertical motion decrease in magnitude and vary less dramatically in space. The pattern of subsidence and uplift in the modeled vertical motion rates also shifts slightly downstream relative to the present-day locations of the Hawaiian Islands for rates averaged over longer timescales. For instance, maximum subsidence rates occur nearer the present-day location of Maui for rates averaged over 1.9 Myr.

Modeled rates of island vertical motion due to migration over swell topography $\dot{z}_s$ are generally a small fraction of the observed rates. Rates due to swell migration are $|\dot{z}_s| \leq 0.28 \text{ mm yr}^{-1}$ for spectral filtering and $|\dot{z}_s| \leq 0.45 \text{ mm yr}^{-1}$ for spatial median filtering. This is typically a small fraction of the total vertical motion rates, which can be as fast as several millimeters per year (Fig. 6).

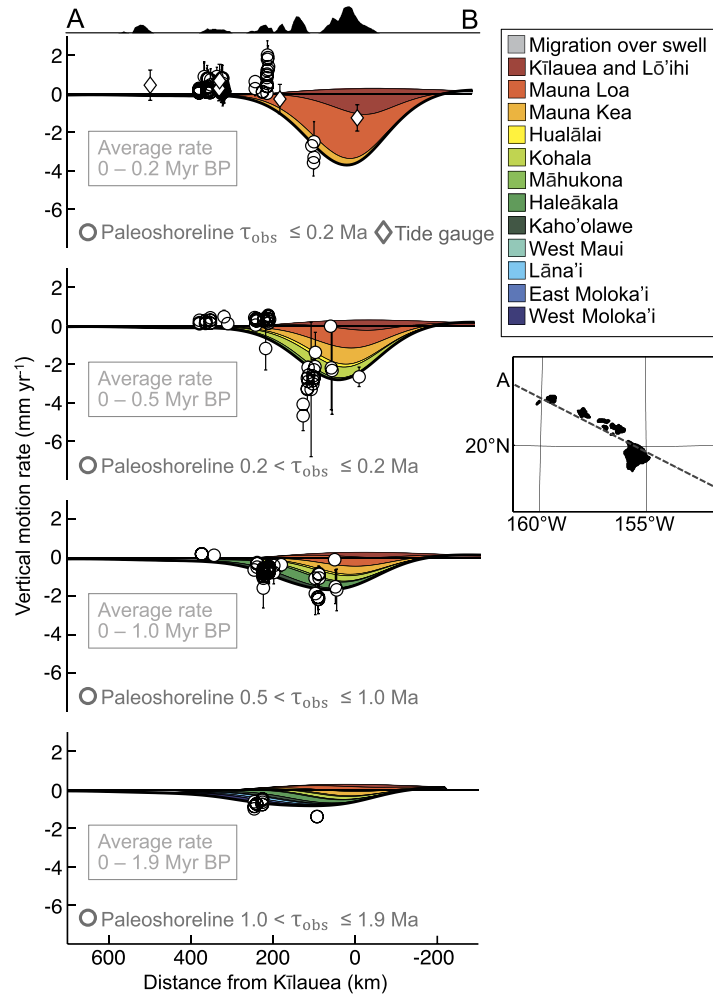


Fig. 8. Comparison of observed rates of vertical motion from tide gauge records (white diamond markers) and dated paleoshorelines (white circular markers) to modeled rates of vertical motion. Modeled rates due to migration over swell topography and flexural isostatic response of the lithosphere ($T_e = 40$ km) to individual volcanic loads (different colors) are stacked to give the total modeled rate (thick black line) along a transect A–B. The four panels show modeled rates averaged over four timescales, $\tau_v = [0.2, 0.5, 1.0, 1.9]$ Myr and compared to observed rates averaged over shorter or equal timescales τ_{obs} . Uplift from migration over swell topography (gray shading) contributes up to 0.5 mm yr^{-1} but is difficult to discern in the figure because this contribution is small relative to the vertical motion from volcanic loading.

Moreover, the modeled rates due to swell migration are inconsistent with the observed patterns of uplift and subsidence. The fastest modeled uplift due to swell migration $\dot{z}_s$ occurs at the locations of the fastest observed subsidence $\dot{z}_{obs}$ (Fig. 8) because swell slopes are steepest on the front face of the swell near Kilauea and shallow as they near the swell apex in the central Hawaiian Islands. Where uplift is observed, modeled uplift rates for migration over swell topography are $\leq 0.05 \text{ mm yr}^{-1}$ for both filtering methods – approximately five times slower than the median observed uplift rate.

Maximum vertical motion rates due to volcanic loading $\dot{z}_v$ are more than an order of magnitude faster than maximum vertical motion rates due to migration over swell topography $\dot{z}_s$. Unlike rates predicted for migration over swell topography, the spatial pattern of rates predicted for volcanic loading generally match the observed magnitudes and patterns of uplift and subsidence. Maximum observed and modeled subsidence rates averaged over short timescales occur near the Big Island of Hawai'i. Averaged over longer timescales, maximum subsidence rates occur nearer Maui. Uplift rates of lesser magnitude occur in the central Hawaiian Islands and, like the subsidence rates, shift downstream to O'ahu and decrease in absolute magnitude when averaged over longer timescales. This close match of observed vertical motion with modeled vertical motion due to volcanic loading suggests that

flexural response of the lithosphere to volcanic loading is the dominant source of Hawaiian Island vertical motion.

6. Constraints on plate strength from vertical motion rates

Because of the dominant influence of volcanic loading, the vertical motion rate observations provide an additional dataset to constrain the elastic strength of the lithosphere at the Hawaiian hotspot. Unbroken elastic plate models constrained by bathymetry, seismic reflection profiles, and gravity anomaly data indicate that the effective elastic plate thickness of the lithosphere at the Hawaiian hotspot lies in the 20 km to 45 km range (Watts, 1978; McNutt and Shure, 1986; Watts and ten Brink, 1989; Wessel, 1993b).

Constrained by the vertical motion rate observations, an effective elastic plate thickness T_e of 40 km minimizes the model misfit ($\Phi = 0.1079$) for spectral filtering. Models with effective elastic thicknesses T_e between 30 km and 80 km provide a better fit to the rate observations than the mean observed rate ($\langle \dot{z}_{obs} \rangle = -0.30 \text{ mm yr}^{-1}$, $\Phi = 0.1316$; Fig. 9), but only models with T_e between 35 km and 45 km provide a better fit to the rate observations than the mean observed rate within each volcanic loading interval ($\Phi = 0.1111$; Fig. 9). This indicates that, over a range of effective elastic thicknesses, the observed rates are more consis-

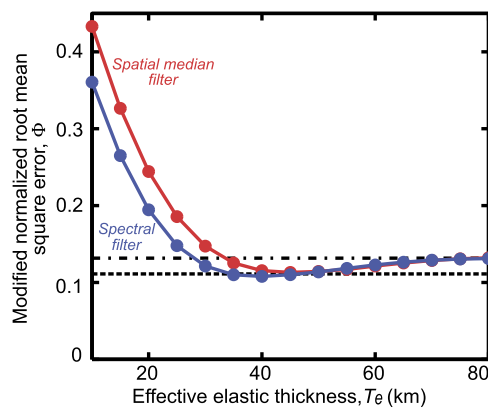


Fig. 9. Effective elastic thickness of the lithosphere T_e versus modified normalized root mean square error Φ of modeled rates of vertical motion in comparison to observed rates of vertical motion from tide gauge records and dated paleoshorelines. Φ is minimized at $T_e = 40$ km ($\Phi = 0.1079$) for swell topography and volcanic loads from spectral filtering of the bathymetry and $T_e = 45$ km ($\Phi = 0.1131$) for swell topography and volcanic loads from spatial median filtering of the bathymetry. Model Φ for spatial median filtering and effective elastic thicknesses T_e between 35 km and 75 km or spectral filtering and T_e between 30 km and 80 km is lower than the $\Phi = 0.1316$ obtained when the modeled rates are set to the mean observed vertical motion rate ($\dot{z}_{\text{obs}}$) (dash-dotted line), but only spectral filtering and T_e between 35 km and 45 km provide a better fit to the rate observations than the mean observed rates within each volcanic loading interval (dotted line, $\Phi = 0.1111$).

tent with the spatially variable deflection predicted for lithospheric flexure than they are with spatially invariant vertical motion. For spatial median filtering, a higher T_e of 45 km minimizes the model misfit ($\Phi = 0.1131$).

These results are on the high end, but within the range of effective elastic thicknesses implied by bathymetry, seismic refraction profiles, and gravity anomalies (20 km to 45 km). Because viscoelastic relaxation of the lithosphere may occur over million year timescales (Watts and Zhong, 2000; Zhong and Watts, 2013) and the vertical motion rates record the response of the lithosphere to loads emplaced in the past <2 Myr, our results may overestimate the “relaxed” effective elastic thickness of the lithosphere. The effective elastic thicknesses obtained from the rate observations are also consistent with the 38 km to 40 km depth of the 600 °C isotherm predicted for 80 Myr to 90 Myr-old oceanic lithosphere by a plate cooling model (Parsons and Sclater, 1977). In a global compilation, 72% of effective elastic thickness estimates for oceanic lithosphere fall within depth of the 300 °C to 600 °C oceanic isotherms (Watts and Zhong, 2000).

Our primary result – that vertical motions of the Hawaiian Islands follow a relatively predictable trajectory of vertical motion set by the history of volcanism at the Hawaiian hotspot – is robust with respect to (a) various interpretations of the radiometric ages τ_v and their relationship to the sequence of volcanic loading and deflection (Appendix D.1), (b) different binnings of the rate observations (Appendix D.2), and (c) reweighting of the rate observations to account for oversampling of paleoshorelines on Lānaʻi and Oʻahu (Appendix D.3).

7. Anomalous uplift at Lānaʻi and Molokaʻi

Modeled rates of vertical motion are inconsistent with the rapid uplift rates observed at Lānaʻi and Molokaʻi over the past 200 kyr (Fig. 8). Uplift rates calculated from presumed paleoshorelines on Lānaʻi and Molokaʻi are as fast as 2.02 ± 0.74 mm yr^{−1}, nearly six times faster than the 0.34 mm yr^{−1} maximum uplift rates that our models can produce at Lānaʻi, using spatial median filtering and $T_e = 10$ km.

The inability of our models to explain these rapid uplift rates may reflect inaccurate interpretations of the elevated coral deposits on Lānaʻi and Molokaʻi. Various non-tectonic emplacement mechanisms have been proposed to explain the modern elevation of coral on Molokaʻi and Lānaʻi. Moore and Moore (1984) proposed that elevated corals on Molokaʻi and Lānaʻi were deposited by a giant tsunami or series of tsunamis generated by a flank collapse of the southeastern edifice of Lānaʻi. However, subsequent workers (Grigg and Jones, 1997; Rubin et al., 2000) have refuted the “giant wave hypothesis” based on (1) the presence of coral in growth position, wave-cut notches and terraces, and horizontally bedded deposits on Lānaʻi and Molokaʻi, (2) sedimentary and chronostratigraphic evidence for multiple depositional events within the elevated Hulopoʻe gravel formation on Lānaʻi, (3) the trend of increasing coral age with elevation above present sea level in deposits on Molokaʻi, and (4) calculations yielding insufficient landslide-triggered tsunami wave heights to explain the present elevations of emerged deposits. These studies propose instead that elevated coral deposits on Molokaʻi and Lānaʻi were either uplifted by lithospheric flexure (Grigg and Jones, 1997) or by some combination of flexure and run-up from seasonal waves, storm events, or “normal” tsunamis (Rubin et al., 2000). Some of the deposits may have also been transported upslope by humans or wind (Grigg and Jones, 1997).

Our results are consistent with the suggestion that migration over swell topography may help explain some observed uplift rates ≤ 0.2 mm yr^{−1} at Lānaʻi and Molokaʻi that cannot be accounted for by flexure alone (Zhong and Watts, 2002). Uplift rates associated with migration over the swell topography $\dot{z}_s$ may be as fast as 0.05 mm yr^{−1} at the paleoshoreline locations and timescales on Lānaʻi and Molokaʻi. Yet, our calculations indicate that a low effective elastic thicknesses ($T_e = 15$ km for spectral filtering or $T_e = 20$ km for median filtering) is required to generate uplift >0.2 mm yr^{−1} on Lānaʻi and Molokaʻi, and the pattern and magnitudes of modeled uplift and subsidence for these low T_e values are inconsistent with rate observations elsewhere along the Hawaiian chain. Neither lithospheric flexure nor swell migration can explain the fastest uplift rates on Lānaʻi and Molokaʻi.

Assuming all elevated coral deposits on Lānaʻi and Molokaʻi were emplaced above sea level by seasonal, storm, or tsunami waves (Moore and Moore, 1984, 1988; Rubin et al., 2000) or transported upslope by humans or wind (Grigg and Jones, 1997), we repeated our calculations excluding the 44 vertical motion rates $\dot{z}_{\text{obs}}$ calculated from uplifted deposits on Lānaʻi and Molokaʻi. Comparison of the model rates with the mean rate did not change significantly when the observed uplift rates at Lānaʻi and Molokaʻi were excluded: T_e between 35 km and 80 km for spatial median filtering and T_e between 30 km and 80 km for spectral filtering provided a better fit to the remaining 198 observed rates than their mean value ($\langle \dot{z}_{\text{obs}} \rangle = -0.48$ mm yr^{−1}, $\Phi = 0.1470$). However, excluding Lānaʻi and Molokaʻi uplift rates expanded the range of effective elastic plate thicknesses T_e with lower misfit to the observed rates than the mean rate within each volcanic loading interval ($\Phi = 0.1321$): T_e between 40 km and 70 km for spatial median filtering and T_e between 35 km and 65 km for spectral filtering. Excluding Lānaʻi and Molokaʻi uplift rates also yielded slightly higher optimal effective elastic thicknesses T_e of 45 km ($\Phi = 0.0999$) and 50 km ($\Phi = 0.0982$) for spectral and spatial median filtering, respectively.

8. Swell evolution suggested by non-flexural vertical motion rates

If elevated coral deposits on Lānaʻi and Molokaʻi were not transported upslope by humans, waves, or wind, their anomalous

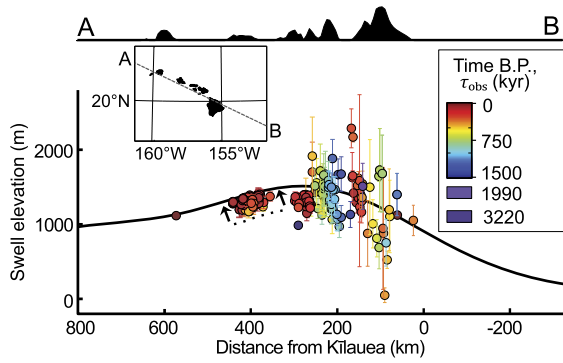


Fig. 10. Paleo swell elevations ($\pm 2\sigma$ error bars, measured relative to the regional seafloor depth beyond the extents of the swell) estimated by multiplying non-flexural rates of vertical motion (observed rates $\dot{z}_{\text{obs}}$ minus rates predicted for volcanic loading $\dot{z}_v$) by the timescale τ_{obs} over which each rate is averaged. $\dot{z}_v$ was calculated from the flexure model that best fits the rate observations ($T_e = 40$ km, spectral filtering of the bathymetry). Each paleo swell elevation corresponds to the elevation of the swell at the location of a paleoshoreline τ_{obs} years before present assuming constant 92 mm yr^{-1} plate velocity along a 304° transect A–B. Paleoshoreline locations are plotted at their distance from Kilauea along the transect A–B. Color of marker indicates time before present τ_{obs} of each paleo swell elevation estimate, and black line displays modern swell elevations. Dotted line and arrows schematically illustrate a possible trend of increasing swell elevation on the downstream slope of the swell suggested by systematically lower ~ 500 kyr paleo swell elevations than < 500 kyr paleo swell elevations at locations on O’ahu.

lous uplift could have resulted from transient changes in swell topography. Numerical models have examined the dynamics of hot plume material that is delivered to the top of the asthenosphere, ponded and entrained beneath the moving lithosphere, and carried downstream of the Hawaiian hotspot (Ribe and Christensen, 1994; Moore et al., 1998). These models typically identify three regimes for the exchange of heat and material between the lithosphere and the plume: (1) an “underplating” regime in which plume material stays in the asthenosphere and forms a steady plume train that exchanges only minor amounts of heat with the lithosphere by conduction; (2) a steady shear-aligned roll regime in which a time-invariant plume train exchanges heat with the lowermost lithosphere by convection, sustained by the descent of thin sheets of cool, delaminated lithosphere aligned with plate motion; and, finally, (3) an unsteady regime in which vigorous secondary convection escalates with the downwelling of thicker sheets and drips of lithospheric material and increased heat exchange to a significantly thinned lithosphere.

High downstream variability in heat flux (Von Herzen et al., 1989), high admittance (Marks and Sandwell, 1991), and seismic observations (Wolfe et al., 2009) support a compensation model for the Hawaiian Swell with both dynamic and thermal contributions (regimes (2) or (3)), but provide no time-varying observations to constrain the behavior of secondary convection (i.e. (2) vs. (3)). To determine if transient changes in swell topography might be recorded in our data (e.g. Ramalho et al., 2010a; Ramalho, 2011), we multiplied the non-flexural rates of vertical motion ($\dot{z}_{\text{obs}} - \dot{z}_v$) by the timescale τ_{obs} to determine the residual vertical displacement Δz_s that might be associated with migration over constant swell topography or transient changes in swell amplitude. If Δz_s resembles the present-day shape of the swell, it suggests that the swell topography has been constant through time. If Δz_s deviates from the present-day shape of the swell, it provides an estimate of transient changes in the swell.

The paleo swell elevations thus computed (Fig. 10) imply that swell elevations near the present-day locations of Moloka’i (~ 300 km from Kilauea along the transect A–B) and O’ahu (~ 400 km from Kilauea along the transect A–B) have increased

more rapidly than would be expected from passing of a constant swell over the past ~ 500 kyr (Fig. 10). This growth might be related to the increase in magma supply rates during construction of the Big Island of Hawai’i over the past ~ 1 Myr compared to supply rates in the preceding period (Robinson and Eakins, 2006). However, paleo swell elevation estimated over longer timescales and nearer the island of Hawai’i do not display coherent signals of transient swell morphology. In some locations, the uncertainties in these estimates span ~ 3 km due to the large uncertainties in the observed rates of vertical motion (Fig. 10). We conclude that it may be infeasible to detect changes in swell topography near the Hawaiian hotspot where the contribution of swell topography to rates of island vertical motion is much smaller than the contribution from volcanic loading.

9. Conclusions

Our comparisons of modeled and observed uplift and subsidence rates in the Hawaiian Islands suggest that flexural response of the lithosphere to volcanic loading is the primary mechanism of island vertical motion. Island migration over the Hawaiian Swell makes a relatively minor contribution to uplift and subsidence. The dominant influence of volcanic loading makes it possible to recover an effective elastic plate thickness of the lithosphere from an inversion of rate observations. Rates of island vertical motion from dated paleoshorelines and tide gauge records constrain the effective elastic thickness of the lithosphere near the Hawaiian hotspot to 40 km to 45 km, consistent with values predicted by plate cooling models and determined previously using gravity and seismic data.

Neither volcanic loading nor migration over swell topography can explain high uplift rates in the central Hawaiian Islands. These anomalous rates may result from transient changes in swell topography. Residual vertical motion rates suggest a possible trend of downstream swell growth over the past ~ 500 kyr but they reveal no coherent signals of swell evolution nearer the Hawaiian hotspot or over longer timescales.

Overall, these results imply that Hawaiian Islands follow a relatively predictable trajectory of vertical motion over their initial ~ 3 Myr set by the history of volcanic loading at the Hawaiian hotspot. Once islands have passed beyond the flexural bulge of volcanic loading, migration over swell topography, thermal contraction of the lithosphere if it has been drastically reheated by the mantle plume (Menard, 1983; Caplan-Auerbach et al., 2000) or viscoelastic relaxation of the lithosphere under the load of the volcanoes (Watts and Zhong, 2000) may be the dominant mechanisms of island vertical motion. Our approach of modeling island vertical motion as a consequence of plate motion relative to an evolving lithospheric deflection field offers a framework for modeling vertical motion in numerous other hotspot island chains over timescales relevant to the topographic, microclimatic, and biogeographic evolution of volcanic ocean islands.

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Appendix A. Vertical motion rates from dated paleoshorelines

See Table 1.

Table 1
Vertical motion rates from dated paleoshorelines.

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
	Slope break			−300	Kilauea K-Ar	100 $\pm$ 10 ^a	−2.75 $\pm$ 0.43	Moore and Clague (1992)
T276-R5	Coralline algae (nodule)	19.47	−155.94	−150	Mauna Loa ¹⁴ C	11.82 $\pm$ 0.97 ^b	−5.48 $\pm$ 4.86	Webster et al. (2004)
P5-67-3	<i>Porites</i>	19.47	−156.1	−150 ^c	¹⁴ C	13.48 $\pm$ 0.43 ^b	−2.55 $\pm$ 2.51	Moore et al. (1990b)
P5-67-5a	Coralline algae (crust limestone)	19.47	−156.1	−150 ^c	¹⁴ C	12.72 $\pm$ 0.96 ^b	−3.43 $\pm$ 4.27	Moore et al. (1990b)
T291-R4	<i>Porites</i>	19.47	−155.94	−150 ^c	U–Th	15.24 $\pm$ 0.1	−1.97 $\pm$ 1.99	Webster et al. (2004)
T291-R9	<i>Porites</i>	19.46	−155.94	−150 ^c	U–Th	15.07 $\pm$ 0.11	−1.98 $\pm$ 2.02	Webster et al. (2004)
T291-R16	<i>Porites</i>	19.46	−155.94	−150 ^c	U–Th	14.74 $\pm$ 0.09	−1.99 $\pm$ 2.06	Webster et al. (2004)
T291-R17A	<i>Porites</i>	19.48	−155.96	−150	U–Th	15.07 $\pm$ 0.14	−1.98 $\pm$ 2.02	Webster et al. (2004)
T291-R7A	<i>Porites</i>	19.47	−155.94	−150 ^c	U–Th	15.25 $\pm$ 0.14	−1.97 $\pm$ 2	Webster et al. (2004)
T291-R8A	<i>Porites</i>	19.47	−155.94	−150 ^c	U–Th	14.87 $\pm$ 0.11	−1.99 $\pm$ 2.04	Webster et al. (2004)
T291-R8	Coralline algae (crust limestone)	19.47	−155.94	−150 ^c	¹⁴ C	14.29 $\pm$ 1.05 ^b	−2.11 $\pm$ 2.54	Webster et al. (2004)
M-162-3	<i>Porites</i>	19.47	−155.94	−150 ^c	¹⁴ C	14.72 $\pm$ 1.01 ^b	−1.99 $\pm$ 2.32	Moore and Fornari (1984)
M-170-5A	<i>Porites</i>	19.48	−156.1	−150 ^c	¹⁴ C	15.8 $\pm$ 0.5 ^b	−1.95 $\pm$ 2.19	Moore et al. (1990b)
M-170-5B	<i>Porites</i>	19.48	−156.1	−150 ^c	¹⁴ C	15.7 $\pm$ 0.5 ^b	−1.96 $\pm$ 2.12	Moore et al. (1990b)
M-170-5C	<i>Porites</i>	19.48	−156.1	−150 ^c	¹⁴ C	15.8 $\pm$ 0.46 ^b	−1.95 $\pm$ 2.16	Moore and Fornari (1984)
M-169-1	<i>Porites</i>	19.47	−155.94	−150 ^c	¹⁴ C	14.12 $\pm$ 1.23 ^b	−2.2 $\pm$ 2.72	Moore and Fornari (1984)
T291-R12D	Coralline algae (crust limestone)	19.47	−155.94	−150 ^c	¹⁴ C	13.8 $\pm$ 0.44 ^b	−2.37 $\pm$ 2.45	Webster et al. (2004)
P5-78-5	<i>Porites lobata</i> ?	18.86	−155.94	−150 ^c	U–Th	13.6 $\pm$ 0.3	−2.49 $\pm$ 2.4	Moore et al. (1990b)
P5-78-4	<i>Porites</i>	18.86	−155.87	−150 ^c	¹⁴ C	14.08 $\pm$ 1.04 ^b	−2.22 $\pm$ 2.65	Moore et al. (1990b)
P5-77-7	Coralline algae (crust limestone)	18.89	−155.66	−150 ^c	¹⁴ C	11.5 $\pm$ 0.76 ^b	−6.18 $\pm$ 4.27	Moore et al. (1990b)
P5-77-6	Coralline algae (nodule limestone)	18.89	−155.66	−150 ^c	¹⁴ C	10.29 $\pm$ 0.42 ^b	−8.17 $\pm$ 3.64	Moore et al. (1990b)
P5-77-3	Coralline algae (nodule limestone)	18.88	−155.67	−150 ^c	¹⁴ C	12.02 $\pm$ 0.53 ^b	−5 $\pm$ 3.91	Moore et al. (1990b)
M-170-5	<i>Porites</i>	19.48	−156.1	−150 ^c	²³⁴ U/ ²³⁸ U	17 $\pm$ 5	−0.81 $\pm$ 3.72	Ludwig et al. (1991)
M-170-5	<i>Porites</i>	19.48	−156.1	−150 ^c	U–Th	15.8 $\pm$ 7	−1.95 $\pm$ 6.37	Ludwig et al. (1991)
P5-78-5	<i>Porites lobata</i>	18.86	−155.69	−150 ^c	²³⁴ U/ ²³⁸ U	19 $\pm$ 5	−0.21 $\pm$ 2.69	Ludwig et al. (1991)
P5-78-5	<i>Porites lobata</i>	18.86	−155.69	−150 ^c	U–Th	14 $\pm$ 5	−2.26 $\pm$ 7.01	Ludwig et al. (1991)
T291-R16	<i>Porites</i>	19.48	−155.96	−150 ^c	⁸⁷ Sr/ ⁸⁶ Sr	234 $\pm$ 340		Webster et al. (2010)
T291-R4	Coralline algae?	19.47	−155.94	−150 ^c	⁸⁷ Sr/ ⁸⁶ Sr	506 $\pm$ 320	−0.09 $\pm$ 0.23	Webster et al. (2010)
T291-R4	Coralline algae?	19.47	−155.94	−150 ^c	⁸⁷ Sr/ ⁸⁶ Sr	264 $\pm$ 330		Webster et al. (2010)
T291-R8	Coralline algae?	19.47	−155.94	−150 ^c	⁸⁷ Sr/ ⁸⁶ Sr	445 $\pm$ 250	−0.03 $\pm$ 0.22	Webster et al. (2010)
R25-0.80	Coral fragments	19.73	−155.05 ^d	−27	U–Th	1.62 $\pm$ 0.11	−1.48 $\pm$ 18.86	Moore et al. (1996)
R25-1.45	Coral fragments	19.73	−155.05 ^d	−27.2	U–Th	1.74 $\pm$ 0.1	−1.73 $\pm$ 17.49	Moore et al. (1996)
R34-1.10	Coral fragments	19.73	−155.05 ^d	−42.4	U–Th	8.19 $\pm$ 0.34	1.31 $\pm$ 4.34	Moore et al. (1996)
R34-1.50	Coral fragments	19.73	−155.05 ^d	−42.6	U–Th	8.35 $\pm$ 0.5	1.51 $\pm$ 4.58	Moore et al. (1996)
R40-0.00	Coral fragments	19.73	−155.05 ^d	−52.2	U–Th	9.69 $\pm$ 0.11	1.17 $\pm$ 3.15	Moore et al. (1996)
R25-0.08	Coral fragments	19.73	−155.05 ^d	−27	¹⁴ C	1.92 $\pm$ 0.16 ^b	−1.38 $\pm$ 15.97	Moore et al. (1996)
R26-1.00	Coral fragments	19.73	−155.05 ^d	−27.2	¹⁴ C	2.37 $\pm$ 0.17 ^b	−0.66 $\pm$ 12.94	Moore et al. (1996)
R29-0.80	Coral fragments	19.73	−155.05 ^d	−30.8	¹⁴ C	2.55 $\pm$ 0.17 ^b	−1.83 $\pm$ 12.13	Moore et al. (1996)
R34-1.10	Coral fragments	19.73	−155.05 ^d	−42.4	¹⁴ C	8.03 $\pm$ 0.19 ^b	1.08 $\pm$ 4.12	Moore et al. (1996)
R36-2.00	Coral fragments	19.73	−155.05 ^d	−42.7	¹⁴ C	8.19 $\pm$ 0.15 ^b	1.28 $\pm$ 3.95	Moore et al. (1996)
R35-3.00	Coral fragments	19.73	−155.05 ^d	−46.1	¹⁴ C	8.26 $\pm$ 0.16 ^b	0.96 $\pm$ 3.94	Moore et al. (1996)
R37-1.20	Coral fragments	19.73	−155.05 ^d	−48.6	¹⁴ C	8.21 $\pm$ 0.13 ^b	0.58 $\pm$ 3.91	Moore et al. (1996)
R39-2.40	Coral fragments	19.73	−155.05 ^d	−51.4	¹⁴ C	9.28 $\pm$ 0.21 ^b	1.18 $\pm$ 3.34	Moore et al. (1996)
R40-0.00	Coral fragments	19.73	−155.05 ^d	−52.2	¹⁴ C	10.11 $\pm$ 0.17 ^b	1.24 $\pm$ 3.07	Moore et al. (1996)

Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
R40-0.20	Coral fragments	19.73	−155.05 ^d	−52.2	¹⁴ C	9.95 $\pm$ 0.25 ^b	1.21 $\pm$ 3.13	Moore et al. (1996)
R65-5.00	Coral fragments	19.73	−155.05 ^d	−86.5	¹⁴ C	10.65 $\pm$ 0.12 ^b	−1.56 $\pm$ 2.97	Moore et al. (1996)
HSDP-2	Subaerial- submarine erupted basalt	19.73	−155.05 ^d	−1079	⁴⁰ Ar/ ³⁹ Ar	402 $\pm$ 50	−2.66 $\pm$ 0.51	Sharp and Renne (2005)
	Slope break			−500	K-Ar <i>Mauna Kea</i>	200 $\pm$ 20 ^a	−2.53 $\pm$ 0.45	Moore and Clague (1992)
Dr-9-A	<i>Porites lobata</i>	20.04	−156.07	−693 ^c	²³⁴ U/ ²³⁸ U	225 $\pm$ 12	−2.74 $\pm$ 0.44	Ludwig et al. (1991)
Dr-9-A	<i>Porites lobata</i>	20.04	−156.07	−693 ^c	U–Th	261 $\pm$ 55	−2.3 $\pm$ 0.78	Ludwig et al. (1991)
Dr-9-F	<i>Porites</i>	20.04	−156.07	−693 ^c	²³⁴ U/ ²³⁸ U	226 $\pm$ 13	−2.72 $\pm$ 0.45	Ludwig et al. (1991)
Dr-9-F	<i>Porites</i>	20.04	−156.07	−693 ^c	U–Th	>260		Ludwig et al. (1991)
P5-69-4	<i>Porites lobata</i>	19.99	−156.13	−945 ^c	²³⁴ U/ ²³⁸ U	314 $\pm$ 10	−2.86 $\pm$ 0.3	Ludwig et al. (1991)
P5-69-4	<i>Porites lobata</i>	19.99	−156.13	−945 ^c	U–Th	337 $\pm$ 79	−2.62 $\pm$ 0.87	Ludwig et al. (1991)
P5-69-6	<i>Porites lobata</i>	19.98	−156.13	−945 ^c	²³⁴ U/ ²³⁸ U	287 $\pm$ 10	−3.03 $\pm$ 0.28	Ludwig et al. (1991)
T278-R25	Branching <i>Porites</i>	20	−156.03	−430 ^c	⁸⁷ Sr/ ⁸⁶ Sr	53 $\pm$ 390		Webster et al. (2010)
T286-R4	<i>Porites compressa</i>	19.94	−156.09	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	555 $\pm$ 230	−1.05 $\pm$ 0.58	Webster et al. (2010)
T286-R21	<i>Porites compressa</i>	19.94	−156.09	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	422 $\pm$ 250	−1.39 $\pm$ 1.04	Webster et al. (2010)
T286-R26	<i>Porites compressa</i>	19.94	−156.09	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	−418 $\pm$ 560		Webster et al. (2010)
T291-R9	Coralline algae, nodule facies	19.92	−156.1	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	295 $\pm$ 330		Webster et al. (2010)
T283-R6	Matrix sediments	19.92	−156.1	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	234 $\pm$ 340		Webster et al. (2010)
T283-R27	Matrix sediments	19.92	−156.1	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	127 $\pm$ 370		Webster et al. (2010)
T283-R25	Coralline algae, nodule facies	19.92	−156.1	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	658 $\pm$ 210	−0.86 $\pm$ 0.38	Webster et al. (2010)
T279-R20	<i>Porites compressa</i>	20.01	−156.07	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	160 $\pm$ 360		Webster et al. (2010)
T279-R11	<i>Porites</i>	20.01	−156.07	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	727 $\pm$ 260	−0.81 $\pm$ 0.39	Webster et al. (2010)
T279-R3	<i>Porites compressa</i>	20.01	−156.07	−693 ^c	⁸⁷ Sr/ ⁸⁶ Sr	199 $\pm$ 350		Webster et al. (2010)
	Slope break			−375	K-Ar <i>Hualālai</i>	130 $\pm$ 13 ^a	−2.54 $\pm$ 0.96	Moore and Clague (1992)
P5-70-4A	<i>Porites lobata</i>	19.86	−156.13	−430 ^c	²³⁴ U/ ²³⁸ U	133 $\pm$ 10	−2.49 $\pm$ 1.05	Ludwig et al. (1991)
P5-70-4A	<i>Porites lobata</i>	19.86	−156.13	−430 ^c	U–Th	112 $\pm$ 6	−3.3 $\pm$ 0.61	Ludwig et al. (1991)
P5-68-2B	<i>Porites lobata</i>	19.88	−156.13	−693 ^c	²³⁴ U/ ²³⁸ U	276 $\pm$ 9	−2.13 $\pm$ 0.24	Ludwig et al. (1991)
P5-68-2B	<i>Porites lobata</i>	19.88	−156.13	−693 ^c	U–Th	177 $\pm$ 15	−3.58 $\pm$ 0.69	Ludwig et al. (1991)
	Slope break			−430	Stratigraphic <i>Kohala</i>	200 $\pm$ 20 ^a	−2.18 $\pm$ 0.41	Robinson and Eakins (2006)
M-287-5	Coralline algae	20.33	−155.93 ^d	−360	U–Th	120 $\pm$ 5	−2.7 $\pm$ 0.45	Szabo and Moore (1986)
M-287-3	Coralline algae	20.33	−155.93 ^d	−360	U–Th	99 $\pm$ 4	−3.05 $\pm$ 0.57	Szabo and Moore (1986)
T301-R31	<i>Porites</i>	20.27	−155.49	−945 ^c	⁸⁷ Sr/ ⁸⁶ Sr	395 $\pm$ 350	−2.2 $\pm$ 2.18	Webster et al. (2010)
T301-R39	<i>Leptastrea</i> <i>transversa</i>	20.27	−155.49	−945 ^c	⁸⁷ Sr/ ⁸⁶ Sr	555 $\pm$ 270	−1.5 $\pm$ 0.88	Webster et al. (2010)
T301-R46	<i>Porites lobata</i> ?	20.27	−155.49	−945 ^c	⁸⁷ Sr/ ⁸⁶ Sr	234 $\pm$ 440		Webster et al. (2010)
T302-R1	<i>Porites compressa</i>	20.26	−155.47	−945 ^c	⁸⁷ Sr/ ⁸⁶ Sr	506 $\pm$ 280	−1.66 $\pm$ 1.08	Webster et al. (2010)
T302-R30	<i>Montipora</i>	20.25	−155.48	−945 ^c	⁸⁷ Sr/ ⁸⁶ Sr	359 $\pm$ 310	−2.33 $\pm$ 2.28	Webster et al. (2010)
T302-R35	<i>Porites compressa</i>	20.25	−155.48	−945 ^c	⁸⁷ Sr/ ⁸⁶ Sr	359 $\pm$ 360		Webster et al. (2010)
T302-R38	<i>Porites compressa</i>	20.25	−155.48	−945 ^c	⁸⁷ Sr/ ⁸⁶ Sr	127 $\pm$ 370		Webster et al. (2010)
T304-R3	<i>Porites compressa</i>	20.24	−155.7	−150 ^c	⁸⁷ Sr/ ⁸⁶ Sr	270 $\pm$ 480		Webster et al. (2010)
P5-59	<i>Porites compressa</i>	20.47	−156.02 ^d	−945	ESR	248 $\pm$ 78	−3.27 $\pm$ 1.35	Jones (1995)
P5-59	<i>Porites compressa</i>	20.47	−156.02 ^d	−945	²³⁴ U/ ²³⁸ U	259 $\pm$ 6	−3.32 $\pm$ 0.26	B. Szabo cited in Jones (1995)
	Slope break			−1000	K-Ar <i>Māhukona</i>	430 $\pm$ 40	−2.05 $\pm$ 0.38	Moore and Clague (1992)
D-21	<i>Porites lobata</i>	20.1	−156.21	−1146 ^c	²³⁴ U/ ²³⁸ U	406 $\pm$ 12	−2.73 $\pm$ 0.22	Ludwig et al. (1991)
P5-73-2	<i>Porites lobata</i>	20.08	−156.2	−1146 ^c	²³⁴ U/ ²³⁸ U	360 $\pm$ 12	−2.9 $\pm$ 0.24	Ludwig et al. (1991)
P5-73-1	<i>Porites lobata</i>	20.09	−156.2	−1146 ^c	²³⁴ U/ ²³⁸ U	475 $\pm$ 38	−2.17 $\pm$ 0.33	Ludwig et al. (1991)

(continued on next page)

Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
P5-73-1	<i>Porites lobata</i>	20.09	-156.2	-1146 ^c	U-Th	>320		Ludwig et al. (1991)
P5-75-3	<i>Porites lobata</i>	19.98	-156.24	-1336 ^c	²³⁴ U/ ²³⁸ U	463 $\pm$ 8	-2.57 $\pm$ 0.36	Ludwig et al. (1991)
T293-R29	<i>Porites compressa</i>	19.99	-156.17	-1146 ^c	⁸⁷ Sr/ ⁸⁶ Sr	996 $\pm$ 230	-1.06 $\pm$ 0.31	Webster et al. (2010)
T293-R47	<i>Porites</i>	19.99	-156.16	-1146 ^c	⁸⁷ Sr/ ⁸⁶ Sr	580 $\pm$ 300	-1.85 $\pm$ 1.1	Webster et al. (2010)
T293-R12	Branching <i>Porites</i>	19.99	-156.17	-1146 ^c	⁸⁷ Sr/ ⁸⁶ Sr	55 $\pm$ 320		Webster et al. (2010)
T293-R5	<i>Porites</i>	19.99	-156.17	-1146 ^c	⁸⁷ Sr/ ⁸⁶ Sr	330 $\pm$ 320	-3.32 $\pm$ 3.49	Webster et al. (2010)
	Slope break			-1250	K-Ar	463 $\pm$ 16	-2.49 $\pm$ 0.13	Moore and Clague (1992)
					<i>Haleākala</i>			
P5-64		20.55	-156.08 ^d	-1650	ESR	338 $\pm$ 36	-4.69 $\pm$ 0.77	Jones (1995)
P5-64		20.55	-156.08 ^d	-1650	²³⁴ U/ ²³⁸ U	380 $\pm$ 7	-4.09 $\pm$ 0.17	B. Szabo cited in Jones (1995)
D32	Coralline material	20.65	-155.75 ^d	-1650 ^c	²³⁴ U/ ²³⁸ U	750 $\pm$ 13	-2.04 $\pm$ 0.09	Moore et al. (1990a)
F2-88-D32A	<i>Leptoseris</i>	20.65	-155.73	-1750	⁸⁷ Sr/ ⁸⁶ Sr ⁱ	752 $\pm$ 224	-2.15 $\pm$ 0.74	Faichney et al. (2010)
	<i>hawaiiensis</i>							
F2-88-D32B	<i>Leptoseris</i>	20.65	-155.73	-1750	⁸⁷ Sr/ ⁸⁶ Sr ⁱ	1188 $\pm$ 157	-1.39 $\pm$ 0.24	Faichney et al. (2010)
	<i>hawaiiensis</i>							
F2-88-D32S	<i>Leptoseris</i>	20.65	-155.73	-1750	⁸⁷ Sr/ ⁸⁶ Sr ⁱ	814 $\pm$ 214	-2.05 $\pm$ 0.63	Faichney et al. (2010)
	<i>hawaiiensis</i>							
F2-88-D32A	<i>Leptoseris</i>	20.65	-155.73	-1750	⁸⁷ Sr/ ⁸⁶ Sr	752 $\pm$ 224	-2.15 $\pm$ 0.74	Faichney et al. (2010)
	<i>hawaiiensis</i>							
F2-88-D32B	<i>Leptoseris</i>	20.65	-155.73	-1750	⁸⁷ Sr/ ⁸⁶ Sr	1188 $\pm$ 157	-1.39 $\pm$ 0.24	Faichney et al. (2010)
	<i>hawaiiensis</i>							
F2-88-D32S	<i>Leptoseris</i>	20.65	-155.73	-1750	⁸⁷ Sr/ ⁸⁶ Sr	814 $\pm$ 214	-2.05 $\pm$ 0.63	Faichney et al. (2010)
	<i>hawaiiensis</i>							
	Slope break			-1900	K-Ar	950 $\pm$ 95 ^a	-1.98 $\pm$ 0.25	Moore and Clague (1992)
				-1600	West Mau'i K-Ar	1320 $\pm$ 80	-1.18 $\pm$ 0.1	Clague and Dalrymple (1987)
					<i>Kaho'olawe</i>			
T312-R3	Matrix sediments, pelagic facies	20.64	-156.71	-286.4	⁸⁷ Sr/ ⁸⁶ Sr	295 $\pm$ 370		Webster et al. (2010)
P5-254-1A	<i>Porites lobata</i> ?	20.62	-156.71	-305	⁸⁷ Sr/ ⁸⁶ Sr	533 $\pm$ 310	-0.35 $\pm$ 0.36	Webster et al. (2010)
MVD9-R2	Coralline algae?	20.61	-156.91	-483	⁸⁷ Sr/ ⁸⁶ Sr	504 $\pm$ 280	-0.76 $\pm$ 0.59	Webster et al. (2010)
MVD9-R4	<i>Porites compressa</i> ?	20.61	-156.91	-483	⁸⁷ Sr/ ⁸⁶ Sr	912 $\pm$ 190	-0.4 $\pm$ 0.16	Webster et al. (2010)
T312-R3	Coralline algae	20.64	-156.71	-286	¹⁴ C	31.05 $\pm$ 0.13 ^b	-6.59 $\pm$ 1	Webster et al. (2006)
	Slope break			-1600	K-Ar	1030 $\pm$ 360	-1.48 $\pm$ 0.55	Clague and Dalrymple (1987)
					<i>Lana'i</i>			
T295-R25	Matrix sediments	20.68	-157	-487	⁸⁷ Sr/ ⁸⁶ Sr	270 $\pm$ 330		Webster et al. (2010)
T295-R20A	Coralline algae	20.68	-157	-525	⁸⁷ Sr/ ⁸⁶ Sr	836 $\pm$ 240	-0.52 $\pm$ 0.24	Webster et al. (2010)
T295-R20B	<i>Leptastrea</i> ?	20.68	-157	-525	⁸⁷ Sr/ ⁸⁶ Sr	676 $\pm$ 240	-0.6 $\pm$ 0.32	Webster et al. (2010)
T295-R21	Coralline algae	20.68	-157	-525	⁸⁷ Sr/ ⁸⁶ Sr	555 $\pm$ 270	-0.74 $\pm$ 0.51	Webster et al. (2010)
T295-R12	<i>Porites compressa</i> ?	20.68	-157	-533.7	⁸⁷ Sr/ ⁸⁶ Sr	-375 $\pm$ 540		Webster et al. (2010)
T295-R6	<i>Porites compressa</i> ?	20.68	-157	-533.9	⁸⁷ Sr/ ⁸⁶ Sr	580 $\pm$ 300	-0.79 $\pm$ 0.55	Webster et al. (2010)
T295-R7	<i>Porites compressa</i> ?	20.68	-157	-534	⁸⁷ Sr/ ⁸⁶ Sr	295 $\pm$ 330		Webster et al. (2010)
T295-R10	<i>Porites compressa</i> ?	20.68	-157	-534	⁸⁷ Sr/ ⁸⁶ Sr	984 $\pm$ 210	-0.45 $\pm$ 0.17	Webster et al. (2010)
T295-R8	<i>Porites compressa</i> ?	20.68	-157	-534	⁸⁷ Sr/ ⁸⁶ Sr	359 $\pm$ 260	-1.19 $\pm$ 1.12	Webster et al. (2010)
T295-R13	<i>Porites compressa</i> ?	20.68	-157	-534	⁸⁷ Sr/ ⁸⁶ Sr	996 $\pm$ 200	-0.44 $\pm$ 0.16	Webster et al. (2010)
T294-R21#1	Echinoid spine	20.67	-157.03	-564	⁸⁷ Sr/ ⁸⁶ Sr	676 $\pm$ 310	-0.66 $\pm$ 0.42	Webster et al. (2010)
T294-R21#2	Echinoid spine	20.67	-157.03	-564	⁸⁷ Sr/ ⁸⁶ Sr	609 $\pm$ 250	-0.78 $\pm$ 0.45	Webster et al. (2010)
T294-R7	Matrix sediments	20.67	-157.03	-584	⁸⁷ Sr/ ⁸⁶ Sr	748 $\pm$ 250	-0.62 $\pm$ 0.31	Webster et al. (2010)
T294-R7#2	Echinoid spine	20.67	-157.03	-584	⁸⁷ Sr/ ⁸⁶ Sr	912 $\pm$ 190	-0.51 $\pm$ 0.19	Webster et al. (2010)
T294-R8A	Coralline algae	20.67	-157.03	-584	⁸⁷ Sr/ ⁸⁶ Sr	676 $\pm$ 210	-0.69 $\pm$ 0.32	Webster et al. (2010)
T294-R8S	Echinoid spine	20.67	-157.03	-584	⁸⁷ Sr/ ⁸⁶ Sr	887 $\pm$ 220	-0.52 $\pm$ 0.21	Webster et al. (2010)
T294-R8S#2	Echinoid spine	20.67	-157.03	-584	⁸⁷ Sr/ ⁸⁶ Sr	676 $\pm$ 270	-0.69 $\pm$ 0.39	Webster et al. (2010)
T309-R34	Massive <i>Porites</i>	20.68	-157.14	-591.8	⁸⁷ Sr/ ⁸⁶ Sr	748 $\pm$ 220	-0.63 $\pm$ 0.29	Webster et al. (2010)

Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
T309-R31	<i>Porites compressa</i> ?	20.68	−157.14	−599.2	⁸⁷ Sr/ ⁸⁶ Sr	270 $\pm$ 380		Webster et al. (2010)
T309-R31	Matrix sediments	20.68	−157.14	−599	⁸⁷ Sr/ ⁸⁶ Sr	609 $\pm$ 290	−0.84 $\pm$ 0.53	Webster et al. (2010)
T309-R25	Matrix sediments	20.68	−157.14	−601	⁸⁷ Sr/ ⁸⁶ Sr	941 $\pm$ 210	−0.57 $\pm$ 0.2	Webster et al. (2010)
T309-R28	<i>Porites compressa</i> ?	20.68	−157.14	−601	⁸⁷ Sr/ ⁸⁶ Sr	984 $\pm$ 210	−0.52 $\pm$ 0.18	Webster et al. (2010)
T309-R24	<i>Pocillopora</i>	20.68	−157.14	−647	⁸⁷ Sr/ ⁸⁶ Sr	1232 $\pm$ 200	−0.47 $\pm$ 0.13	Webster et al. (2010)
T309-R24#2	Echinoid spine	20.68	−157.14	−647	⁸⁷ Sr/ ⁸⁶ Sr	1123 $\pm$ 180	−0.47 $\pm$ 0.13	Webster et al. (2010)
T309-R21	Matrix sediments	20.68	−157.14	−708.9	⁸⁷ Sr/ ⁸⁶ Sr	771 $\pm$ 250	−0.8 $\pm$ 0.36	Webster et al. (2010)
T309-R21	Coralline algae	20.68	−157.14	−709	⁸⁷ Sr/ ⁸⁶ Sr	793 $\pm$ 220	−0.72 $\pm$ 0.3	Webster et al. (2010)
T309-R18	Coralline algae	20.68	−157.14	−712	⁸⁷ Sr/ ⁸⁶ Sr	811 $\pm$ 240	−0.76 $\pm$ 0.32	Webster et al. (2010)
T309-R13	Branching <i>Porites</i>	20.68	−157.14	−764	⁸⁷ Sr/ ⁸⁶ Sr	912 $\pm$ 220	−0.71 $\pm$ 0.25	Webster et al. (2010)
T309-R15	<i>Leptastrea</i> ?	20.68	−157.14	−764	⁸⁷ Sr/ ⁸⁶ Sr	1063 $\pm$ 220	−0.65 $\pm$ 0.2	Webster et al. (2010)
T309-R8	<i>Porites compressa</i> ?	20.68	−157.14	−860	⁸⁷ Sr/ ⁸⁶ Sr	705 $\pm$ 270	−1.05 $\pm$ 0.51	Webster et al. (2010)
T309-R7	Matrix sediments	20.68	−157.14	−889.1	⁸⁷ Sr/ ⁸⁶ Sr	504 $\pm$ 280	−1.56 $\pm$ 1.04	Webster et al. (2010)
T309-R7	<i>Porites compressa</i> ?	20.68	−157.14	−889	⁸⁷ Sr/ ⁸⁶ Sr	1467 $\pm$ 200	−0.55 $\pm$ 0.12	Webster et al. (2010)
T309-R7	<i>Porites compressa</i> ?	20.68	−157.14	−889	⁸⁷ Sr/ ⁸⁶ Sr	836 $\pm$ 240	−0.96 $\pm$ 0.37	Webster et al. (2010)
T309-R6	<i>Porites compressa</i> ?	20.68	−157.14	−890.3	⁸⁷ Sr/ ⁸⁶ Sr	793 $\pm$ 220	−0.95 $\pm$ 0.36	Webster et al. (2010)
T309-R6	<i>Porites compressa</i> ?	20.68	−157.14	−890	⁸⁷ Sr/ ⁸⁶ Sr	912 $\pm$ 220	−0.85 $\pm$ 0.28	Webster et al. (2010)
T309-R4	Matrix sediments	20.68	−157.14	−896.8	⁸⁷ Sr/ ⁸⁶ Sr	1080 $\pm$ 140	−0.77 $\pm$ 0.16	Webster et al. (2010)
T309-R5	<i>Porites</i>	20.68	−157.14	−896.5	⁸⁷ Sr/ ⁸⁶ Sr	1029 $\pm$ 140	−0.76 $\pm$ 0.17	Webster et al. (2010)
P191-7	Massive <i>Porites</i>	20.89	−157.18	−845	⁸⁷ Sr/ ⁸⁶ Sr	1063 $\pm$ 190	−0.72 $\pm$ 0.2	Webster et al. (2010)
P191-7B	<i>Porites compressa</i> ?	20.89	−157.18	−845	⁸⁷ Sr/ ⁸⁶ Sr	234 $\pm$ 390		Webster et al. (2010)
P191-4	<i>Acropora</i>	20.87	−157.2	−880	⁸⁷ Sr/ ⁸⁶ Sr	1080 $\pm$ 160	−0.75 $\pm$ 0.17	Webster et al. (2010)
P191-3B	<i>Acropora</i>	20.87	−157.2	−925	⁸⁷ Sr/ ⁸⁶ Sr	1141 $\pm$ 180	−0.72 $\pm$ 0.17	Webster et al. (2010)
T310-R21	Oolite	20.71	−156.9	−142	¹⁴ C	17.87 $\pm$ 0.56 ^b	0.1 $\pm$ 1.87	Webster et al. (2006)
T310-R11	Coralline algae	20.71	−156.9	−150	¹⁴ C	29.76 $\pm$ 0.13 ^b	−1.97 $\pm$ 1.07	Webster et al. (2006)
T310-R19	Benthic foraminifera	20.71	−156.9	−150	¹⁴ C	12.2 $\pm$ 0.33 ^b	−4.58 $\pm$ 3.33	Webster et al. (2006)
T310-R13	<i>Porites</i>	20.71	−156.9	−153	¹⁴ C	33.34 $\pm$ 0.21 ^b	−1.28 $\pm$ 1.1	Webster et al. (2006)
T310-R1	Coralline algae	20.71	−156.9	−154	¹⁴ C	29.98 $\pm$ 0.12 ^b	−2.16 $\pm$ 1.06	Webster et al. (2006)
T311-R22	Coralline algae	20.69	−156.71	−163	¹⁴ C	22.29 $\pm$ 0.78 ^b	−1.3 $\pm$ 1.73	Webster et al. (2006)
T311-R23	<i>Leptoseris hawaiiensis</i>	20.69	−156.71	−166	¹⁴ C	21.36 $\pm$ 0.72 ^b	−1.22 $\pm$ 1.56	Webster et al. (2006)
T311-R24A	Coralgal nodule limestone	20.69	−156.71	−166	¹⁴ C	13.49 $\pm$ 0.33 ^b	−3.74 $\pm$ 2.47	Webster et al. (2006)
T311-R24B	Coralline algae	20.69	−156.71	−166	¹⁴ C	16.7 $\pm$ 0.48 ^b	−2.06 $\pm$ 2.39	Webster et al. (2006)
T311-R21	<i>Halimeda distorta</i>	20.68	−156.71	−167	¹⁴ C	14.13 $\pm$ 0.31 ^b	−3.4 $\pm$ 2.33	Webster et al. (2006)
T311-R20	<i>Halimeda distorta</i>	20.68	−156.71	−168	¹⁴ C	12.97 $\pm$ 0.18 ^b	−4.3 $\pm$ 2.65	Webster et al. (2006)
T311-R4	Coralline algae	20.68	−156.72	−230	¹⁴ C	16.71 $\pm$ 0.48 ^b	−5.88 $\pm$ 2.49	Webster et al. (2006)
T311-R8A	Coralline algae	20.68	−156.72	−231	¹⁴ C	29.92 $\pm$ 0.29 ^b	−4.72 $\pm$ 1.16	Webster et al. (2006)
T311-R8B	Coralline algae	20.68	−156.72	−231	¹⁴ C	17.54 $\pm$ 0.53 ^b	−5.06 $\pm$ 2.12	Webster et al. (2006)
	Aragonitic coral	20.77	−156.85	115	U–Th	108 $\pm$ 5	1.27 $\pm$ 0.15	B. Szabo cited in Moore and Moore (1988)
	Aragonitic coral	20.75	−156.85	120	U–Th	101 $\pm$ 4	1.41 $\pm$ 0.16	B. Szabo cited in Moore and Moore (1988)
	Aragonitic coral	20.75	−156.87	155	U–Th	134 $\pm$ 7	1.88 $\pm$ 0.62	B. Szabo cited in Moore and Moore (1988)
L-10	Limestone conglomerate w/ coral fragments	20.74	−156.9 ^d	12	ESR	210 $\pm$ 36	0.25 $\pm$ 0.25	Jones (1993a)
L-21S	Hulopo'e gravel	20.74	−156.9 ^d	16	ESR	185 $\pm$ 40	0.4 $\pm$ 0.39	Jones (1993a)
L-21	Hulopo'e gravel	20.74	−156.9 ^d	16	ESR	202 $\pm$ 15	0.12 $\pm$ 0.19	Jones (1993a)
L-41B		20.81	−156.98 ^d	23	ESR	200 $\pm$ 23	0.09 $\pm$ 0.2	Jones (1993a)
L-41A		20.81	−156.98 ^d	23	ESR	223 $\pm$ 18	0.31 $\pm$ 0.19	Jones (1993a)
L-43AX		20.81	−156.98 ^d	27	ESR	205 $\pm$ 31	0.3 $\pm$ 0.25	Jones (1993a)
L-42AR		20.81	−156.98 ^d	27	ESR	197 $\pm$ 16	0.12 $\pm$ 0.23	Jones (1993a)

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Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
L-42AS	Platygyra, growth position	20.81	−156.98 ^d	27	ESR	215 $\pm$ 15	0.23 $\pm$ 0.16	Jones (1993a)
L-42B		20.81	−156.98 ^d	27	ESR	224 $\pm$ 39	0.35 $\pm$ 0.32	Jones (1993a)
L-95		20.74	−156.9 ^d	29	ESR	201 $\pm$ 17	0.15 $\pm$ 0.21	Jones (1993a)
L-95+		20.74	−156.9 ^d	29	ESR	173 $\pm$ 15	0.38 $\pm$ 0.29	Jones (1993a)
L-95		20.74	−156.9 ^d	29	ESR	181 $\pm$ 49	0.39 $\pm$ 0.46	Jones (1993a)
L-47	Coral clast in marine mudstone	20.83	−156.99 ^d	40	ESR	37 $\pm$ 11	3.09 $\pm$ 1.57	Jones (1993a)
L-t.c.		20.74	−156.9 ^d	44	ESR	140 $\pm$ 14	1.2 $\pm$ 0.53	Jones (1993a)
L-23B		20.74	−156.9 ^d	47	ESR	235 $\pm$ 33	0.47 $\pm$ 0.3	Jones (1993a)
L-23A		20.74	−156.9 ^d	47	ESR	233 $\pm$ 21	0.5 $\pm$ 0.31	Jones (1993a)
L-26		20.76	−156.85 ^d	52	ESR	217 $\pm$ 19	0.32 $\pm$ 0.22	Jones (1993a)
L-24S	Limestone conglomerate w/ coral fragments	20.75	−156.9 ^d	74	ESR	195 $\pm$ 38	0.38 $\pm$ 0.34	Jones (1993a)
L-24R		20.75	−156.9 ^d	74	ESR	223 $\pm$ 22	0.54 $\pm$ 0.23	Jones (1993a)
L-32A		20.75	−156.88 ^d	170	ESR	137 $\pm$ 14	2.02 $\pm$ 0.74	Jones (1993a)
L-32B		20.75	−156.88 ^d	170	ESR	112 $\pm$ 12	1.79 $\pm$ 0.37	Jones (1993a)
Lan3-3-1		20.74	−156.9 ^d	3.8	U–Th	134.6 $\pm$ 0.9	0.85 $\pm$ 0.16	Rubin et al. (2000)
Lan1-0-1	Littoral and alluvial deposits	20.74	−156.9 ^d	8	U–Th	196.2 $\pm$ 1.3	0.03 $\pm$ 0.02	Rubin et al. (2000)
Lan2-8-1		20.74	−156.9 ^d	35	U–Th	211.4 $\pm$ 1.5	0.35 $\pm$ 0.02	Rubin et al. (2000)
Lan1-2h-1		20.74	−156.9 ^d	20	U–Th	130.9 $\pm$ 1.5	0.54 $\pm$ 0.1	Rubin et al. (2000)
Lan3-5-3		20.74	−156.9 ^d	23	U–Th	136.4 $\pm$ 0.9	1.03 $\pm$ 0.11	Rubin et al. (2000)
Lan2-4-3		20.74	−156.9 ^d	19	U–Th	136.7 $\pm$ 1	0.96 $\pm$ 0.12	Rubin et al. (2000)
Lan1-4-1	Littoral and alluvial deposits	20.74	−156.9 ^d	28	U–Th	230.2 $\pm$ 3.9	0.35 $\pm$ 0.07	Rubin et al. (2000)
Lan3-5-1		20.74	−156.9 ^d	17.7	U–Th	204.1 $\pm$ 1.4	0.24 $\pm$ 0.06	Rubin et al. (2000)
Lan1-6-1		20.75	−156.9 ^d	58	U–Th	222.4 $\pm$ 1.5	0.45 $\pm$ 0.07	Rubin et al. (2000)
Lan2-6-2		20.75	−156.9 ^d	58.9	U–Th	235.9 $\pm$ 1.5	0.49 $\pm$ 0.09	Rubin et al. (2000)
	Slope break			−1250	K–Ar	1280 $\pm$ 80	−0.94 $\pm$ 0.08	Clague and Dalrymple (1987)
T300-R29	<i>Acropora cerealis?</i>	21.31	−157.02	−1284.1	East Moloka'i ⁸⁷ Sr/ ⁸⁶ Sr	1371 $\pm$ 165	−0.86 $\pm$ 0.15	Faichney et al. (2010)
T300-R30	<i>Porites</i>	21.31	−157.02	−1287.9		1992 $\pm$ 383	−0.62 $\pm$ 0.15	Faichney et al. (2010)
T300-R31	<i>Porites</i>	21.31	−157.02	−1281.4		1449 $\pm$ 194	−0.85 $\pm$ 0.16	Faichney et al. (2010)
T300-R35	<i>Acropora</i>	21.31	−157.02	−1279		1232 $\pm$ 194	−0.99 $\pm$ 0.21	Faichney et al. (2010)
91WA-1BA	<i>Porites compressa?</i>	21.02	−157.12	−342.5		912 $\pm$ 220	−0.25 $\pm$ 0.14	Webster et al. (2010)

Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
91WA-1BB	Coralline algae	21.02	–157.12	–342.5	⁸⁷ Sr/ ⁸⁶ Sr	504 $\pm$ 280	–0.48 $\pm$ 0.44	Webster et al. (2010)
91WA-1G	Echinoid spine	21.02	–157.12	–342.5	⁸⁷ Sr/ ⁸⁶ Sr	633 $\pm$ 250	–0.3 $\pm$ 0.24	Webster et al. (2010)
91WA-1G#2	Echinoid spine	21.02	–157.12	–342.5	⁸⁷ Sr/ ⁸⁶ Sr	811 $\pm$ 180	–0.3 $\pm$ 0.16	Webster et al. (2010)
T308-R2	Matrix sediments	21.04	–157.19	–604.9	⁸⁷ Sr/ ⁸⁶ Sr	811 $\pm$ 240	–0.63 $\pm$ 0.28	Webster et al. (2010)
Mkk-41		21.1	–157.03 ^d	13	ESR	163 $\pm$ 14	0.64 $\pm$ 0.31	Jones (1993a)
Mkk-31		21.09	–157.02 ^d	14	ESR	258 $\pm$ 17	0.29 $\pm$ 0.22	Jones (1993a)
Mkk-31a		21.09	–157.02 ^d	14	ESR	237 $\pm$ 40	0.24 $\pm$ 0.3	Jones (1993a)
Mkk-31b		21.09	–157.02 ^d	14	ESR	272 $\pm$ 32	0.35 $\pm$ 0.21	Jones (1993a)
Mkk-31c		21.09	–157.02 ^d	14	ESR	245 $\pm$ 21	0.29 $\pm$ 0.25	Jones (1993a)
Mkk-tw		21.09	–157.02 ^d	14	ESR	253 $\pm$ 12	0.37 $\pm$ 0.25	Jones (1993a)
Mkk-32a		21.1	–157.04 ^d	15	ESR	232 $\pm$ 18	0.34 $\pm$ 0.28	Jones (1993a)
Mkk-32b		21.1	–157.04 ^d	15	ESR	241 $\pm$ 18	0.03 $\pm$ 0.24	Jones (1993a)
Mkk-42l		21.1	–157.02 ^d	18	ESR	177 $\pm$ 22	0.27 $\pm$ 0.37	Jones (1993a)
Mkk-33b		21.1	–157.04 ^d	30	ESR	290 $\pm$ 31	0.26 $\pm$ 0.16	Jones (1993a)
Mkk-20		21.12	–157.06 ^d	58	ESR	303 $\pm$ 25	0.42 $\pm$ 0.15	Jones (1993a)
	Slope break			–1850	K-Ar	1760 $\pm$ 140	–1.03 $\pm$ 0.11	Clague and Dalrymple (1987)
					West Moloka'i			
	Slope break			–1050	K-Ar	1900 $\pm$ 120	–0.55 $\pm$ 0.05	Clague and Dalrymple (1987)
					Ko'olau			
LJ-570	Limpets	21.71	–157.98	1.5	¹⁴ C	8.14 $\pm$ 0.62 ^b	3.01 $\pm$ 1.45	Hubbs et al. (1965)
LJ-948	Gastropod	21.4	–157.72	1.5	¹⁴ C	43.17 $\pm$ 2.25 ^b	2.02 $\pm$ 0.44	Hubbs et al. (1965)
1	Marine shells	21.71	–157.98 ^d	2	¹⁴ C	23.14 $\pm$ 3.64 ^b	4.23 $\pm$ 1.76	Stearns (1974)
2	Marine shells	21.71	–157.98 ^d	–1	¹⁴ C	25.43 $\pm$ 0.37 ^b	3.21 $\pm$ 0.1	Stearns (1974)
3	Shells and coral	21.32	–157.67 ^d	5	¹⁴ C	27.27 $\pm$ 2.7 ^b	2.89 $\pm$ 0.88	Stearns (1974)
C1	Porites	21.26	–157.79 ^d	7.6	²³¹ Pa/ ²³⁵ U	105 $\pm$ 30	0.25 $\pm$ 0.45	Ku et al. (1974)
C1	Porites	21.26	–157.79 ^d	7.6	U–Th	115 $\pm$ 12	0.28 $\pm$ 0.21	Ku et al. (1974)
C1	Porites	21.26	–157.79 ^d	7.6	U–Th	118 $\pm$ 14	0.08 $\pm$ 0.26	Ku et al. (1974)
C1	Porites	21.26	–157.79 ^d	7.6	²³¹ Pa/ ²³⁵ U	120 $\pm$ 22	0.12 $\pm$ 0.66	Ku et al. (1974)
C10	Porites	21.45	–157.73 ^d	7.8	U–Th	131 $\pm$ 16	0.45 $\pm$ 0.73	Ku et al. (1974)
C11	Pocillopora	21.45	–157.73 ^d	11	U–Th	134 $\pm$ 14	0.81 $\pm$ 0.71	Ku et al. (1974)
C12	Pocillopora	21.71	–157.97 ^d	0.5	U–Th	137 $\pm$ 22	0.78 $\pm$ 0.7	Ku et al. (1974)
C13	Pocillopora	21.45	–157.74 ^d	1.5	U–Th	118 $\pm$ 18	0.03 $\pm$ 0.48	Ku et al. (1974)
C13	Pocillopora	21.45	–157.74 ^d	1.5	²³¹ Pa/ ²³⁵ U	105 $\pm$ 30	0.19 $\pm$ 0.44	Ku et al. (1974)
C14	Porites	21.26	–157.8 ^d	1.8	U–Th	137 $\pm$ 22	0.79 $\pm$ 0.7	Ku et al. (1974)
C15	Porites	21.26	–157.8 ^d	2.2	U–Th	124 $\pm$ 18	0.04 $\pm$ 0.67	Ku et al. (1974)
C16	Porites	21.26	–157.81 ^d	2.5	U–Th	121 $\pm$ 8	0.09 $\pm$ 0.2	Ku et al. (1974)
C16	Porites	21.26	–157.81 ^d	2.5	²³¹ Pa/ ²³⁵ U	96 $\pm$ 28	0.43 $\pm$ 0.45	Ku et al. (1974)
C16	Porites	21.26	–157.81 ^d	2.5	U–Th	121 $\pm$ 8	0.09 $\pm$ 0.2	Ku et al. (1974)
C16	Porites	21.26	–157.81 ^d	2.5	²³¹ Pa/ ²³⁵ U	91 $\pm$ 20	0.36 $\pm$ 0.32	Ku et al. (1974)
C17	Porites	21.26	–157.81 ^d	0.1	U–Th	128 $\pm$ 16	0.24 $\pm$ 0.71	Ku et al. (1974)
C17	Porites	21.26	–157.81 ^d	0.1	²³¹ Pa/ ²³⁵ U	109 $\pm$ 30	0.23 $\pm$ 0.58	Ku et al. (1974)
C18	Porites	21.71	–157.97 ^d	1	U–Th	115 $\pm$ 12	0.22 $\pm$ 0.2	Ku et al. (1974)
C2	Porites	21.26	–157.79 ^d	7.6	U–Th	118 $\pm$ 6	0.08 $\pm$ 0.2	Ku et al. (1974)
C3	Porites	21.26	–157.79 ^d	7.6	²³¹ Pa/ ²³⁵ U	105 $\pm$ 38	0.25 $\pm$ 0.69	Ku et al. (1974)
C3	Porites	21.26	–157.79 ^d	7.6	²³¹ Pa/ ²³⁵ U	138 $\pm$ 160		Ku et al. (1974)
C3	Porites	21.26	–157.79 ^d	7.6	U–Th	112 $\pm$ 12	0.34 $\pm$ 0.22	Ku et al. (1974)
C3	Porites	21.26	–157.79 ^d	7.6	U–Th	121 $\pm$ 14	0.13 $\pm$ 0.45	Ku et al. (1974)
C4	Porites	21.26	–157.79 ^d	6.7	²³¹ Pa/ ²³⁵ U	94 $\pm$ 22	0.52 $\pm$ 0.34	Ku et al. (1974)
C4	Porites	21.26	–157.79 ^d	6.7	²³¹ Pa/ ²³⁵ U	105 $\pm$ 38	0.24 $\pm$ 0.69	Ku et al. (1974)
C4	Porites	21.26	–157.79 ^d	6.7	U–Th	115 $\pm$ 12	0.27 $\pm$ 0.21	Ku et al. (1974)
C4	Porites	21.26	–157.79 ^d	6.7	U–Th	118 $\pm$ 14	0.07 $\pm$ 0.26	Ku et al. (1974)
C5	Porites	21.26	–157.79 ^d	7.6	²³¹ Pa/ ²³⁵ U	130 $\pm$ 84	0.4 $\pm$ 0.73	Ku et al. (1974)
C5	Porites	21.26	–157.79 ^d	7.6	U–Th	128 $\pm$ 16	0.3 $\pm$ 0.72	Ku et al. (1974)
C6	Porites	21.28	–157.74 ^d	8.1	U–Th	121 $\pm$ 14	0.14 $\pm$ 0.45	Ku et al. (1974)

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Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
C6	<i>Porites</i>	21.28	-157.74 ^d	8.1	²³¹ Pu/ ²³⁵ U	120 $\pm$ 36	0.13 $\pm$ 0.61	Ku et al. (1974)
C7	<i>Pocillopora</i>	21.28	-157.74 ^d	6.1	U–Th	121 $\pm$ 8	0.12 $\pm$ 0.2	Ku et al. (1974)
C8	<i>Porites</i>	21.28	-157.76 ^d	7.5	U–Th	121 $\pm$ 14	0.13 $\pm$ 0.45	Ku et al. (1974)
C8	<i>Porites</i>	21.28	-157.76 ^d	7.5	U–Th	118 $\pm$ 6	0.08 $\pm$ 0.2	Ku et al. (1974)
C8	<i>Porites</i>	21.28	-157.76 ^d	7.5	²³¹ Pu/ ²³⁵ U	113 $\pm$ 50	0.34 $\pm$ 0.74	Ku et al. (1974)
C8	<i>Porites</i>	21.28	-157.76 ^d	7.5	²³¹ Pu/ ²³⁵ U	124 $\pm$ 64	0.08 $\pm$ 0.54	Ku et al. (1974)
11	Algae	21.27	-157.7 ^d	0.2	¹⁴ C	2.17 $\pm$ 0.5 ^b	-1.94 $\pm$ 1.38	Easton and Olson (1976)
12A	<i>Cyphastrea</i>	21.27	-157.7 ^d	-0.6	¹⁴ C	2.75 $\pm$ 0.19 ^b	-1.23 $\pm$ 0.5	Easton and Olson (1976)
13	Algae	21.27	-157.7 ^d	-1.2	¹⁴ C	4.11 $\pm$ 0.35 ^b	-0.19 $\pm$ 0.35	Easton and Olson (1976)
I-1618	<i>Porites</i>	21.27	-157.7 ^d	-1.2	¹⁴ C	3.68 $\pm$ 0.29 ^b	-0.46 $\pm$ 0.37	Easton and Olson (1976)
13-B	Algae	21.27	-157.7 ^d	-1.2	¹⁴ C	3.15 $\pm$ 0.52 ^b	-0.91 $\pm$ 0.8	Easton and Olson (1976)
14	<i>Cyphastrea</i>	21.27	-157.7 ^d	-3.7	¹⁴ C	4.5 $\pm$ 0.28 ^b	-0.53 $\pm$ 0.25	Easton and Olson (1976)
I-1619	<i>Porites</i>	21.27	-157.7 ^d	-4.6	¹⁴ C	4.46 $\pm$ 0.4 ^b	-0.76 $\pm$ 0.33	Easton and Olson (1976)
15	<i>Porites</i>	21.27	-157.7 ^d	-6.5	¹⁴ C	5.76 $\pm$ 0.23 ^b	-1.15 $\pm$ 0.2	Easton and Olson (1976)
17	<i>Pocillopora</i>	21.27	-157.7 ^d	-7.6	¹⁴ C	6.07 $\pm$ 0.25 ^b	-1.23 $\pm$ 0.47	Easton and Olson (1976)
18	<i>Pocillopora</i>	21.27	-157.7 ^d	-8.2	¹⁴ C	5.94 $\pm$ 0.29 ^b	-1.46 $\pm$ 0.36	Easton and Olson (1976)
33	<i>Porites</i>	21.27	-157.7 ^d	-0.3	¹⁴ C	1.41 $\pm$ 0.15 ^b	-3.19 $\pm$ 1.11	Easton and Olson (1976)
33	<i>Pocillopora</i>	21.27	-157.7 ^d	-2.1	¹⁴ C	0.9 $\pm$ 0.18 ^b	-5.15 $\pm$ 2.33	Easton and Olson (1976)
53	Beachrock	21.27	-157.7 ^d	0.2	¹⁴ C	0.9 $\pm$ 0.35 ^b	-2.5 $\pm$ 3.03	Easton and Olson (1976)
54	<i>Porites</i>	21.27	-157.7 ^d	0.2	¹⁴ C	1.46 $\pm$ 0.24 ^b	-2.82 $\pm$ 1.44	Easton and Olson (1976)
I-1509	<i>Porites</i>	21.27	-157.7 ^d	-1.7	¹⁴ C	3.84 $\pm$ 0.39 ^b	-0.47 $\pm$ 0.43	Easton and Olson (1976)
56	<i>Porites</i>	21.27	-157.7 ^d	-1.7	¹⁴ C	3.54 $\pm$ 0.19 ^b	-0.69 $\pm$ 0.32	Easton and Olson (1976)
61-B	Algae	21.27	-157.7 ^d	-0.5	¹⁴ C	3.1 $\pm$ 0.26 ^b	-0.75 $\pm$ 0.48	Easton and Olson (1976)
61	Algae	21.27	-157.7 ^d	-1.5	¹⁴ C	2.27 $\pm$ 0.24 ^b	-2.47 $\pm$ 0.82	Easton and Olson (1976)
62	Algae	21.27	-157.7 ^d	-1.5	¹⁴ C	3.3 $\pm$ 0.26 ^b	-0.83 $\pm$ 0.43	Easton and Olson (1976)
63	Algae	21.27	-157.7 ^d	-3.3	¹⁴ C	4.07 $\pm$ 0.29 ^b	-0.72 $\pm$ 0.35	Easton and Olson (1976)
64	<i>Porites</i>	21.27	-157.7 ^d	-4.2	¹⁴ C	5.21 $\pm$ 0.3 ^b	-0.69 $\pm$ 0.23	Easton and Olson (1976)
65	<i>Porites</i>	21.27	-157.7 ^d	-8.2	¹⁴ C	6.07 $\pm$ 0.25 ^b	-1.32 $\pm$ 0.47	Easton and Olson (1976)
71	Algae	21.27	-157.7 ^d	-1.1	¹⁴ C	2.99 $\pm$ 0.27 ^b	-1.06 $\pm$ 0.57	Easton and Olson (1976)
71-A	Algae	21.27	-157.7 ^d	-1.1	¹⁴ C	2.76 $\pm$ 0.22 ^b	-1.39 $\pm$ 0.56	Easton and Olson (1976)
72	Algae	21.27	-157.7 ^d	-0.2	¹⁴ C	0.54 $\pm$ 0.15 ^b	-2.69 $\pm$ 2.72	Easton and Olson (1976)
73	<i>Pocillopora</i>	21.27	-157.7 ^d	-0.8	¹⁴ C	3.93 $\pm$ 0.33 ^b	-0.18 $\pm$ 0.35	Easton and Olson (1976)
74	Algae	21.27	-157.7 ^d	-1.7	¹⁴ C	5.02 $\pm$ 0.31 ^b	-0.15 $\pm$ 0.21	Easton and Olson (1976)
75	Algae	21.27	-157.7 ^d	-3.2	¹⁴ C	5.28 $\pm$ 0.36 ^b	-0.51 $\pm$ 0.24	Easton and Olson (1976)
77	Algae	21.27	-157.7 ^d	-4.1	¹⁴ C	5.42 $\pm$ 0.5 ^b	-0.7 $\pm$ 0.31	Easton and Olson (1976)
78	Algae	21.27	-157.7 ^d	-4.7	¹⁴ C	5.65 $\pm$ 0.3 ^b	-0.84 $\pm$ 0.22	Easton and Olson (1976)
82	Algae	21.27	-157.7 ^d	-1.4	¹⁴ C	1.72 $\pm$ 0.24 ^b	-3.84 $\pm$ 1.18	Easton and Olson (1976)
82	Algae	21.27	-157.7 ^d	-1.4	¹⁴ C	2 $\pm$ 0.26 ^b	-3.16 $\pm$ 1.06	Easton and Olson (1976)
82-B	Algae	21.27	-157.7 ^d	-3.2	¹⁴ C	4.26 $\pm$ 0.38 ^b	-0.57 $\pm$ 0.39	Easton and Olson (1976)
84	<i>Cyphastrea</i>	21.27	-157.7 ^d	-4.4	¹⁴ C	4.35 $\pm$ 0.37 ^b	-0.79 $\pm$ 0.37	Easton and Olson (1976)
85-A	<i>Pocillopora</i>	21.27	-157.7 ^d	-7.9	¹⁴ C	7.6 $\pm$ 0.23 ^b	1.08 $\pm$ 0.54	Easton and Olson (1976)
85-B	<i>Porites</i>	21.27	-157.7 ^d	-9.5	¹⁴ C	7.56 $\pm$ 0.18 ^b	0.8 $\pm$ 0.44	Easton and Olson (1976)
86	<i>Porites</i>	21.27	-157.7 ^d	-9.8	¹⁴ C	7.51 $\pm$ 0.33 ^b	0.69 $\pm$ 0.73	Easton and Olson (1976)
90	Algae	21.27	-157.7 ^d	-1.1	¹⁴ C	2.47 $\pm$ 0.53 ^b	-1.89 $\pm$ 1.33	Easton and Olson (1976)
90	Algae	21.27	-157.7 ^d	-1.1	¹⁴ C	3.1 $\pm$ 0.28 ^b	-0.92 $\pm$ 0.53	Easton and Olson (1976)
91	Algae	21.27	-157.7 ^d	-1.7	¹⁴ C	2.89 $\pm$ 0.35 ^b	-1.41 $\pm$ 0.75	Easton and Olson (1976)
91	Algae	21.27	-157.7 ^d	-1.7	¹⁴ C	2.83 $\pm$ 0.27 ^b	-1.5 $\pm$ 0.65	Easton and Olson (1976)
92	<i>Cyphastrea</i>	21.27	-157.7 ^d	-7.2	¹⁴ C	5.4 $\pm$ 0.3 ^b	-1.26 $\pm$ 0.25	Easton and Olson (1976)
92	<i>Cyphastrea</i>	21.27	-157.7 ^d	-7.2	¹⁴ C	5.21 $\pm$ 0.27 ^b	-1.25 $\pm$ 0.25	Easton and Olson (1976)
93	<i>Porites</i>	21.27	-157.7 ^d	-8.1	¹⁴ C	6.09 $\pm$ 0.24 ^b	-1.25 $\pm$ 0.5	Easton and Olson (1976)
93	<i>Porites</i>	21.27	-157.7 ^d	-8.1	¹⁴ C	6.38 $\pm$ 0.36 ^b	-0.76 $\pm$ 0.78	Easton and Olson (1976)
93	<i>Porites</i>	21.27	-157.7 ^d	-8.1	¹⁴ C	6.61 $\pm$ 0.28 ^b	-0.37 $\pm$ 0.59	Easton and Olson (1976)
93	<i>Porites</i>	21.27	-157.7 ^d	-8.1	¹⁴ C	6.6 $\pm$ 0.33 ^b	-0.38 $\pm$ 0.68	Easton and Olson (1976)

Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
93	Porites	21.27	-157.7 ^d	-8.1	¹⁴ C	6.08 $\pm$ 0.47 ^b	-1.29 $\pm$ 0.75	Easton and Olson (1976)
94	Porites	21.27	-157.7 ^d	-8.7	¹⁴ C	6.64 $\pm$ 0.31 ^b	-0.43 $\pm$ 0.64	Easton and Olson (1976)
97	Sand	21.27	-157.7 ^d	-10.5	¹⁴ C	6.04 $\pm$ 0.2 ^b	-1.76 $\pm$ 0.4	Easton and Olson (1976)
101	Algae	21.27	-157.7 ^d	-0.3	¹⁴ C	0.29 $\pm$ 0.23 ^b	-2.66 $\pm$ 6.95	Easton and Olson (1976)
102	Algae	21.27	-157.7 ^d	-3	¹⁴ C	2.59 $\pm$ 0.18 ^b	-2.43 $\pm$ 0.59	Easton and Olson (1976)
104	Algae	21.27	-157.7 ^d	-6.1	¹⁴ C	4.8 $\pm$ 0.54 ^b	-1.02 $\pm$ 0.34	Easton and Olson (1976)
105	Porites	21.27	-157.7 ^d	-7.9	¹⁴ C	5.92 $\pm$ 0.35 ^b	-1.4 $\pm$ 0.41	Easton and Olson (1976)
106	Porites	21.27	-157.7 ^d	-9.8	¹⁴ C	6.05 $\pm$ 0.2 ^b	-1.61 $\pm$ 0.4	Easton and Olson (1976)
107	Porites	21.27	-157.7 ^d	-12.2	¹⁴ C	5.99 $\pm$ 0.45 ^b	-2.12 $\pm$ 0.68	Easton and Olson (1976)
108	Porites	21.27	-157.7 ^d	-13.1	¹⁴ C	5.76 $\pm$ 0.37 ^b	-2.3 $\pm$ 0.34	Easton and Olson (1976)
111	Algae	21.27	-157.7 ^d	-0.3	¹⁴ C	0.23 $\pm$ 0.17 ^b	-2.48 $\pm$ 6.93	Easton and Olson (1976)
117	Algae	21.27	-157.7 ^d	-0.3	¹⁴ C	0.67 $\pm$ 0.16 ^b	-3.04 $\pm$ 2.43	Easton and Olson (1976)
117	Algae	21.27	-157.7 ^d	-0.3	¹⁴ C	0.74 $\pm$ 0.18 ^b	-3.07 $\pm$ 2.34	Easton and Olson (1976)
118	Pocillopora	21.27	-157.7 ^d	-0.3	¹⁴ C	0.75 $\pm$ 0.17 ^b	-3.07 $\pm$ 2.23	Easton and Olson (1976)
74-6-9-3	Porites	21.45	-157.74 ^e	5.8	U-Th	122 $\pm$ 11	0.1 $\pm$ 0.31	Easton and Ku (1981)
74-7-20-3	Porites	21.66	-157.92 ^e	0.5	U-Th	120 $\pm$ 9	0.06 $\pm$ 0.21	Easton and Ku (1981)
KP3-2		21.71	-157.98 ^d	3	ESR	139 $\pm$ 14	0.81 $\pm$ 0.53	Jones (1993a)
LL-3		21.28	-157.69 ^d	3	ESR	312 $\pm$ 79	0.11 $\pm$ 0.2	Jones (1993a)
Beta-76292	Porites compressa	21.43	-157.79 ^e	2.1	¹⁴ C	2.95 $\pm$ 0.17 ^b	-0.03 $\pm$ 0.36	Grossman and Fletcher (1998)
Beta-76293	Porites compressa	21.43	-157.79 ^e	2.1	¹⁴ C	3.08 $\pm$ 0.2 ^b	0.08 $\pm$ 0.38	Grossman and Fletcher (1998)
Beta-79905	Porites compressa	21.43	-157.79 ^e	1.7	¹⁴ C	3.4 $\pm$ 0.18 ^b	0.18 $\pm$ 0.29	Grossman and Fletcher (1998)
AA-18218	Porites compressa	21.43	-157.79 ^e	1.7	¹⁴ C	0.31 $\pm$ 0.2 ^b	3.79 $\pm$ 6.67	Grossman and Fletcher (1998)
AA-18219	Porites compressa	21.43	-157.79 ^e	1.7	¹⁴ C	3.87 $\pm$ 0.15 ^b	0.42 $\pm$ 0.24	Grossman and Fletcher (1998)
AA-18221	Porites compressa	21.43	-157.79 ^e	1.5	¹⁴ C	3.87 $\pm$ 0.16 ^b	0.38 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18223	Porites reticulata	21.43	-157.79 ^e	1.5	¹⁴ C	3.89 $\pm$ 0.16 ^b	0.39 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18225	Porites compressa	21.43	-157.79 ^e	1.4	¹⁴ C	2.61 $\pm$ 0.19 ^b	-0.67 $\pm$ 0.49	Grossman and Fletcher (1998)
AA-18226	Porites reticulata	21.43	-157.79 ^e	1.4	¹⁴ C	3.54 $\pm$ 0.14 ^b	0.18 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18227	Porites compressa	21.43	-157.79 ^e	1.8	¹⁴ C	4.09 $\pm$ 0.19 ^b	0.54 $\pm$ 0.26	Grossman and Fletcher (1998)
AA-18228	Porites compressa	21.43	-157.79 ^e	1.8	¹⁴ C	3.71 $\pm$ 0.17 ^b	0.38 $\pm$ 0.27	Grossman and Fletcher (1998)
AA-18229	Porites compressa	21.43	-157.79 ^e	2.1	¹⁴ C	3.63 $\pm$ 0.15 ^b	0.43 $\pm$ 0.26	Grossman and Fletcher (1998)
AA-18231	Porites compressa	21.43	-157.79 ^e	2.4	¹⁴ C	2.97 $\pm$ 0.16 ^b	0.09 $\pm$ 0.35	Grossman and Fletcher (1998)
AA-18232	Pocillopora	21.43	-157.79 ^e	2.7	¹⁴ C	2.37 $\pm$ 0.15 ^b	-0.48 $\pm$ 0.46	Grossman and Fletcher (1998)
AA-18233	Porites reticulata	21.43	-157.79 ^e	2.7	¹⁴ C	3.51 $\pm$ 0.13 ^b	0.54 $\pm$ 0.26	Grossman and Fletcher (1998)
AA-18234	Porites compressa	21.43	-157.79 ^e	1.9	¹⁴ C	3.94 $\pm$ 0.27 ^b	0.51 $\pm$ 0.34	Grossman and Fletcher (1998)
AA-18235	Cypraea	21.43	-157.79 ^e	1.9	¹⁴ C	3.78 $\pm$ 0.15 ^b	0.44 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18236	Porites reticulata	21.43	-157.79 ^e	1.9	¹⁴ C	3.5 $\pm$ 0.13 ^b	0.3 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18237	Porites compressa	21.43	-157.79 ^e	2.3	¹⁴ C	2.94 $\pm$ 0.15 ^b	0.03 $\pm$ 0.34	Grossman and Fletcher (1998)
AA-18238	Porites compressa	21.43	-157.79 ^e	1.8	¹⁴ C	3.82 $\pm$ 0.15 ^b	0.42 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18239	Porites reticulata	21.43	-157.79 ^e	1.8	¹⁴ C	3.53 $\pm$ 0.13 ^b	0.28 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18240	Porites compressa	21.43	-157.79 ^e	2.2	¹⁴ C	3.59 $\pm$ 0.14 ^b	0.42 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18241	Porites reticulata	21.43	-157.79 ^e	2.2	¹⁴ C	3.34 $\pm$ 0.17 ^b	0.29 $\pm$ 0.29	Grossman and Fletcher (1998)
AA-18242	Porites reticulata	21.43	-157.79 ^e	2.5	¹⁴ C	2.85 $\pm$ 0.21 ^b	0.01 $\pm$ 0.42	Grossman and Fletcher (1998)
AA-18243	T. variabilis	21.43	-157.79 ^e	2.5	¹⁴ C	3.52 $\pm$ 0.13 ^b	0.49 $\pm$ 0.25	Grossman and Fletcher (1998)
AA-18244	Porites compressa	21.43	-157.79 ^e	2.5	¹⁴ C	2.73 $\pm$ 0.16 ^b	-0.11 $\pm$ 0.38	Grossman and Fletcher (1998)
AA-18245	Pocillopora	21.43	-157.79 ^e	2.8	¹⁴ C	2.32 $\pm$ 0.15 ^b	-0.52 $\pm$ 0.46	Grossman and Fletcher (1998)
AA-18246	Pocillopora	21.43	-157.79 ^e	2.8	¹⁴ C	1.42 $\pm$ 0.11 ^b	-1.01 $\pm$ 0.73	Grossman and Fletcher (1998)
Beta-80884	Porites compressa	21.39	-157.7 ^e	0.5	¹⁴ C	0.51 $\pm$ 0.09 ^b	-1.37 $\pm$ 1.95	Grossman and Fletcher (1998)
19-Pr1	Porites	21.45	-157.72 ^d	7.3	U-Th	129.9 $\pm$ 1.5	0.39 $\pm$ 0.1	Szabo et al. (1994)
19-Pr2	Porites	21.45	-157.72 ^d	7.3	U-Th	131 $\pm$ 2.7	0.45 $\pm$ 0.22	Szabo et al. (1994)
19-Pc2	Pocillopora	21.45	-157.72 ^d	8.6	U-Th	125 $\pm$ 3.8	0.11 $\pm$ 0.19	Szabo et al. (1994)
1-Pr1	Porites	21.71	-157.98 ^d	3	U-Th	120.3 $\pm$ 2.5	0.09 $\pm$ 0.05	Szabo et al. (1994)
2-Pr1	Porites	21.71	-157.98 ^d	3	U-Th	122.1 $\pm$ 1.3	0.08 $\pm$ 0.03	Szabo et al. (1994)
13-Pr1	Porites	21.71	-157.98 ^d	3	U-Th	121.7 $\pm$ 1.5	0.08 $\pm$ 0.03	Szabo et al. (1994)

(continued on next page)

Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
25-Pc1	<i>Pocillopora</i>	21.26	−157.79 ^d	17	U–Th	247 $\pm$ 69	0.46 $\pm$ 0.34	Szabo et al. (1994)
HA-18D(1)	<i>Pocillopora</i>	21.45	−157.72 ^d	12.4	U–Th	127 $\pm$ 4	0.29 $\pm$ 0.27	Muhs and Szabo (1994)
HA-18A(2)	<i>Pocillopora</i>	21.45	−157.72 ^d	12.4	U–Th	134 $\pm$ 4	0.82 $\pm$ 0.44	Muhs and Szabo (1994)
GX-2673	Coral in beachrock	21.27	−157.7 ^e	2.7	¹⁴ C	3.47 $\pm$ 0.41 ^b	0.51 $\pm$ 0.51	Stearns (1977)
Beta-55347	<i>Pocillopora</i>	21.43	−157.79 ^d	−1.5	¹⁴ C	0.35 $\pm$ 0.15 ^b	−6.22 $\pm$ 5.82	Fletcher and Jones (1996)
Beta-55346	<i>Porites compressa</i>	21.43	−157.79 ^d	2.8	¹⁴ C	2.92 $\pm$ 0.16 ^b	0.17 $\pm$ 0.36	Fletcher and Jones (1996)
Beta-55351	<i>Porites compressa</i>	21.43	−157.79 ^d	2.8	¹⁴ C	1.89 $\pm$ 0.23 ^b	−1.29 $\pm$ 0.69	Fletcher and Jones (1996)
Beta-55350	<i>Porites compressa</i>	21.43	−157.79 ^d	2.8	¹⁴ C	2.9 $\pm$ 0.15 ^b	0.16 $\pm$ 0.36	Fletcher and Jones (1996)
Beta-55348	<i>Porites compressa</i>	21.43	−157.79 ^d	2.5	¹⁴ C	2.4 $\pm$ 0.19 ^b	−0.54 $\pm$ 0.51	Fletcher and Jones (1996)
Beta-55354	<i>Spondylus</i>	21.43	−157.79 ^d	1.6	¹⁴ C	2.02 $\pm$ 0.21 ^b	−1.62 $\pm$ 0.76	Fletcher and Jones (1996)
Beta-55345	<i>Porites compressa</i>	21.43	−157.79 ^d	1.6	¹⁴ C	3.13 $\pm$ 0.18 ^b	−0.03 $\pm$ 0.34	Fletcher and Jones (1996)
Beta-55352	<i>Periglypta</i>	21.43	−157.79 ^d	1.6	¹⁴ C	3.52 $\pm$ 0.28 ^b	0.22 $\pm$ 0.36	Fletcher and Jones (1996)
Beta-55356	<i>Cellana</i>	21.39	−157.7 ^d	1.4	¹⁴ C	2.56 $\pm$ 0.19 ^b	−0.76 $\pm$ 0.49	Fletcher and Jones (1996)
Beta-55357	<i>Porites compressa</i>	21.39	−157.7 ^d	1.4	¹⁴ C	3.21 $\pm$ 0.21 ^b	−0.03 $\pm$ 0.35	Fletcher and Jones (1996)
Beta-62123	<i>Porites compressa</i>	21.39	−157.7 ^d	1.4	¹⁴ C	2.88 $\pm$ 0.13 ^b	−0.35 $\pm$ 0.34	Fletcher and Jones (1996)
Beta-62124	<i>Porites compressa</i>	21.39	−157.7 ^d	1.4	¹⁴ C	2.39 $\pm$ 0.16 ^b	−1.01 $\pm$ 0.5	Fletcher and Jones (1996)
Beta-62125	<i>Porites compressa</i>	21.39	−157.7 ^d	1.4	¹⁴ C	3.42 $\pm$ 0.13 ^b	0.11 $\pm$ 0.24	Fletcher and Jones (1996)
Beta-62126	<i>Cypraea</i>	21.39	−157.7 ^d	1.4	¹⁴ C	3.11 $\pm$ 0.18 ^b	−0.12 $\pm$ 0.35	Fletcher and Jones (1996)
Beta-62127	<i>Cypraea</i>	21.39	−157.7 ^d	1.4	¹⁴ C	2.37 $\pm$ 0.15 ^b	−1.05 $\pm$ 0.49	Fletcher and Jones (1996)
Beta-62128	<i>Cypraea</i>	21.39	−157.7 ^d	1.4	¹⁴ C	2.76 $\pm$ 0.14 ^b	−0.49 $\pm$ 0.37	Fletcher and Jones (1996)
HA-3-A	<i>Porites</i>	21.4	−157.73 ^d	3	U–Th	119.4 $\pm$ 3	0.06 $\pm$ 0.05	Muhs et al. (2002)
HA-3-A dup	<i>Porites</i>	21.4	−157.73 ^d	3	U–Th	121.5 $\pm$ 5	0.09 $\pm$ 0.07	Muhs et al. (2002)
HA-3-B	<i>Porites</i>	21.4	−157.73 ^d	3	U–Th	133 $\pm$ 0.8	0.54 $\pm$ 0.16	Muhs et al. (2002)
HA-3-B dup	<i>Porites</i>	21.4	−157.73 ^d	3	U–Th	134.5 $\pm$ 1.4	0.82 $\pm$ 0.24	Muhs et al. (2002)
HA-3-D	<i>Porites</i>	21.4	−157.73 ^d	3	U–Th	129.9 $\pm$ 0.9	0.36 $\pm$ 0.06	Muhs et al. (2002)
HA-13-B	<i>Porites</i>	21.71	−157.98 ^d	3	U–Th	113 $\pm$ 0.7	0.3 $\pm$ 0.01	Muhs et al. (2002)
HA-13-C	<i>Porites</i>	21.71	−157.98 ^d	3	U–Th	121 $\pm$ 1.3	0.1 $\pm$ 0.02	Muhs et al. (2002)
3-Pr1(C)	<i>Porites</i>	21.4	−157.73 ^d	3	U–Th	130.2 $\pm$ 2.5	0.38 $\pm$ 0.16	Szabo et al. (1994)
3-Pc1(C)	<i>Pocillopora</i>	21.4	−157.73 ^d	3	U–Th	128.5 $\pm$ 6.1	0.29 $\pm$ 0.47	Szabo et al. (1994)
2-Pc1(C)	<i>Pocillopora</i>	21.71	−157.98 ^d	3	U–Th	114.1 $\pm$ 1.8	0.3 $\pm$ 0.11	Szabo et al. (1994)
	Slope break			−950	K–Ar	2600 $\pm$ 200	−0.35 $\pm$ 0.04	Clague and Dalrymple (1987)
					Wai'anae			
LJ-899	Shells	21.4	−158.18	3.5	¹⁴ C	21.02 $\pm$ 1.39 ^b	5.41 $\pm$ 0.57	Hubbs et al. (1965)
OAHU22	<i>Leptastrea</i>	21.59	−158.11 ^e	1.5	U–Th	140 $\pm$ 30	0.9 $\pm$ 0.71	Veeh (1966)
OAHU22	<i>Leptastrea</i>	21.59	−158.11 ^e	1.5	U–Th	140 $\pm$ 50	0.9 $\pm$ 0.77	Veeh (1966)
C9	<i>Pocillopora</i>	21.56	−158.25 ^d	3.7	U–Th	115 $\pm$ 12	0.24 $\pm$ 0.21	Ku et al. (1974)
74-6-19-1a	<i>Porites</i>	21.32	−157.92 ^e	2.8	U–Th	135 $\pm$ 6	0.9 $\pm$ 0.49	Easton and Ku (1981)
74-6-8-1	<i>Porites</i>	21.35	−158.13 ^e	11.6	U–Th	142 $\pm$ 12	0.78 $\pm$ 0.37	Easton and Ku (1981)
BPC-10		21.31	−158.09 ^d	5	ESR	122 $\pm$ 8	0.1 $\pm$ 0.22	Jones (1993a)
BPC-2A		21.31	−158.09 ^d	5	ESR	142 $\pm$ 20	0.73 $\pm$ 0.57	Jones (1993a)
BPC-9		21.31	−158.09 ^d	5	ESR	152 $\pm$ 25	0.47 $\pm$ 0.42	Jones (1993a)
WB		21.33	−158.1 ^d	14	ESR	196 $\pm$ 29	0.06 $\pm$ 0.2	U. Radtke cited in Jones (1993b)
WHC		21.43	−158.18 ^d	26.5	ESR	468 $\pm$ 136	0.28 $\pm$ 0.2	Jones (1993a)
KAE-3		21.57	−158.27 ^d	26.5	ESR	547 $\pm$ 82	0.19 $\pm$ 0.1	Jones (1993a)
BPC-6		21.31	−158.09 ^d	2	ESR	562 $\pm$ 96	0.13 $\pm$ 0.1	Jones (1993a)
D-1407	Coral	21.57	−158.27 ^e	30	ESR	507 $\pm$ 50	0.21 $\pm$ 0.08	Bruckner and Radtke (1989)
D-1437	Coral	21.57	−158.27 ^e	30	ESR	499 $\pm$ 50	0.16 $\pm$ 0.08	Bruckner and Radtke (1989)
D-1509-1	Coral	21.57	−158.27 ^e	30	ESR	449 $\pm$ 45	0.28 $\pm$ 0.19	Bruckner and Radtke (1989)
D-1509-2	Coral	21.57	−158.27 ^e	30	ESR	406 $\pm$ 40	0.09 $\pm$ 0.2	Bruckner and Radtke (1989)
D-1635a	Coral	21.57	−158.27 ^e	30	ESR	547 $\pm$ 55	0.2 $\pm$ 0.07	Bruckner and Radtke (1989)
20-Pr1	<i>Porites</i>	21.58	−158.25 ^d	3	U–Th	110.5 $\pm$ 3.8	0.29 $\pm$ 0.09	Szabo et al. (1994)
21-Pr1	<i>Porites</i>	21.58	−158.25 ^d	3	U–Th	114.8 $\pm$ 2.1	0.26 $\pm$ 0.18	Szabo et al. (1994)
27-Pc1	<i>Pocillopora</i>	21.57	−158.27 ^d	24.4	U–Th	532 $\pm$ 130	0.21 $\pm$ 0.16	Szabo et al. (1994)

Table 1 (continued)

Sample	Type	Lat.	Long.	Elev. (m)	Method	Age $\pm 2\sigma$ (kyr)	Vertical motion rate $\pm 2\sigma$ (mm yr ⁻¹)	Reference
HA-7X	<i>Pocillopora</i>	21.35	−158.13 ^d	11	U–Th	120 $\pm$ 3	0.15 $\pm$ 0.05	Muhs and Szabo (1994)
HA-7Y	<i>Porites</i>	21.35	−158.13 ^d	11	U–Th	134 $\pm$ 4	0.81 $\pm$ 0.43	Muhs and Szabo (1994)
MA15-1S1		21.41	−158.19 ^d	−10	U–Th	223.3 $\pm$ 1.4	0.17 $\pm$ 0.04	Sherman et al. (1999)
WA18-3S1A		21.45	−158.21 ^d	−27	U–Th	96.8 $\pm$ 6	0.09 $\pm$ 0.16	Sherman et al. (1999)
WA16-1S1		21.45	−158.21 ^d	−30	U–Th	82.8 $\pm$ 5	−0.13 $\pm$ 0.29	Sherman et al. (1999)
HA-7-Y	<i>Porites</i>	21.35	−158.13 ^d	11	U–Th	119.5 $\pm$ 1.4	0.13 $\pm$ 0.06	Muhs et al. (2002)
HA-7-Y dup	<i>Porites</i>	21.35	−158.13 ^d	11	U–Th	119.6 $\pm$ 1.1	0.14 $\pm$ 0.05	Muhs et al. (2002)
HA-7-Y dup#2	<i>Porites</i>	21.35	−158.13 ^d	11	U–Th	119.4 $\pm$ 0.6	0.13 $\pm$ 0.03	Muhs et al. (2002)
HA-15-(1)	<i>Porites</i>	21.58	−158.25 ^d	5	U–Th	126.2 $\pm$ 0.7	0.18 $\pm$ 0.07	Muhs et al. (2002)
HA-15-3	<i>Pocillopora</i>	21.58	−158.25 ^d	5	U–Th	115.2 $\pm$ 0.9	0.23 $\pm$ 0.11	Muhs et al. (2002)
HA-15-(4)	<i>Porites</i>	21.58	−158.25 ^d	5	U–Th	124.3 $\pm$ 1.2	0.06 $\pm$ 0.04	Muhs et al. (2002)
HA-15-(4) dup	<i>Porites</i>	21.58	−158.25 ^d	5	U–Th	122.1 $\pm$ 0.9	0.09 $\pm$ 0.02	Muhs et al. (2002)
HA-20-B	<i>Porites</i>	21.58	−158.25 ^d	2	U–Th	117.9 $\pm$ 0.6	0.04 $\pm$ 0.02	Muhs et al. (2002)
HA-20-(2)	<i>Porites</i>	21.58	−158.25 ^d	2	U–Th	119.3 $\pm$ 0.8	0.05 $\pm$ 0.04	Muhs et al. (2002)
HA-21-(1)	<i>Porites</i>	21.58	−158.25 ^d	2	U–Th	126.2 $\pm$ 0.7	0.15 $\pm$ 0.07	Muhs et al. (2002)
26-Pc1(C)	<i>Pocillopora</i>	21.58	−158.25 ^d	3	U–Th	121.3 $\pm$ 1.4	0.09 $\pm$ 0.02	Szabo et al. (1994)
8-Pr1(C)	<i>Porites</i>	21.35	−158.13 ^d	11	U–Th	117.6 $\pm$ 1.9	0.12 $\pm$ 0.07	Szabo et al. (1994)
7-Pr1(C)	<i>Porites</i>	21.35	−158.13 ^d	11	U–Th	113.3 $\pm$ 4.2	0.37 $\pm$ 0.18	Szabo et al. (1994)
7-Pc1(C)	<i>Pocillopora</i>	21.35	−158.13 ^d	11	U–Th	117.8 $\pm$ 1.7	0.12 $\pm$ 0.05	Szabo et al. (1994)
7-Pc2(C)	<i>Pocillopora</i>	21.35	−158.13 ^d	11	U–Th	110.9 $\pm$ 3.5	0.37 $\pm$ 0.08	Szabo et al. (1994)
OA-GM-75	<i>Porites</i>	21.44	−158.18 ^d	17	U–Th	334.2 $\pm$ 17	0.11 $\pm$ 0.23	McMurtry et al. (2010)
OA-GM-93	<i>Porites</i>	21.35	−158.13 ^d	12.5	U–Th	123.6 $\pm$ 0.9	0.13 $\pm$ 0.02	McMurtry et al. (2010)
OA-GM-94	<i>Porites</i>	21.35	−158.13 ^d	12.5	U–Th	120.3 $\pm$ 0.81	0.17 $\pm$ 0.03	McMurtry et al. (2010)
OA-GM-95A	<i>Leptoseris</i>	21.35	−158.13 ^d	12.5	U–Th	125.5 $\pm$ 0.96	0.18 $\pm$ 0.1	McMurtry et al. (2010)
OA-GM-95B	<i>Pocillopora</i>	21.35	−158.13 ^d	12.5	U–Th	122.7 $\pm$ 0.96	0.15 $\pm$ 0.02	McMurtry et al. (2010)
OA-GM-95C	<i>Porites</i>	21.35	−158.13 ^d	12.5	U–Th	121.4 $\pm$ 1.3	0.17 $\pm$ 0.03	McMurtry et al. (2010)
OA-GM-96	<i>Porites brighami</i>	21.34	−158.11 ^d	22.8	U–Th	334.6 $\pm$ 10	0.13 $\pm$ 0.2	McMurtry et al. (2010)
OA-GM-97	<i>Porites</i>	21.34	−158.11 ^d	22.8	U–Th	330.5 $\pm$ 12	0.13 $\pm$ 0.14	McMurtry et al. (2010)
OA-GM-98	<i>Porites</i>	21.34	−158.11 ^d	22.8	U–Th	363.2 $\pm$ 20	0.26 $\pm$ 0.1	McMurtry et al. (2010)
OA-GM-101	<i>Porites</i>	21.34	−158.11 ^d	22.8	U–Th	342.5 $\pm$ 16	0.41 $\pm$ 0.23	McMurtry et al. (2010)
OA-GM-102A	<i>Porites</i>	21.34	−158.11 ^d	22.8	U–Th	302.1 $\pm$ 17	0.3 $\pm$ 0.12	McMurtry et al. (2010)
	Slope break			−1050	K–Ar	3700 $\pm$ 200	−0.29 $\pm$ 0.02	Clague and Dalrymple (1987)
					<i>Kaua'i</i>			
LJ-917	Beachrock	21.96	−159.73	1	¹⁴ C	1.26 $\pm$ 0.34 ^b	−2.14 $\pm$ 2.01	Hubbs et al. (1965)
LJ-753	<i>Porites</i>	22.14	−159.76	−18	¹⁴ C	9.02 $\pm$ 0.57 ^b	1.57 $\pm$ 0.57	Hubbs et al. (1965)
LJ-916	Sandstone	22.12	−159.74	1.8	¹⁴ C	17.79 $\pm$ 1.27 ^b	6.51 $\pm$ 1.1	Hubbs et al. (1965)
	<i>Tellina</i>	22.1	−159.31 ^e	1.1	¹⁴ C	3.93 $\pm$ 0.17 ^b	0.3 $\pm$ 0.25	Matsumoto and Kayanne (1988)
	<i>Tellina</i>	22.1	−159.31 ^e	0.8	¹⁴ C	3.85 $\pm$ 0.22 ^b	0.18 $\pm$ 0.28	Matsumoto and Kayanne (1988)
	<i>Leptachatina</i>	22.1	−159.31 ^e	1.6	¹⁴ C	2.65 $\pm$ 0.17 ^b	−0.56 $\pm$ 0.44	Matsumoto and Kayanne (1988)
	<i>Tellina</i>	22.1	−159.31 ^e	1.2	¹⁴ C	4.04 $\pm$ 0.18 ^b	0.37 $\pm$ 0.25	Matsumoto and Kayanne (1988)
Beta-41951	<i>Porites</i>	22.21	−159.49 ^d	1.8	¹⁴ C	4.39 $\pm$ 0.2 ^b	0.65 $\pm$ 0.26	Jones (1992)
Beta-41952	<i>Porites</i>	22.21	−159.49 ^d	1.5	¹⁴ C	3.7 $\pm$ 0.21 ^b	0.29 $\pm$ 0.29	Jones (1992)
Beta-41953	<i>Porites</i>	22.21	−159.49 ^d	1.5	¹⁴ C	3.16 $\pm$ 0.22 ^b	−0.03 $\pm$ 0.37	Jones (1992)
Beta-41952	<i>Porites</i>	22.21	−159.49 ^d	1.5	¹⁴ C	4.23 $\pm$ 0.23 ^b	0.52 $\pm$ 0.28	Jones (1992)
HTZ-705/710	Shallow marine sediments	22	−159.39	−132	⁴⁰ Ar/ ³⁹ Ar	3220 $\pm$ 260	−0.04 $\pm$ 0.01	Izuka and Sherrod (2011)
				−1100	K–Ar	5100 $\pm$ 400	−0.22 $\pm$ 0.02	Clague and Dalrymple (1987)

^a Assumed 10% 2 σ age uncertainty.^b Calibrated age computed from reported radiocarbon age using the program CALIB 6.1 and the Marine09 calibration curve for carbonate material (Reimer et al., 2009), assuming a local ocean reservoir deviation ΔR for Hawai'i of -38 ± 3 (Druffel et al., 2001).^c Upslope terrace elevation.^d Sample location inferred from map.^e Sample location estimated, from described location.

Appendix B. Vertical motion rates from tide gauge records

See Table 2.

Table 2

Vertical motion rates from tide gauge records.

Station	Observation period	Months of record	Relative sea level change (mm yr ⁻¹) ^a	Vertical motion rate (mm yr ⁻¹) ^b
Hilo, Hawai'i	1946–2011	773	3.05 ± 0.33	−1.25 ± 0.68
Kahulunui, Maui	1947–2011	540	2.07 ± 0.47	−0.27 ± 0.76
Mokuoloe, O'ahu	1957–2011	593	1.12 ± 0.61	0.68 ± 0.86
Honolulu, O'ahu	1905–2011	1208	1.42 ± 0.23	0.28 ± 0.64 ^c
Nawiliwili, Kaua'i	1954–2011	682	1.35 ± 0.51	0.45 ± 0.79

^a Reported error is the 95% confidence interval on the regression slope of month-averaged hourly sea level measurements detrended for regular seasonal fluctuations.

^b Calculated as the difference between station relative sea level change and 1.8 ± 0.3 mm yr⁻¹ global-averaged sea level change, 1950–2004, from monthly tide gauge records and satellite altimetry (estimated s.e.). Reported error is the station sea level change 95% confidence interval and the 2σ global sea level change error summed in quadrature.

^c Calculated using 1.7 ± 0.3 mm yr⁻¹ global-averaged sea level change, 1870–2004, from monthly tide gauge records and satellite altimetry (estimated s.e.).

Appendix C. Timescale dependence of island vertical motion rates

Rates of sedimentary, geomorphic, and paleontologic processes often decrease as a power-law function of their measurement interval (Sadler, 1993). This negative power-law dependence on measurement interval, or “Sadler effect”, has been attributed to non-steady-state behavior: averaged over longer periods of time, rates of sediment deposition, bedrock river incision, and other episodic processes incorporate longer hiatuses, resulting in an apparent decline in their tempo over longer timescales (Sadler, 1993; Finnegan et al., 2014).

To determine if Hawaiian Island vertical motion rates vary systematically as a function of averaging time, we computed the power-law relationship between cumulative uplift or subsidence and measurement interval (Fig. 11). Plotting the cumulative uplift or subsidence ($z_{\text{sample}} + z_{\text{crit}} - z_{\text{psi}}$), rather than the rates of uplift or subsidence $\dot{z}_{\text{mod}}$ against measurement interval τ_{obs} permits the timescale dependence of the rates to be analyzed without introducing spurious correlation from regressing the rates against their own denominator. The timescale dependence of the vertical motion rates is then obtained by subtracting one from the cumulative uplift or subsidence versus measurement interval power-law exponent β (Finnegan et al., 2014).

Both uplift and subsidence display power-law exponents $\beta < 1$, corresponding to a negative power-law scaling between vertical motion rates and their measurement intervals over 10^2 to 10^6 year timescales. However, β is close to one for both uplift and subsidence rates, indicating only a weak dependence of vertical motion rates on measurement interval. Numerous factors may contribute to the signature of unsteadiness in the vertical motion rate observations. Drill cores from Hilo, Hawai'i indicate high variability in accumulation rates for lava flows erupted from Mauna Loa over the past several thousand years (Lipman and Moore, 1996). Because space geodesy and geochronology of geomagnetic reversals suggest that seafloor spreading rates are steady (within 2% to 6%) over several year to several million-year timescales (DeMets et al., 1994), the observed variations are probably not due to unsteadiness in plate migration rates.

Uplift rates in the Hawaiian Islands display a stronger timescale dependence ($\beta = 0.86 \pm 0.04$, corresponding to a rate-interval power-law exponent of -0.14 ± 0.04) than subsidence rates ($\beta = 0.96 \pm 0.03$, corresponding to a rate-interval power-law exponent of -0.04 ± 0.03). The stronger timescale dependence of uplift

rates could result from a preservation bias associated with glacio-eustatic sea level change. Most uplifted paleoshorelines in the Hawaiian Islands formed during interglacial sea level highstands while a lesser majority of drowned coral terraces formed during glacial lowstands (Appendix A). Because subsidence rates in the Hawaiian Islands can significantly exceed even the fastest rates of glacio-eustatic sea level fall, coral reefs can drown even when sea level is falling. Maximum uplift rates, on the other hand, are slower than typical rates of glacio-eustatic sea level rise, so corals can only be stranded above sea level during periods of sea level fall or relatively slow sea level rise. This implies that island subsidence can be recorded more steadily through time as sea level fluctuates, whereas island uplift may only be selectively recorded by paleoshorelines. This could increase the timescale dependence of uplift rates if anomalous episodes of rapid island uplift are required to overcome glacio-eustatic sea level rise and generate emerged paleoshorelines. The greater potential for uplifted shoreline deposits to be reworked, weathered, and effectively erased from record may further enhance the timescale dependence of uplift rates if, for instance, only the highest paleoshorelines associated with episodes of rapid island uplift are preserved in the paleoshoreline record.

Appendix D. Sensitivity of model results to volcanic loading timescales and regional oversampling

D.1. Averaging timescales of vertical motion rates due to volcanic loading

We used K-Ar ages of shield stage flows τ_v (Robinson and Eakins, 2006) to constrain the timescales of vertical motion for volcanic loading $\dot{z}_v$. With the exception of Hualālai, where no shield-stage flows are exposed, and Ka'ula, where only post-shield flows have been dated, these ages represent the most frequent (modal) radiometric ages determined for shield-stage flows on the surface of each volcano. Because accessible surface flows for each volcano should correspond to the end of shield building (to the extent that (a) shield stage flows are exposed beneath post-shield flows and (b) the youngest shield-stage flows have withstood surface weathering and erosion) our calculation of vertical motion due to volcanic loading $\dot{z}_v$ (the sum of the deflection for all volcanoes with ages less than or equal to a given age τ_v divided by that age τ_v) assumes that all deflection resulting from the construction of a volcano occurs after the end of its shield-building phase. Although we assumed in the flexure calculations that the upper mantle behaves as an inviscid fluid over the timescales of interest, this calculation represents the case where lithospheric deformation lags load emplacement, as expected for a viscous substratum. It therefore yields an upper bound on vertical motion rates $\dot{z}_v$ averaged to the present.

To calculate a lower bound on vertical motion rates due to volcanic loading $\dot{z}_v$, we divided the deflection resulting from all volcanoes with ages less than a given age τ_v by that age τ_v . This calculation assumes that all deflection resulting from the construction of a given volcano occurs before the end of its shield building phase. The modeled rates $\dot{z}_{\text{mod}}$ calculated with these volcanic loading rates $\dot{z}_v$ yield slightly lower best-fit effective elastic thicknesses T_e of 35 km ($\Phi = 0.1106$) and 40 km ($\Phi = 0.1122$) for spectral and spatial median filtering, respectively. Model Φ is lower than the Φ of the mean observed vertical motion rate ($\langle \dot{z}_{\text{obs}} \rangle = -0.30$ mm yr⁻¹, $\Phi = 0.1316$) over a slightly larger range of effective elastic thicknesses than for the upper bound rates: T_e between 30 km and 80 km for spatial median filtering and T_e between 25 km and 80 km for spectral filtering. Comparing model Φ to the Φ obtained for the mean observed vertical motion rate of each specific volcanic loading interval ($\Phi = 0.1111$), only spec-

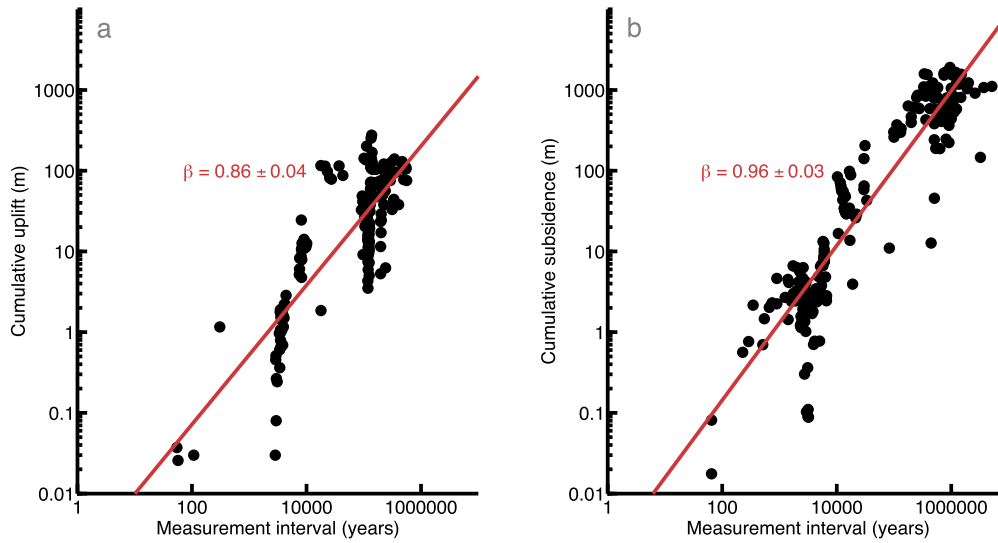


Fig. 11. Cumulative uplift (a) or cumulative subsidence (b) of paleoshorelines in the Hawaiian Islands ($n = 459$) plotted against their radiometric age (measurement interval). Solid line displays the ordinary least-squares regression between log-transformed cumulative uplift/subsidence and measurement interval with slope β ($\pm$ s.e.). $\beta = 1$ corresponds to no dependence of vertical motion rate on measurement interval.

tral filtering and $T_e = 35$ km provided a better fit to the observed rates.

To further test the sensitivity of our results to the averaging timescales of volcanic loading, we assumed that shield building began 500 kyr prior to the radiometric ages for each volcano τ_v and calculated rates of vertical motion for volcanic loading $\dot{z}_v$ by dividing the deflection resulting from all volcanoes with ages less than or equal to a given age τ_v by $(\tau_v + 500 \text{ kyr})$. This calculation assumes a fixed duration of shield building relative to each radiometric age τ_v but, by including the approximate duration of shield building events within the averaging periods, makes no assumptions about the response time of the lithosphere to load emplacement. For these estimates, an effective elastic thickness T_e of 45 km produces the best fit to the rate observations for both the spectral filter ($\Phi = 0.1049$) and the spatial median filter ($\Phi = 0.1026$). Here, model Φ is lower than the Φ of the mean observed vertical motion rate over a wide range of effective elastic thicknesses: T_e between 20 km and 80 km for spatial median filtering and T_e between 15 km and 80 km for spectral filtering. Effective elastic thicknesses T_e between 30 km and 55 km also provide a better fit to the rate observations than the timescale-binned mean vertical motion rates for both spectral filtering and spatial median filtering.

D.2. Binning of rate observations

Without a time-resolved history of the volcanism at the Hawaiian hotspot, the rates that we calculate for volcanic loading $\dot{z}_v$ are intended to average over the entire period of shield building for individual volcanoes, to the extent that this can be constrained by their radiometric ages, plus any preceding or subsequent quiescent periods of pre-shield or post-shield stage volcanism. The modeled rates $\dot{z}_v$ therefore average over a subset of specific timescales τ_v that do not correspond exactly to the averaging timescales of the rate observations τ_{obs} . To compare the rate observations to the modeled rates, we binned the rate observations in three different ways.

Initially, we compared each observed rate $\dot{z}_{\text{obs}}$ to a modeled rate for volcanic loading $\dot{z}_v$ averaged over the nearest timescale τ_v greater than or equal to the observed rate averaging timescale τ_{obs} . To test the sensitivity of our results to binning of the observed rates, we also compared each rate observation to modeled

rates averaged over the nearest (shorter or longer) timescale τ_v to each τ_{obs} . Here we found best-fit effective elastic thicknesses T_e of 45 km ($\Phi = 0.1122$) and 50 km ($\Phi = 0.1165$) for the spectral and spatial median filtering, respectively, and narrower ranges of effective elastic thicknesses (T_e between 40 km and 75 km for spatial median filtering and T_e between 35 km and 75 km for spectral filtering) yielding lower Φ than the mean observed vertical motion rate. In contrast, a larger range of effective elastic thicknesses (T_e between 45 km and 55 km for spatial median filtering and T_e between 40 km and 55 km for spectral filtering) yield model Φ lower than the Φ of the mean observed vertical motion rates within each volcanic loading interval ($\Phi = 0.1192$).

Finally, we restricted our analysis to include only the 69 rate observations averaged over timescales τ_{obs} within ± 50 kyr of the modeled rate timescales τ_v . This yielded a best-fit elastic thickness T_e of 30 km for both the spectral filter ($\Phi = 0.1526$) and the spatial median filter ($\Phi = 0.1585$). Here, model Φ is lower than the Φ of both the mean of the 69 observed rates ($\Phi = 0.1932$) and the mean observed rate within each volcanic loading time interval ($\Phi = 0.1821$) for spatial median filtering and T_e between 30 km and 80 km or for spectral filtering and T_e between 25 km and 80 km.

D.3. Geographic distribution of rate observations

Uplift rates from the intensely studied emerged terraces on O'ahu ($n = 106$) and the elevated deposits on Lāna'i ($n = 33$) are disproportionately represented in the dataset of rate observations $\dot{z}_{\text{obs}}$, while tide gauge records at Nāwiliwili and shallow marine sediments in a Hanana'ulu well drill core provide the only vertical motion rates for Kaua'i. To test the sensitivity of our results to the uneven geographic distribution of rate observations, we compared modeled rates $\dot{z}_{\text{mod}}$ to the mean and median observed rate on each volcano for each averaging time period τ_v . For volcanoes and time periods with multiple rate observations, we used the standard deviation of the rate observations to calculate Φ rather than the uncertainties $\delta \dot{z}_{\text{obs}}$ on individual rates. We found best-fit effective elastic thicknesses T_e of 40 km (mean $\dot{z}_{\text{mod}}$ $\Phi = 0.1239$, median $\dot{z}_{\text{mod}}$ $\Phi = 0.1249$) and 45 km (mean $\dot{z}_{\text{mod}}$ $\Phi = 0.1265$, median $\dot{z}_{\text{mod}}$ $\Phi = 0.1277$) using the mean and the median observed rate for each volcano and observation period, respectively. Using the mean observed rate for each volcano and time period ($n = 37$), filtering methods, model Φ is lower than the Φ of their mean value

($\dot{z}_{\text{obs}} = -0.99 \text{ mm yr}^{-1}$, $\Phi = 0.2093$) for either filtering method and T_e between 30 km and 80 km. Using the median observed rate for each volcano and time period, model Φ is also lower than the Φ of their mean value ($\dot{z}_{\text{obs}} = -1.00 \text{ mm yr}^{-1}$, $\Phi = 0.2124$) for either filtering method and T_e between 30 km and 80 km.

Appendix E. Supplementary material

Supplementary material related to this article can be found online at <http://dx.doi.org/10.1016/j.epsl.2015.02.027>. These data include the Google map of the most important areas described in this article.

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