The Innovation, Volume *4*

Supplemental Information

An emerging plume head interacting with the Hawaiian plume tail

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1. Other double chains in the Pacific plate

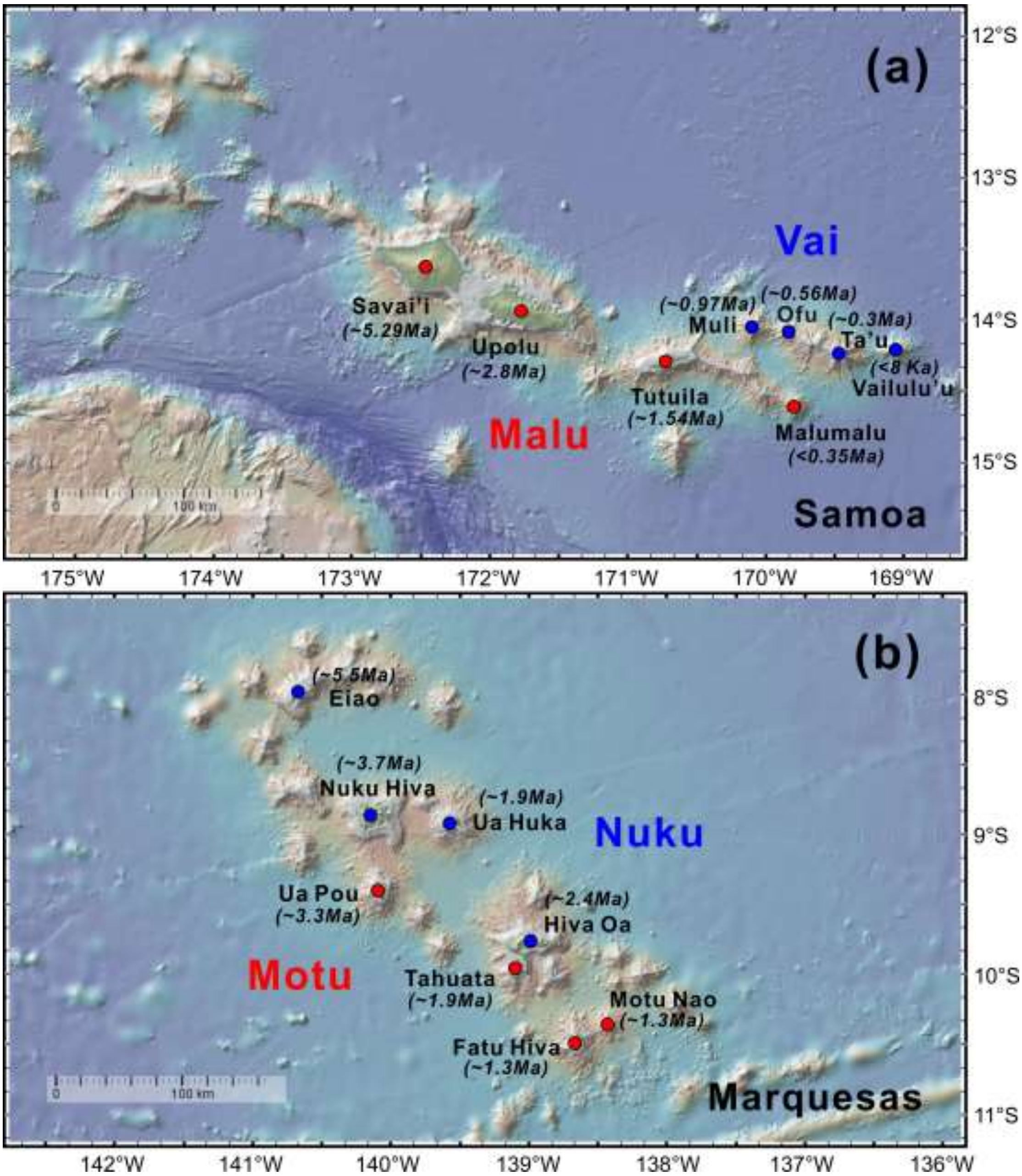

Figure S1 Map showing the other double volcano chains of **(a)** the Samoa and **(b)** the Marquesas plumes. In contrast to the Hawaiian plume, these two plumes are not bended nor increased in eruption. Also, these double chains appeared later than the Hawaiian double chain. The base maps are from the software of GeoMapApp. Age data are from literature as follow.

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2. Geochemical data for each island of the Hawaiian chain

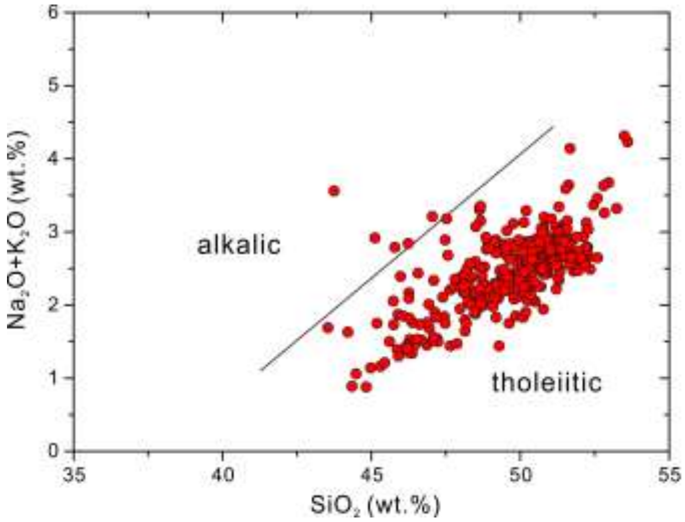

Figure S2 SiO_2 vs. total alkalis diagram. All the samples for which element concentrations are available were calculated as tholeiites. The tholeiitic/alkalic boundary is from Macdonald and Katsura (1964). The references used to determine the composition of each seamount are as follows.

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3. Geodynamic model setup

To study plume-plume interaction, we used a user-updated version of 3D spherical finite element code CitcomS. On finite-element mesh, we solved the conservation equations of mass, momentum, and energy, under the Boussinesq approximation:

$$\nabla \cdot \vec{u} = 0$$

$$-\nabla P + \nabla \cdot [\eta(\nabla\vec{u} + \nabla^T\vec{u})] + \rho_m \alpha \Delta T \vec{g} = 0$$

$$\frac{\partial T}{\partial t} + \vec{u} \cdot \nabla T = \kappa \nabla^2 T$$

where $\vec{u}$ is velocity, P is dynamic pressure, η is effective viscosity, ρ_m is reference mantle density, α is thermal expansion coefficient, ΔT is thermal anomaly, $\vec{g}$ is gravitational acceleration, T is temperature, and κ is thermal diffusivity, respectively.

The model spans 30° × 30° × 2000 km in longitude × latitude × radius, a region much larger than the Hawaiian volcanic chain. The vertical and horizontal resolution are 15 km and 6 km, respectively. Passive tracers are used to track the evolution of two plumes, assuming no extra chemical buoyancy of plume material.

Boundary conditions involve temperature and velocity boundary conditions at surface. The surface potential temperature is set to be 0 °C, beneath the surface we set up a thermal oceanic lithosphere based on half-space cooling model, assuming 80 Myr old seafloor. The surface has been prescribed a constant westward plate motion at 10 cm/yr through the whole simulation.

To model realistic plume structure above 1000 km depth, we fixed heat sources as cones at depth from 1400 km to 1500 km. The resultant plume conduits have spatial averaged excess temperatures varying from 100 °C to 300 °C at LAB depth (~100 km). The present plume conduit under Hawaii has a spatial averaged excess temperature around 300 °C, which is consistent with literature estimates[5,6]. The heat source for Loa trend is located 220 km south to the one for Kea trend, and its base radius is twice as large as that for Kea trend. To simulate the evolution of Hawaiian volcanic chain, we started with the heat source for Kea trend only. After 40 Myr, the Kea trend is formed and became steady. At this time, the heat source for Loa trend was added (Figure S3).

Temperature- and depth-dependent Newtonian rheology is applied for the whole model domain. The background mantle viscosity at the ambient mantle temperature has a 3 layer-profile: lithosphere (0-100km), asthenosphere (100-410 km), and below (410-2000 km). Their respective viscosity values are $0.1*\eta_0$ (10^{20} Pa s), $0.1*\eta_0$ (10^{20} Pa s), η_0 (10^{21} Pa s), where η_0 is the reference viscosity (10^{21} Pa s). The temperature-dependent Newtonian rheology follows:

$$\eta = \eta_{b.g.}(\mathrm{r})\exp(\frac{E(r)}{T+T_0} - \frac{E(r)}{T_m+T_0})$$

where η is effective viscosity, $\eta_{b.g.}$ is background mantle viscosity, E is activation energy, T is temperature, T_0 is temperature offset, and T_m is ambient mantle temperature. The activation energy used in this study is the same as in Hu et al. (2018). With strong temperature-dependent Newtonian rheology, our strong lithosphere reaches 10^{23} Pa s in the upper part and smoothly transient to weak asthenosphere. Other physical parameters used are shown in table S2.

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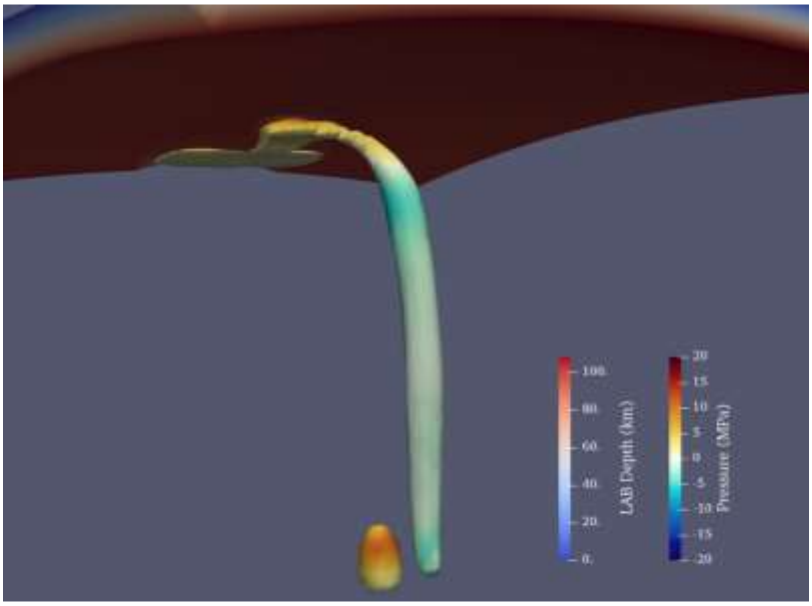

Figure S3 3D view of the Hawaiian plume system at 40 Myr from starting time, when the Loa plume emerged from the lower mantle. View towards northwest direction. The right plume is established Kea plume, the left one is newly formed Loa plume which has a larger size than the Kea plume. Plumes are presented by isothermal surface with 50 ˚C excess temperature.

Table S2 Physical parameters used in geodynamic model

Parameter	Symbol	Value	Unit
Earth radius	R	6371.0	km
Gravitational acceleration	g	9.81	m/s^2
Reference mantle density	ρ	3340	kg/m^3
Reference viscosity	η_0	10^{21}	Pa s
Background viscosity	$\eta_{b.k.}$	10^{20}, 10^{20}, 10^{21}	Pa s
Thermal diffusivity	κ	10^{-6}	m^2/s
Thermal expansivity	α	3.0×10^{-5}	1/°C
Rayleigh number	Ra	5.0e8	-
Activation energy	E	100, 166, 100	kJ/mol
Minimum viscosity	η_{min}	1.0e19	Pa s
Maximum viscosity	η_{max}	1.0e23	Pa s

4. Migration history of plume system center

We identified the centers of Hawaiian plume system in our geodynamic model at 200 km depth which can be used to show the migration of surface volcanic activities. The hottest part of the Hawaiian plume system is defined as the center of the plume system at each output timestep. Since melting processes highly depend on the P-T condition, the hottest part may well represent the center of the plume system with the highest melt production rate at this depth. After identification, the centers were assumed to move with the oceanic lithosphere at the same speed. The temporal and spatial evolution of plume system center is shown in Figure 5 in the main text.