Seismic stratigraphy of Detroit Seamount, Hawaiian-Emperor seamount chain: Post-hot-spot shield-building volcanism and deposition of the Meiji drift

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[1] Detroit Seamount, one of the northernmost seamounts of the Hawaiian-Emperor seamount chain, was formed at ca. 76 Ma. New seismic data suggest renewed volcanism as late as 25 m.y. after initial seamount formation. We use high-resolution single-channel seismic (SCS) data acquired over the summit of Detroit Seamount in 2001 on Ocean Drilling Program (ODP) Leg 197, supplemented by older SCS data acquired as part of the GLORIA mapping program of the U.S. Geological Survey, to characterize the seismic stratigraphy of Detroit Seamount. Volcanic edifices occur on the summit of the seamount and are older than the oldest beds of the Meiji drift (early Oligocene: ca. 34 Ma). On the basis of ash layers in ODP drill holes, we suggest the edifices were active throughout much of the Eocene (ca. 52–34 Ma), with activity possibly extending into the early Oligocene (<34 Ma). Hence the age difference between the shield-building lavas and the postshield cones on Detroit is far greater than the shield/postshield age differences observed on the Hawaiian Islands, suggesting that renewed volcanic activity and tectonic collapse may be possible on any of the Hawaiian Islands. We confirm earlier assertions that the thick sediment cap, Oligocene and younger in age, was deposited by an ocean-bottom current with a southeastward flow direction, along the northeast facing flank of the Emperor Seamount chain. This sediment cap, the Meiji drift, was deposited by a lower-velocity current than many other sediment drifts. A low-angle normal fault, dipping ~19°, suggests topographic collapse of Detroit seamount sometime during the Eocene or late Cretaceous.

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1. Introduction

[1] Detroit Seamount is one of the northernmost surviving volcanoes in the Hawaiian-Emperor Seamount chain (Figure 1). It is a broad volcanic plateau approximately 10,000 km² in area [Lonsdale et al., 1993], with Campanian crestal lava flows (ca. 76 Ma) [Duncan and Keller, 2004],
that rises from abyssal depths to \( \sim 1500 \) m deep (Figure 2). Lonsdale et al. [1993] refers to the shallowest cone as Detroit Seamount and calls the broader plateau Detroit Plateau. However, in this paper we refer to the broader plateau as Detroit Seamount, following the usage of Rea et al. [1993]. A sequence of sediment as thick as 800–900 m blankets most of the edifice [Rea et al., 1993; Shipboard Scientific Party, 2002]. The Meiji drift, constructed of Oligocene to Quaternary pelagic and hemipelagic mud deposited by ocean-bottom currents, forms the bulk of the sediment cap [Creager et al., 1973; Rea et al., 1995a, 1995b; Shipboard Scientific Party, 2002]. Post-hot-spot shield-building volcanic cones rise above the summit platform on Detroit Seamount [Lonsdale et al., 1993]. Some of the cones peak \( \sim 1–2 \) km above the summit platform, and many of the cones crop out above the Meiji drift [Lonsdale et al., 1993].

[3] In July 2001, the JOIDES Resolution obtained basalt cores from two holes drilled near the summit of Detroit Seamount as part of ODP Leg 197’s goal to constrain the paleolatitude of the Hawaiian hot spot [Tarduno et al., 2003]. Drill locations were placed away from faults and eruptive edifices to ensure that the emplacement orientation of the basalt flow units had not been disturbed, which would degrade the accuracy of paleolatitude measurements. To locate and characterize suitable drilling sites we used a Seismograph Services Inc. (SSI) 80-in\(^3\) watergun to collect high-resolution single-channel seismic profiles.

[4] In addition to the digital seismic data gathered by ODP Leg 197, previous survey lines recorded...
over Detroit Seamount were integrated into our study. The U.S. Geological Survey (USGS) collected digital seismic data, as well as gravity, magnetic, and bathymetry recordings over Detroit Seamount as part of the GLORIA Mapping Program [U.S. Geological Survey, 1987] (contact Jon Childs (jchilds@usgs.gov) for seismic availability and information, or see http://walrus.wr.usgs.gov/infobank/f/f287aa/html/f-2-87-aa.meta.html). Lonsdale et al. [1993] collected lines of SeaBeam, magnetic, and analog seismic data in addition to rock dredging over Detroit Seamount.

Figure 2. Map of Detroit Seamount with seismic lines from Leg 197 and lines F41 and F43 of R/V Farnella cruise F-2-87-AA. Detroit, Windsor, and Wayne denote cones imaged by Lonsdale et al. [1993] (black dots); 882, 883, and 884 are ODP Leg 145 site locations, and 1203 and 1204 are ODP Leg 197 site locations. Seismic lines are marked in 10 km increments. Contour map from Rea et al. [1993]. Color map from Smith and Sandwell [1997]. Contour interval is 100 m, and contour labels are in meters.
Following a description of the geographic, physiographic, and geologic setting of Detroit Seamount, including new and older seismic, and relevant drilling data, we explore two principal topics in this paper, the age of postshield volcanism at the summit of Detroit Seamount, and the depositional processes involved in forming its thick mantle of sedimentary deposits. We confirm the conclusion of Lonsdale et al. [1993] that the age difference between the shield-building lavas and the postshield cones on Detroit is far greater than the shield/postshield age differences observed on the Hawaiian Islands. We extend the time of postshield volcanism to at least the end of the Eocene. Second we corroborate the views of other authors [Scholl et al., 2003] that Detroit’s sediment cap is largely an Oligocene and younger drift deposit: the Meiji drift. The geometry and size of bedforms surprisingly imply that the depositing drift current was not particularly swift and that its vigor has decreased with time.

1.1. Geologic Setting of Detroit Seamount

Detroit Seamount lies at the northwestern end of the nearly 6,000-km-long Hawaiian-Emperor volcanic chain (Figure 1). The chain records the relative movement of the Pacific plate over the Hawaiian hot spot. The age of the volcanoes increases north and west of the island of Hawaii, the current locus of eruptive activity.

Detroit Seamount, a large plateau roughly the size of the Big Island of Hawaii, is centered at approximately 51.25°N, 167.5°E (Figure 2). The plateau has a broadly smooth surface disrupted by (1) troughs cut by ocean-bottom currents; (2) fault scarps; and (3) cones or domes that were subaqueously emplaced and never leveled at wave-base [Lonsdale et al., 1993] (Figure 2). Some of these cones, not covered by sediment, have been dredged [Lonsdale et al., 1993]. The summit structure and stratigraphy of Detroit Seamount generally consist of a relatively flat basement surface deeply buried (up to 840 m) by a channelized sediment drift deposit, the Meiji drift [Rea et al., 1993; Shipboard Scientific Party, 2002]. The Shipboard Scientific Party [2002] confirm that Detroit lavas erupted intermittently in shallow marine and nonmarine environments. No significant time hiatus has been found between the youngest shield lava flows and the overlying Late Cretaceous sediment at Sites 883 and 1204 [Rea et al., 1993; Shipboard Scientific Party, 2002], suggesting that, unlike the Hawaiian Islands, a large subaerial edifice did form above the cresting summit of Detroit Seamount.

Plate reconstructions predict that the Hawaiian hot spot was near a spreading ridge at ca. 80 Ma [Rea and Dixon, 1983; Mammerickx and Sharman, 1988], so Detroit Seamount formed on young (~10 m.y. old), therefore relatively warm and thin oceanic lithosphere. In contrast, the island of Hawaii is built on much older oceanic lithosphere (ca. 90–100 Ma) [Müller et al., 1997]. The resulting differences in subsidence rate have implications that are discussed later in the paper.

Other undated seamounts exist ~100 km northeast of the Emperor Seamount chain (Figure 1), but it is not clear if they are related to the Hawaiian hot spot. Lonsdale et al. [1993] suggests that these undated seamounts are likely no older than Cenozoic (early Tertiary) in age, on the basis of SeaBeam data showing that the summit of one of these seamounts was never wave-base leveled, and is less than 2 km below sea level.

1.2. New Seismic Data

Leg 197 data were collected using a single-channel streamer to provide high vertical and lateral resolution to assist detailed analysis of Detroit’s stratigraphic rock and sedimentary sequence. The 80-in³ SSI watergun used by the Resolution provides higher frequencies (spectral peak ~40–45 Hz) than conventional airguns. Slow ship speeds during the surveys enabled relatively closely spaced shot intervals that ranged between 15–20 m. The F-2-87-AA data recorded an 80-in³ conventional airgun source using two single-channel streamers at shot intervals between 40–50 m. The streamers were towed from opposite sides of the ship, and the active section of both streamers was offset 400 m behind the airgun. Leg 197 data were collected in two separate surveys, referred to as Survey 1203 (about ODP Site 1203) and Survey 1204 (about ODP Sites 883/1204). Lines 1203-1, 1203-3, and 1203-6 all intersect at Site 1203; lines 1204-1 and 1204-3 intersect at ODP Site 883, ~500 m northwest of Site 1204. The two USGS or F-2-87-AA lines (Lines F41 and F43) are subparallel, separated by ~25 km. Line F41 intersects Lines 1204-1 and 1204-3, and Line F43 passes within 1 km of Lines 1203-3 and 1203-4, allowing limited correlation with these profiles also (Figure 2).
1.2.1. Data Processing

To maximize our ability to correlate sediment horizons from line to line, we attempted to process both the Leg 197 and the F-2-87-AA data sets equivalently. The different source signatures of the Leg 197 and F-2-87-AA data sets (nonminimum phase watergun versus minimum phase airgun, respectively), and different frequency contents (Leg 197 peaks between 40–45 Hz and the F-2-87-AA data peaks at 22 and 44 Hz with a 30 Hz...
notch), limited the degree to which the processing sequences could be matched. We attempted a source-sig\nture deconvolution on both Leg 197 surveys (in order to convert the data to minimum phase) prior to predictive deconvolution. The lack of a true far-field source-sig\nture recording of the watergun source and a significant loss of signal-to-noise ratio due to source-sig\nture deconvolution artifacts, rendered source-sig\nture deconvolution impractical for Survey 1204 data. A persistent receiver ghost, prominent just below the seafloor in the Survey 1203 data, required a signature deconvolution for which we simulated the far-field source-re\n\nciver signature using the water-bottom reflecti\ng by stacking 40 traces from a thick (~800 ms twtt) part of the sediment cap. On all of the data sets, we performed normal moveout (NMO) corrections, predictive deconvolution, bandpass filtering, true amplitude recovery, and Stolt F-K migration. We used a constant velocity of 1,500 m/s for the NMO and the true amplitude corrections, as well as the migration. Additional processing details and data examples are presented by Kerr et al. [2005].

2. Analysis and Interpretation

2.1. Volcanic Edifices

[12] The shield stage of Hawaiian volcano evolution (e.g., Mauna Loa and Kilauea on Hawaii), during which the bulk of each island edifice is built by tholeiitic basalt flows, lasts for <1 m.y. with eruption intervals on the order of 1–10 years [Moore and Clague, 1992; DePaolo and Stolper, 1996; Clague, 1987]. After shield-building ends, alkaline basalt, hawaiite and trachyte lavas erupt on the shield summit over a period as long as ~0.2 m.y. with eruption intervals of $10^2$–$10^3$ years (e.g., Mauna Kea on Hawaii) [Moore and Clague, 1992; DePaolo and Stolper, 1996; Clague, 1987]. After a hiatus of ~0.25–2.5 m.y., a rejuvenated stage of alk\nkaline volcanism erupts melilite, nephelineite, basanite, and alkaline basalt over a duration as long as 3 m.y., with eruption intervals of $10^4$–$10^5$ years (e.g., Honolulu volcanics on Oahu) [Moore and Clague, 1992; DePaolo and Stolper, 1996; Clague, 1987; Langenheim and Clague, 1987]. Rejuvenated-stage lavas appear to be typical along the Hawaiian chain as they have been sampled from the older, more northwestern islands and seamounts (e.g., Necker Island, Ladd Bank, and Colahan Seamount) [Clague and Dalrymple, 1987]. In summary, within about 5 m.y. the documented volcanic history of each island in the Hawaiian chain is completed.

[13] We interpret the two peaks or highs in the basement topography imaged north of Site 883/1204 (Figure 3) to be volcanic cones akin to those documented by Lonsdale et al. [1993]. Rising approximately 500 m above the surrounding sub-horizontal basement, the edifices in Figure 3 are about one-fourth the height of the tallest cones mapped by Lonsdale et al. [1993]. Line 1204-1 probably does not cross directly over the top of the two buried edifices in Figure 3, so the edifices might actually peak more than 500 m above the surrounding basement. The apparent slope of the north flank of the northern edifice (approximately km 22 on Line 1204-1) is approximately 30°, and the apparent dip of the slopes of the smaller edifice (approximately km 18 on Line 1204-1) is about 10°.

[14] On the basis of the relation of sediment horizons with basement relief, these volcanic edifices formed before significant Meiji sediment accumulation began. At first glance, it appears that the edifices in Figure 3 are younger than the middle, lower, and pre-Meiji sequences as the surface of the edifices (Horizon A, Figure 3b) appears to truncate sediment horizons. If the edifices were older than the sediment, one would expect the sediment horizons, in response to particle rain, to drape or be concordant with the profile of the basement surface as the sediment rains down. However, as we establish later in this paper, an ocean-bottom current influences the deposition of the Meiji drift: Figures 7, 8, and 10 show that Meiji drift deposition is influenced by bottom-current interaction with existing topography, as sediment horizons onlap onto existing topography. As can be seen in Figure 3a (inset), sediment horizons onlap onto the edifices (Horizon A), suggesting the cones are older than the Meiji sedimentation, which began at ca. 34 Ma. The lack of evidence of sills in the sediment also implies the edifices are older than the Meiji drift.

[15] Using thermal subsidence curves from Detrick et al. [1977], Lonsdale et al. [1993] theorized that the volcanic cones that dot the summit of Detroit formed as late as 60 Ma. Lonsdale et al. [1993] presumed that the summit of Detroit was never significantly above wave-base, a presumption supported by the lack of a significant time hiatus between the youngest shield lava flows and the
overlying Cretaceous sediment at Sites 883 and 1204 [Rea et al., 1993; Shipboard Scientific Party, 2002]. Lonsdale et al. [1993] also assumed the cones never grew above wave-base, and that the subsidence rate of Detroit Seamount was not reset by this later volcanism (Figure 4). However, the cones could have formed later, in a deep marine environment. Rocks that Lonsdale et al. [1993] dredged from these cones geochemically resemble late-stage Hawaiian lavas, which therefore might be anticipated to have erupted much earlier than at 60 Ma [Lonsdale et al., 1993]. Along the Hawaiian chain, alkalic postshield stage lavas are typically erupted approximately 1 m.y. after shield building has ceased (up to 4 m.y. for the Honolulu volcanics) [Langenheim and Clague, 1987]. Thus, according to the Hawaiian model of eruption histories, the postshield cones on Detroit Seamount should have formed at ca. 75–71 Ma. However, we suggest below that the summit cones could be even younger than 60 Ma, so that the chemical similarity to late-stage Hawaiian lavas is perplexing, perhaps coincidental. Long time gaps between hot spot-related shield-building volcanism and non-hot-spot-related rejuvenescent volcanism have been reported elsewhere. Dredge samples from Hodgkins Seamount (northwest Pacific) exhibited age differences of 11.5 m.y. [Turner et al., 1980]. Keller et al. [1997] reported drilled basalt cores from the Patton-Murray seamount platform in the Gulf of Alaska with age differences as much as 16 m.y. Keller et al. [1997] attributed the origin of the youngest basalt cores to extension at a location of hot spot-induced plate weakness.

**2.1.1. Indications of Local Volcanic Activity**

[16] Volcanic material recovered during ODP Leg 197 drilling suggest that Detroit’s summit cones (including the edifices imaged on Line 1204-1) might be even younger than Lonsdale et al.’s [1993] 60 Ma estimate. Figure 5 shows the downhole locations of ash layers (as thick as 0.5 m) reported by Rea et al. [1993, 1995a, 1995b] and the Shipboard Scientific Party [2002]. These observations were made on the summit of Detroit (ODP Sites 883/1204) as well as on the flanks to the northeast of the summit platform (ODP Site 884) [Rea et al., 1993, 1995a, 1995b; Shipboard Scientific Party, 2002]. Ash layers occur throughout most of the Eocene section. The earliest Eocene ash layer was deposited at ca. 52 Ma (Site 1204). Glass intraclasts, ca. 48 Ma in age, were found at Site 883 [Rea et al., 1993]. Rea et al. [1993] describe these intraclasts as brown and angular suggesting they are basaltic and were not transported over a long enough distance to round the clasts. On the summit at Site 883, black ash layers occur as late as 32 Ma [Rea et al., 1993]. At Site 884 the ash-rich layers are several meters thick, occurring throughout the Eocene [Rea et al., 1993]. Site 884 is 50 km downslope from Sites 883 and 1204.
Sites 883 and 1204 (~500 m apart) are within 7–11 km of the two summit cones observed on Line 1204-1 of our Leg 197 data. Ash layers recovered from Sites 883 and 1204 that are vitric in texture are also dark in color, suggesting a mafic composition [Rea et al., 1993, 1995a, 1995b; Shipboard Scientific Party, 2002]. Other ash layers have been altered to palagonite and are brown in color, perhaps indicating submarine eruption. In addition, the Eocene vitric ash layers observed at Sites 883 and 884, by Rea et al. [1993, 1995a, 1995b] and at Site 1204 by the Shipboard Scientific Party [2002] are absent at Site 1203 (30 km to the south) [Shipboard Scientific Party, 2002] implying that they were derived locally. Hence it seems likely that these and other summit cones on Detroit are the source of the Eocene ashes cored at Sites 883, 884, and 1204.

The numerous observations of displaced and deformed sediment at Sites 883, 884, and 1204 throughout the Eocene (Figure 5), suggest slope instability induced by rapid deposition of volcanic material, as well as the uplift of that material caused by igneous intrusion associated with the volcanism. Deformed or sediment-mobilized structures include convoluted and contorted bedding, slump blocks and folds, microfaults, load casts, and scour surfaces [Rea et al., 1993]. Disrupted sediment is observed at Site 883 at ca. 45, 47–48, and 51 Ma, some of which are coincident with the glass intraclasts mentioned above [Rea et al., 1993].

Geochemical analysis by Duncan and Keller [2004] reveals alteration of feldspars to sanidine in Site 1204 basalt beginning at ca. 58 Ma and ending at ca. 33 Ma. Duncan and Keller [2004] attribute this age resetting to circulating K-rich fluids produced during postemplacement volcanic activity [see also Dallymple et al., 1980]. We propose that the Eocene ashes, reworked sediment, and chemically altered basement rocks observed at Sites 883,
884 and 1204 are due to a renewed phase of volcanism occurring on Detroit Seamount throughout the Eocene, and possibly into the early Oligocene. Such a young age is consistent with the onlapping of Unit III onto the volcanic edifice, and its eventual overtopping by Unit II, seen on Line 1204-1 (Figure 3b).

[20] Even though we believe the nature of the ash layers implies a local source for these volcanics, the apparent Eocene volcanism on Detroit Seamount might have been part of a Pacific Basin-wide event. During the early and middle Eocene, the present crust of the northwestern Pacific Basin was affected by plate reorganization tectonism as the Kula-Pacific plate boundary immediately east of Detroit rotated by nearly 45° [Lonsdale, 1988]. The Aleutian island arc also formed at this time [Scholl et al., 1987]. It is possible that Detroit Seamount’s proximity to the rotating Pacific-Kula plate boundary changed the regional stress regime such that melt was able to work its way to the surface. We can rule out extension linked to formation of the flexural arch around the islands centered over the active hot spot as a cause of volcanism because by 52 Ma, the Hawaiian hot spot was ~1500 away, between Nintoku and Koko Seamounts (Figure 1) [Shipboard Scientific Party, 2002]. Perhaps significantly, volcanism occurred at an increased rate during the Eocene throughout the Pacific Basin [Kennett et al., 1977]. Seismic activity, associated with increased rates of volcanism, might explain the widespread recording of sediment deformation and displacement [e.g., Schlanger and Premoli Silva, 1981; Thiede, 1981a, 1981b; Lancelot et al., 1990; Storms et al., 1991; Rea et al., 1995a, 1995b].

[21] More work could be done to determine if the summit cones documented by Lonsdale et al. [1993] and imaged along Line 1204-1 on Leg 197 are related to the Eocene ashes, displaced and deformed sediment, and altered basement rock. Geochemical analysis needs to be performed on the ashes from Sites 883, 884, and 1204 to constrain their origin. Specifically, matching the geochemical signatures of the Eocene ashes with signatures measured from Lonsdale et al.’s [1993] rock...
dredges on Detroit Seamount would help confirm our suggestion that the Eocene ashes record localized volcanic activity on its summit.

2.2. Meiji Drift

2.2.1. Sediment Drifts

[22] Sediment drift deposits record important information about global ocean currents such as location, speed, and direction [Dorn and Werner, 1993; Flood, 1994; Manley and Caress, 1994; Cunningham and Barker, 1996; Howe, 1996; McCave and Carter, 1997]. The Meiji sediment drift, up to 1800 m thick, over 1000 km long, and ~350 km wide, has been accumulating since the early Oligocene beneath a southeasterly flowing bottom current in the northwest Pacific [Scholl et al., 1997, 2003; Mammerickx, 1985; Rea et al., 1995a, 1995b]. Thermohaline circulation (THC) drives the current responsible for the Meiji sediment drift [Scholl et al., 2003]. Prior to the late Cenozoic, it is unclear whether the THC responsible for the Meiji sediment drift was generated locally (Bering Sea), or distantly (Pacific Bottom Water), or is a combination of the two source regions [Scholl et al., 2003]. Surface water salinity is currently too low to drive THC in the Bering Sea Basin, hence bottom flow (which has not been instrumentally confirmed) over the Meiji drift is presumably generated distantly by THC in the north Atlantic and peripheral to Antarctica. The rare, but consistent, presence of Arctic-boreal diatoms indicate that at least some of the Meiji drift was sourced from a Bering Sea connection [Barron and Gladenkov, 1995]. Changes in Meiji drift sedimentation rate may correlate with changes in delivery rate and sources of deep waters in response to large-scale shifts in abyssal circulation caused by tectonism and changes in the global climatic regime. Hence our seismic profiles that image the total thickness of the Meiji drift have the potential to date episodes of climate change.

[23] Sediment drift deposits are primarily characterized by wave-like bedforms often referred to as mudwaves, and the Meiji drift exhibits these in abundance. Wynn and Stow [2002] provide a comprehensive summary of the characteristics (wavelength, wave height, wave symmetry, etc.) of mudwaves. 

[24] Horizontal migration of fine-grained mudwave crests and troughs is controlled by the velocity of the bottom current [Wynn and Stow, 2002]. Flood [1988] explains this wave migration using a lee-wave model (Figure 6), in which mudwaves horizontally migrate upcurrent and upslope when bottom-current flow (assumed to be orthogonal to the wave front) velocities are approximately 9–30 cm/s (0.3–1.1 km/hr). At slower bottom-current velocities, sediment tends to accumulate in the wave troughs [Flood, 1988]. If the mean bottom-current velocity is greater than ~16 cm/s (0.6 km/hr), erosion will begin to occur on the downcurrent flank of the mudwave. As the mean bottom-current velocity increases, the zone of erosion increases along the downcurrent flank of the mudwave until sediment accumulates only on the upcurrent flank, so that migration occurs without aggradation [Flood, 1988]. In other words, the wave crests will horizontally migrate, but they will not become taller, even though the sediment supply remains constant. When the mean bottom-current velocity exceeds approximately 30 cm/s (1.1 km/hr), erosion occurs across the entire mudwave. The minimum velocity necessary for migration of mudwaves increases with wavelength (due to drag) and latitude (due to Coriolis effects). The velocity at which erosion begins to occur on the downcurrent flank of the mudwave increases as wave height and length decrease. The velocity at which complete erosion occurs (the maximum velocity at which wave migration is possible) is dependent only on mudwave wavelength [Flood, 1988].

[25] The structure of sediment bedforms proximal to channels is a key paleocurrent indicator. The long travel distances, on the order 104 km, of major ocean-bottom currents make them subject to Coriolis effects, which in the northern hemisphere deflect current flow to the right of the flow direction. For example, a northern-hemisphere bottom current flowing along a southeast-trending channel will have a spillover component above the top of the southwest side of the channel [Normark et al., 2002]. Mudwaves formed on the southwestern bank will migrate against the spillover current toward the channel (Figure 6) [Flood, 1988; Normark et al., 2002].

2.2.2. Meiji Drift Lithostratigraphy on Detroit Seamount

[26] The lithostratigraphy of Detroit Seamount was characterized at Site 883 (Figure 2) by Rea et al. [1993]. The 840-m-thick sediment cap here is divided by Rea et al. [1993] into 5 units on the basis of physical properties measurements, diatom abundance, distinct color changes, and presence of sedimentary structures (e.g., scoured surfaces, lam-
Figure 7
inations, and bored hardgrounds) (see insets to Figures 3b, 7b, and 8b as well as the annotated well log inserted in seismic section of Figure 3b). Unit I (~87 m thick at Site 883, late Pliocene to Quaternary in age) consists of clay with quartz and diatoms. Unit II (~371 m thick, late Miocene to late Pliocene in age) consists of diatom ooze. Vitric ash layers are present in both Unit I and Unit II. Unit III (~194 m thick, early Miocene to late Miocene in age), devoid of distinct volcanic ash layers, is composed of calcareous diatomaceous ooze with interbeds of calcareous chalk. Unit IV (~162 m thick, early Eocene to late Oligocene in age) consists of nannofossil chalk with increasing quantities of altered ash downhole. Altered ash and claystone comprise the bulk of Unit V (~25 m thick, late Cretaceous to Paleocene in age). Basalt underlies Unit V.

[27] We have picked 6 prominent and laterally continuous seismic reflection horizons (Horizons A–F) to help characterize temporal evolution of the seismic stratigraphy along Surveys 1203 and 1204 (insets to Figures 3b, 7b, and 8b). We estimate the ages of these horizons using the sonic logging data from Site 883 (Leg 145) (Figure 3b) to convert two-way travel time to a horizon to that horizon’s depth of recovered core [Rea et al., 1993]. We use the depths of the biostratigraphic datum levels from Barron et al. [1995] to get the age of each seismic horizon. Horizon A marks the contact between volcanic basement and the overlying sediment (base of Unit V). Horizon B (within Unit IV) is the contact between sediment that is late Eocene and older in age, and the Meiji drift sediment of Oligocene and younger age. Horizon C marks the top of highly reflective late Miocene and older Meiji sediment (approx. top of Unit III). Horizon C is ca. 6.2 Ma, and is prominent across the summit of Detroit Seamount on the Leg 197 data (Figures 3, 7, and 8).

[28] Horizon D (5.1–5.3 Ma) is overlain by sediment horizons that downlap on to it at the northeastern ends of Lines 1203-1 and 1203-6 (inset B in Figure 7a, inset in Figure 8a). Horizon D’s high amplitude enables identification of this horizon on Line 1204-1 (Figure 3) where overlying horizons are concordant. Horizon E (within Unit II) is another strong reflector (3.5–3.8 Ma) on which overlying sediment horizons downlap at the north-eastern ends of Lines 1203-1 and 1203-6 (inset B in Figure 7a, inset in Figure 8a). Horizon F is another strong reflector (ca. 1 Ma) near the top of the Meiji drift.

2.2.3. Meiji Drift Imaged on Lines 1203 and 1204

[29] From Figure 7a, it is clear that Horizon B marks a significant erosional event. Pre-Meiji (Paleocene to Eocene) horizons terminate upward into Horizon B between km 34 and km 38. Nannofossil stratigraphy indicates a lower Oligocene to upper Eocene hiatus at Site 883 that likely coincides with Horizon B at Site 1203 [Barron et al., 1995]. The pre-Meiji sediment thickens southward away from Site 1203 (Figures 7b and 8b).

[30] In the vicinity of Site 1203, sediment thins (between Horizons D and E, and Horizons E and F) and horizons downlap (onto Horizons D and E) to the northeast (inset B in Figure 7a, inset in Figure 8a). However, 5 km southwest of Site 1203 (km 38 on Figure 7b, and km 135 Figure 8b), sediment thickness is relatively constant and downlaps are absent. Thinning and downlapping sediment horizons near Site 1203 indicate lower rates of sediment deposition relative to 5 km southwest of Site 1203, indicating greater bottom-current velocities near Site 1203 than 5 km to the southwest.

[31] Approximately 15 km southwest of Site 1203 on lines 1203-1 (between km 24 and km 31, Figure 7) and 1203-6 (between km 122 and km 127, Figure 8), the early Meiji sediment between Horizons B and C onlaps onto the pre-Meiji sediment and basement (Figures 7b and 8b). This is evidence of Meiji sediment horizons prograding upslope against the spillover current from the channel-levee-like system created by a normal fault discussed later (at km 32 on 1203-1 and km 127 on 1203-6).

[32] Figures 7 and 8 show two profiles from Survey 1203 that cross a prominent channel or trough.
controlled by a normal fault (here named Summit Fault; described in the next section) in the basement ~12 km south of Site 1203. Figure 2 shows that the channel and Summit Fault trend northwest-southeast, and are crossed by Lines 1203-1, -6 and -4. Mudwaves are prominent on the southwest side of the channel (upthrown Summit Fault block) in Figure 7 along Line 1203-1.
between km 23 and km 31, and between 3.4 s and 4.1 s two-way travel time (twtt). The crests of the mudwaves migrate upslope (to the northeast) toward the channel. Assuming a constant seismic velocity of 1600 m/s, the amplitudes of mudwave horizons decrease over time. At Horizon C, mudwave amplitudes (half of the peak-to-trough heights) are ~35 m. However, in Figure 7b, mudwave amplitudes at Horizon D are ~22 m, and ~10 m at Horizon E, suggesting a decrease in bottom-current flow energy over time [Normark et al., 1980; Carter et al., 1990]. The likely increase in seismic velocity due to compaction and dewetting is much less than a factor of two (1600 m/s to 3200 m/s), yet the apparent mudwave amplitudes decrease by a factor of 3.5 (40 ms to 12 ms), implying that the real amplitudes also decrease (35 m to <20 m). The section between the horizons C and D is ~100 m thick (assuming a \( V_p = 1600 \text{ m/s} \)), and represents approximately 1 m.y. of deposition. Within this section, wave crests horizontally prograde approximately 250 m upslope, giving a horizontal migration rate of ~0.25 m/ka (250 m/m.y.). This rate is the same order of magnitude as rates recorded in the north Atlantic on the Feni Ridge drift body [Lonsdale and Hollister, 1979], and in nearby Rockall Trough (0.4–0.9 m/ka) [Masson et al., 2002]. Along Line 1203-1 on the summit of Detroit Seamount mudwave wavelengths are approximately 2.7 km, and wavelengths appear to be relatively constant during the late Miocene (despite the 35% decrease in wave amplitude).

[33] The velocity of the bottom current responsible for the mudwaves between Horizons C and D (ca. 6.2–5.2 Ma) was quite slow. Using Flood’s [1988] lee-wave model, the 2.7 km wavelength of the mudwaves between Horizons C and D predicts the mudwave crests would migrate in response to bottom-current velocities between 4–21 cm/s (0.14–0.76 km/hr). Given the range of mudwave amplitudes between Horizons C and D (35 m and 22 m, respectively), erosion of the downcurrent flank of the mudwaves would occur at velocities of 16 cm/s (0.58 km/hr) and 17 cm/s (0.61 km/hr), respectively. Because sediment accumulation is greater in the mudwave troughs than the crests between 6.2 and 5.2 Ma (shown by the decrease in mudwave amplitude from 35 m to 22 m), the velocity of the current across these mudwaves was approximately 4 cm/s (0.14 km/hr). For comparison, Blumsack and Weatherly [1989] recorded a bottom-current velocity of ~10 cm/s (0.36 km/hr) orthogonal to mudwaves in the central

[34] On the southwestern flank of the channel (upthrown side of Summit Fault), marker horizons show clear mudwaves. In contrast, on the northeastern or downthrown side of the channel, horizons are generally flat except on Line 1203-1 where the northeast bank slumps into the channel (Figure 7). This relation implies much lower flow velocity across the northeastern flank of the channel (parallel to Lines 1203-1 and 1203-6), as we would expect to observe mudwaves recording any component of southwest/northeast bottom-current flow. Thus we interpret that the mudwaves on the southwestern side of the channel result from (1) southeasterly channel flow spilling over a

Figure 9. Conceptual diagram illustrating bottom current interaction with Detroit Seamount and Summit Fault. Regional flow represents the main bottom current responsible for the Miji drift. Summit flow is the bottom current flow directed over the summit of Detroit Seamount due to Coriolis forces. Channel flow represents a component of summit flow entrapped by Summit Fault’s scarp. Spillover flow results from the Coriolis effect forcing channel flow over Summit Fault’s scarp. Location of the mudwaves observed in Figures 7 and 8 is shown relative to Summit Fault. Climbing bedforms are seen in Figure 10.
levee, similar to those described by Normark et al. [2002] and by Lewis and Pantin [2002], formed above the upthrown block of a normal fault (Summit Fault), and (2) southwesterly flow over the summit of Detroit Seamount. The velocity of spillover flow over the mudwaves was much greater than the velocity of the summit flow. We infer the relatively slow summit flow velocity because of the lack of mudwaves northeast of Summit Fault. The combined flow along the mudwaves is likely low velocity relative to other mudwave-producing bottom currents as horizontal migration (~250 m) of the mudwaves is roughly 2.5 times vertical accretion (~100 m). In other words, for every meter of vertical accretion, the wave crests have migrated only 2.5 m in the horizontal direction, as shown in Figure 6. This is less than the factor of 5 that Normark et al. [2002] uses to distinguish between vertical wave aggradation (<5) and horizontal migration (>5) of mudwaves in turbidite fan systems.

[35] Figure 9 schematically illustrates bottom-current flow in the vicinity of Detroit Seamount. The Coriolis effect on the regional bottom current produces flow over the summit of Detroit Seamount. Sediment onlapping on the northeast flank of Detroit Seamount basement in Figure 10 is consistent with a southerly-flowing bottom current as sediment is transported up and over the plateau. Otherwise, the lack of a bottom current would result in sediment horizons concordant with the basement. Near-vertical aggradation of the climbing bedforms suggests low current velocities. Summit Fault’s scarp diverts a component of summit flow into southeasterly channel flow along the scarp, and the Coriolis effect on the channel flow creates spillover flow over the scarp (Figure 9).

[36] In summary, the mudwaves observed on our profiles are entirely characteristic of deep-water drifts. The velocity of the constructive current of the Meiji drift (e.g., the summit flow) is not swift enough to generate mudwaves unless the flow interacts with seafloor topography. Hence we are unable to measure the velocity of the regional flow over the summit of Detroit Seamount. So we are predominantly characterizing spillover flows rather than primary flows. However, we can infer that the regional summit flow is significantly slower than the ~4 cm/s velocity measured between Horizons C and D (ca. 5.7 Ma). The velocity of the Meiji current and its evolution with time would best be
determined by new, dedicated seismic profiles recorded parallel to the current in the deep ocean, away from perturbing seamounts, and extending for 10–100 km. In principle, properly oriented and spaced profiles could recognize current velocity changes potentially linked to changes in sources and vigor of bottom water flow with north Pacific or global paleoclimatic implications.

2.2.4. Summit Fault

[37] The normal fault imaged during Survey 1203 (here named Summit Fault) is located ~12 km southwest of Site 1203. The fault trends about 300° (see Figure 2) and dips northeast with apparent dips ranging from 22° in the southeast (Line 1203-4) to 16° in the northwest (Line 1203-6). We calculate apparent fault dip, heave, and throw using top-basement (Horizon A) offsets on all three lines, and converting time to depth using \( V_{\text{sediment}} = 1,600 \text{ m/s} \) and \( V_{\text{water}} = 1,500 \text{ m/s} \). The true dip of the fault, 25°, is likely nearly the same as the apparent dip because all three survey lines are oriented approximately normal to the trend of the fault (Figure 2) (see true-scale image Figure 7c). The apparent vertical offset (throw) of the base- ment increases from ~300 m in the southeast (Line 1203-4) to ~500 m in the northwest (Line 1203-6). On 1201-1, the throws are: 450 m at Horizon A; 300 m at Horizon B; and <50 m on Horizons C, D, E, and F. Smaller normal faults occur northeast of Summit Fault in the downthrown block (Figure 7b). These smaller faults likely formed coincidently with Summit Fault. Because Summit Fault is nonplanar (fault dip varies with apparent erosion of pre-Meiji sediment by bottom current. In order to control bottom-current flow, and hence mudwave formation (as previously discussed), Summit Fault’s scarp must have existed before deposition of the Meiji drift. Because of the apparent erosion of pre-Meiji sediment by bottom currents, and because of the rapidly varying thickness of Meiji deposition, we cannot use the apparent throw of Horizons A to F to more accurately constrain the time interval of Summit Fault’s activity. However, because all sedimentary sequences exhibit a change in thickness across Summit Fault, we recognize that much its vertical offset could have formed in the latest Cretaceous.

[38] Summit Fault was most active sometime between 76 and 34 Ma. The older age limit assumes that the fault is not older than the age of the shield. The younger age limit is the onset of the Meiji bottom current. In order to control bottom-current flow, and hence mudwave formation (as previously discussed), Summit Fault’s scarp must have existed before deposition of the Meiji drift. Because of the apparent erosion of pre-Meiji sediment by bottom currents, and because of the rapidly varying thickness of Meiji deposition, we cannot use the apparent throw of Horizons A to F to more accurately constrain the time interval of Summit Fault’s activity. However, because all sedimentary sequences exhibit a change in thickness across Summit Fault, we recognize that much its vertical offset could have formed in the latest Cretaceous.

[39] We interpret Summit Fault as recording northeast-southwest extension of the southeastern sector of the summit plateau, possibly a manifestation of slope failure. The fault can not be traced across Lines F41 or F43 from cruise F-2-87-AA, so it is likely a plateau-edge fracture, not a result of regional tectonism. Nonetheless, as with the volcanic cones, Summit Fault suggest a record of geological instability long after the Detroit shield volcano would normally have been assumed to be quiescent.

3. Conclusions

[40] It appears Detroit Seamount was the site of episodes of postshield volcanism that occurred much later than is typical for the Hawaiian Islands. Volcanic activity occurring on Detroit Seamount 24–42 m.y. after the end of the main shield-building phase is considerably later than the late-stage volcanism occurring only a few million years after shield-forming ceased on the Hawaiian Islands. Renewed volcanism on Detroit might be related to changes in the motions of the Kula plate during the Eocene. It is interesting and important that the geochemistry of the late-stage Detroit Seamount volcanism matches that of the much shorter-hiatus alkalic rejuvenated stage of Hawaiian volcanism.

[41] By analyzing the relation of strata and mudwave geometries with preexisting bathymetric relief, we provide confirming evidence that the Meiji sediment body is a drift accumulation deposited by an ocean-bottom current flowing southeast along the northeast-facing flank of the Emperor Seamount chain. The estimated bottom current velocity of ~4 cm/s (0.14 km/hr) represents the sum of (1) the relatively slow regional flow over summit and (2) the considerably faster Summit Fault scarp spillover current. We infer that the regional bottom current velocity over the summit of Detroit Seamount is significantly smaller than some other inferred bottom currents. Future seismic profiling efforts in this region might usefully target time variation of the regional current velocity as a proxy for bottom-water composition and hence paleoclimatic events.

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