GEOLOGY, MORPHOLOGY AND ACOUSTIC SIDESCAN (SeaMARC II)

CHARACTERISTICS OF PACIFIC SEAMOUNTS:
ORIGIN AND EVOLUTION OF LANDFORMS AND
REMOTE SENSING GROUND TRUTH

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This thesis is dedicated to
Christopher John Bellaire Zayac

The memory of my brother, Chris, was the source of my strength and persistence in completing my degree after having been ready to give up. He was gifted with a sharp and surprising wit that could turn my blackest mood into a laughing fit, and I would instantly feel better. He helped me learn some important lessons of life: that things are never quite as bad as they seem and that even the most tragic situation can be a source of great inspiration if you choose to make it so.

I found the following quote by Ralph Waldo Emerson hung in Chris' room. It was obviously important to him.

"To laugh often and love much; to win the respect of intelligent persons and the affection of children; to earn the approbation of honest citizens and endure the betrayal of false friends; to appreciate beauty; to find the best in others; to give of one's self; to leave the world a bit better, whether by a healthy child, a garden patch or a redeemed social condition; to have played and laughed with enthusiasm and sung with exultation; to know even one life has breathed easier because you have lived - this is to have succeeded."

Chris, I dedicate this work to you, your garden patch and where it is forever Spring.
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CHAPTER 1
INTRODUCTION

This thesis consists of two related studies which are presented in the following chapters. Chapter 2 serves as a general background source of information about elevations on the seafloor, known as seamounts. It provides a synthesis of information on how environmental and geological processes affect the morphology of seamounts. The chapter is divided into specific subsections pertaining to the process being reviewed.

Chapter 3 makes up the second major portion of this thesis. This is a study to determine the effectiveness of SeaMARC II, a long-range, high-resolution seafloor mapping system, as a geologic mapping tool. The ability of SeaMARC II to distinguish different lithologies was ascertained by correlating ground truth data from ALVIN dive observations, ANGUS towed camera films and rock samples with features derived from the backscattered energy used to produce SeaMARC II images. Chapter 3 describes the methodology and results obtained from the data from Seamount 6, a well-surveyed seamount located near the East Pacific Rise at about 12° N and 102° W.
CHAPTER 2
PROCESSES AFFECTING SEAMOUNT MORPHOLOGY

INTRODUCTION

The morphology of seamounts is a function of a variety of processes. Some are primary controls, such as volcanic processes, tectonics and bathymetry. Other controls are secondary, affecting the shape after the main edifice has formed. These controls include sedimentation and sediment transport, current action, erosion and mass wasting and reef cap growth. In this paper, I will examine how these processes influence the morphology of seamounts.

Guyots, atolls and volcanic islands are included in this discussion since they are subject to the same controls. In general, terms are defined as follows. Seamounts are local, steep-sided bathymetric highs. For our purposes, we are only interested in structures formed by volcanism. Not all seamounts have a volcanic origin; some have formed by serpentinite diapirism in the Mariana fore-arc [Fryer and Fryer, 1987]. Guyots are drowned, ancient islands with flat tops. The flat summits are considered to be the result of wave truncation. Atolls are volcanic islands which have subsided or are subsiding, and are capped with coral reef crowns. Volcanic islands are, of course, edifices which rise above sea level.
Volcanoes in the south and west Pacific Ocean are found predominantly in large, linear island and seamount chains (Figure 1). Most likely, these are of hotspot origin [Batiza, 1989]. In the east Pacific, most seamounts are clustered and isolated (Figures 2 and 3) [Menard, 1964]. Most of these seamounts are probably not hotspot related but are linked to mid-ocean ridge volcanism [Batiza, 1989]. Non-linear volcanoes form where conditions for melt generation and eruption exist. Some forming near the East Pacific Rise are fed by the ridge [Hekinian et al., 1983]. Others result from excess volcanism at ridge crest offsets. Batiza and Vanko [1985] showed seamounts may be produced at the intersections of failed spreading centers and at failed transform faults. It was shown by Batiza et al. [1989a] that non-hotspot volcanoes in the Atlantic Ocean and near the Mid-Atlantic Ridge (Figure 4) are similar in distribution, abundance, morphology and origin to those in the Pacific Ocean.

SIZES AND SHAPES

Seamounts come in a wide range of sizes and shapes. Batiza and Vanko [1983] and Fornari et al. [1987] described the morphology with increasing size, as follows: 1) tiny conical and dome-shaped lava piles (Figure 5), 2) irregular or polygonal shapes in seamounts up to 1 km high or less (Figure 6), 3) circular overturned soupbowl shapes, and 4)
FIGURE 1. Linear island chains in the Pacific from Clague and Jarrard [1973].
FIGURE 2. Map of the Pacific from Udintsev et al. [1976] showing the location of known volcanoes. The sizes of the dots indicate relative volcano volume.
FIGURE 3. Bathymetric map of a typical cluster of seamounts generated near the East Pacific Rise. Notice the characteristic circular shapes, steep outer flanks and the broad, flattish summit regions.
FIGURE 4. The locations of seamounts in the North Atlantic with heights greater than 50 fathoms (shown as dots). Magnetic lineations are from S. Cande (Pers. Comm. to Epp and Smoot). The dashed line shows the location of the topographic high between the Charlie-Gibbs fracture zone and the East Azores fracture zone; it does not correspond to anomaly zero [from Epp and Smoot, 1989].
FIGURE 5. Bathymetric map of a small, conical seamount generated near the East Pacific Rise. In the early stages of its evolutionary growth, it displays the common morphology of many young seamounts.
FIGURE 6. This irregularly shaped "H" seamount may have formed by erupting from parallel fractures. Note the western edge of the "H" is much more developed than the eastern edge, which resembles an earlier evolutionary stage—the conical lava pile. The location is near the East Pacific Rise.
larger seamounts are either simply conical edifices or rift-dominated starfish shapes (Figure 7) [Vogt and Smoot, 1984]. Depending on the size of the seamount, the following features may occur: calderas or summit craters (Figure 8), broad flat summit regions (Figure 9), arcuate summit benches (Figure 10), surrounding satellite cones (Figure 11) and small lateral rift zones [Lonsdale and Spiess, 1979; Batiza et al., 1984; Fornari et al., 1984; Lonsdale and Batiza, 1980]. For instance, medium sized seamounts (1-2 km high) commonly have calderas and craters on their flat summits (Figure 12), and in many cases these are nested and/or breached [Fornari et al., 1987]; flank rift zones are usually absent or poorly developed. Oceanic volcanoes generally retain steep (18°-20°) average slopes until they reach volumes of about 5000 km³ [Vogt and Smoot, 1984]. Beyond this, slopes tend to diminish toward the low values of large shields (8°-10°). Figure 13 shows this decrease in slope angle with an increase in volume.

The smallest Pacific seamounts (50-300 m high cones) tend to be roughly circular or elliptical with steep outer flanks and usually have flat tops (Figure 14) [Fornari et al., 1987; Lonsdale, 1983]. Seamounts generated near the East Pacific Rise (EPR) (Figure 15) are generally < 1-2 km high. Larger volcanoes (up to 1-2 km high) have more complex shapes and cratered summits (Figure 16). The summit calderas (i.e., diameters > 1 km) or central craters are typically several kilometers wide and several hundred
FIGURE 7. Bathymetric chart from Vogt and Smoot [1984] of the Winterer cluster of guyots and seamounts in the western Pacific. Note the distinctive starfish shapes resulting either from rift zone formation or mass wasting.
FIGURE 8. Bathymetric map of two seamounts generated near the East Pacific Rise (EPR). The edifice to the east possesses a summit crater, while the one to the west has a small pit crater. Note also the typical steep outer flanks and relatively flat summit regions of near-EPR seamounts.
FIGURE 9. Bathymetry of small cluster of seamounts formed near the East Pacific Rise, displaying characteristic morphology of such volcanoes—steep outer flanks and broad, flattish summits. Notice two summit craters on the central edifice and the presence of several satellite cones.
FIGURE 10. The central and eastern seamounts in this bathymetric map display arcuate summit benches. Evident on the easternmost seamount is a large caldera. Also notable in these near-EPR seamounts, are the circular forms, steep outer flanks and flattish summit regions.
FIGURE 11. This bathymetric map shows many small satellite cones present about these seamounts located near the East Pacific Rise.
FIGURE 12. Bathymetric map of OCP ridge. The large, rectangular-shaped seamount in this bathymetric map possesses not only a large summit crater, but also an arcuate summit bench. The steep outer flanks, plus the relatively flat summits typical of seamounts generated near the East Pacific Rise are also shown.
FIGURE 14a. This bathymetric map is of a lava field just east of the intersection of the Siqueiros transform and the East Pacific Rise (EPR), displaying the characteristic morphology of small Pacific seamounts—circular or elliptical with steep outer flanks and, usually, flat tops.
FIGURE 14b. This bathymetric map displays characteristic morphology of an area east of the intersection of the Clipperton transform and the East Pacific Rise. See caption for Figure 14a.
FIGURE 15. Bathymetry of typical medium-sized East Pacific seamounts. Note the steep outer flanks and relatively flat summits. The summit of the easternmost seamount is cratered. Also notable are the beginnings of what seem to be northwest trending rift zones on the central and eastern volcanoes.
FIGURE 16. Bathymetry of a large seamount near the East Pacific Rise which may actually be two coalesced volcanoes, accounting for its polygonal shape. Many characteristic morphological features are in evidence here: steep outer flanks, a relatively flat, broad summit region, summit craters, arcuate summit benches and satellite cones.
meters deep with steep walls and irregular floors [Batiza et al., 1984]. Within the calderas, pillow mounds, talus piles and pit craters are seen (Figure 17) [Batiza et al., 1984]. Pit craters are about a hundred meters wide and deep and bounded by fault scarps [Lonsdale and Spiess, 1979]. These larger seamounts were formed from a number of eruptions and have longer and more involved histories [Batiza and Vanko, 1984; Fornari et al., 1988; Fornari et al., 1987]. Small seamounts are present in the inner rift valley and on the flanks of the MAR (mid-Atlantic Ridge), in the FAMOUS area [Walker et al., 1984; Phillips and Fleming, 1978; Bryan and Moore, 1977], and in other parts of the Atlantic. Indeed, Batiza et al. [1989a] have shown that MAR and EPR near-ridge seamounts display similar shapes, sizes, distribution, abundances and chemistry.

This seems to indicate that the production of seamounts is a normal mid-ocean ridge process independent of spreading rate.

Small seamounts are much more abundant than large ones in the Pacific (Figure 2) and the Atlantic Oceans, but the total volume of smaller seamounts is less than that of the larger ones. In the Pacific, seamount abundance increases with crustal age on crust of Eocene and younger age. On lithosphere older than Eocene age, the size of seamounts increases while the abundance decreases with increasing crustal age [Batiza, 1982]. This trend is also seen in small Atlantic seamounts, the abundance falling off with
FIGURE 17. Bathymetry of the summit area of Seamount F [from Batiza et al., 1984], showing some of the typical morphologic features of seamount summits. Northwest to southeast profile at twice vertical exaggeration is constructed from the bottom pinger trace of the camera sled.
increasing lithospheric age [Batiza et al., 1989a]. In the Pacific, at least, the decrease in the number of small seamounts is attributed to sediment burial on the older lithosphere [Batiza, 1982]. Also, the increase in larger volcanoes is probably due to the affect of increasing lithospheric thickness on the maximum size volcanoes may attain [Vogt, 1974b].

PRIMARY VOLCANIC PROCESSES

The primary morphology of volcanoes depends on several factors, including morphology of magma conduits, eruption rate, magma viscosity, previous topography and regional and gravitational stresses within the edifice. Fractures, faults and joints act as conduits, providing preferred paths for the transport of magma. Eruption rate is related to magma availability and magma viscosity which are major controls in producing various lava flow forms. Pre-existing topography also contributes to the morphology of lava flows. Regional and gravitational stresses are important in controlling the shape of the volcano.

Vogt [1974b] suggested that the size of seamounts is controlled by lithospheric thickness, since the largest volcanoes seem to be on the oldest, thickest lithosphere. Nakamura [1977] and Kear [1964] showed how vent alignment and elongate shapes of volcanoes reflect regional tectonic stresses and fault patterns. Fiske and Jackson [1972]
demonstrated that gravitational stress within a volcanic edifice can localize eruptions along linear zones of maximum tension. This model explains linear volcanic rift zones. Simkin and Batiza [1984] used Fiske and Jackson's model to explain flat-topped seamounts. They suggested that a doughnut shape of the edifice is established by cratering, and then gravitational stresses maintain volcanic conduits that are circular in plan view. Extrusions of lava from this ring may then flow into the crater, eventually filling it and forming a flat top (Figure 18).

Fornari et al. [1987] argue that the shape, size and arrangement of magma supply conduits is a very important control on the shape of volcanic landforms. For instance, if the magma conduit is a point source, conical lava piles will result; if it is a fissure, a ridge-like feature will result. They go on to say that the irregular or polygonal shaped seamounts owe their shape mainly to the shape and geometry of these conduits. Irregular shaped seamounts are seen as primary growth forms, probably fed by a series of conduits which are controlled by sets of intersecting seafloor faults and fractures that form the polygonal outline [Batiza and Vanko, 1983]. Examples of possible conduit patterns are shown in Figure 19.

As seamounts grow, their shapes probably evolve. It is possible that the large, circular volcanoes (1-2 km high) begin as irregularly shaped seamounts. Eruption
FIGURE 18. Schematic figure [from Simkin, 1972] illustrating how a flat top may develop because of extrusion from ring-shaped conduits.
FIGURE 19. Examples of inferred conduit shapes and patterns for small volcanoes of given plan form and cross-sectional shape [from Batiza and Vanko, 1984]. The composite fracture pattern is reconstructed from the inferred conduit geometry of individual volcanoes. It represents the composite pattern of fractures which could serve to produce all of the observed shapes of small volcanoes. The orthogonal pattern of fractures are the ridge normal and ridge parallel directions while the oblique direction is that of fractures found only near active transforms and presumably present on inactive fracture zones also.
through intersecting fractures produces intersecting linear
volcanic ridges or a set of ridges that enclose a polygonal
basin. With further eruption from the linear ridges, the
basin fills. During this process, growth of the edifice is
probably limited to a few hundred meters in elevation. The
volcano may be large enough at this stage to possess a
small magma chamber. Subsequent summit collapse due to the
presence of such a magma body would produce a circular
symmetry [Fornari et al., 1987]. This symmetry could lead
to the formation of a new set of conical feeders. Further
growth of the edifice would form larger, circular
volcanoes.

Vogt and Smoot [1984] also propose an evolution of
morphology between small and large volcanic edifices
(Figure 20). They observed that some of the larger
edifices in the Geisha Guyot cluster of the Northwest
Pacific exhibit a distinct starfish-shaped morphology
(Figure 7). In most cases, the starfish "arms", typically
3-6 in number, are thought to be flank rift zones such as
those seen in the Hawaiian Islands [Fiske and Jackson,
1972]. These same starfish shapes are observed on guyots
and volcanic islands larger than the Geisha Guyots, which
are considered intermediate in size. Small seamounts are
diverse in their morphology, but the starfish morphology is
not observed. The length of flank rift zones increase (up
to > 100 km long) as the size of the edifice increases.
These observations suggest an evolutionary growth.
ARE THESE THE SAME FEATURES, WITH VOLCANIC ACTIVITY TERMINATED AT DIFFERENT STAGES OF EDIFICE GROWTH?

SMALL GUYOTS

LARGE GUYOTS

T9

GEBHA CHAIN

SCHEMATIC 500 M CONTOURS
BASE CONTOUR 5500M
PLATEAU BREAK DEPTH 2000 M

FIGURE 20. Schematic drawing [from Vogt and Smoot, 1984] showing possible evolution of small to large oceanic volcanoes. Note that guyots are among the largest individual volcanic features in the Pacific.
In studies of active flank rift zones [Fiske and Jackson, 1972; Ryan et al., 1981], observations indicate that magma rises in a central conduit and is then injected into flank rift zones from a shallow magma chamber. The intruding magma bodies are thought to be steeply dipping, bladelike sheets paralleling and bisecting the rift zone. Some injection events are accompanied by eruptions from the crest of the rift zone. Presumably, flank rift zones develop when the volcanic edifice reaches a critical size at which flank injection becomes easier than a summit eruption from the central conduit. Indeed, small seamounts lack flank rift zones [Batiza and Vanko, 1983], indicating it is easier to erupt from a low summit vent than to initiate rift zones [Vogt and Smoot, 1984]. Rift zones develop in response to the total regional stress field as well as the magma pressure [Vogt and Smoot, 1984].

The continued extension of the flank rift zone as the edifice increases in size reflects the increasing head of the magma column as the height of the summit increases. The higher the summit, the greater the pressure of the magma column at the base of the volcano and the easier it is to pump magma laterally into the flank rift zones. The length of the flank rift zones is limited by viscous losses and the gravitational pressure gradient [Vogt and Smoot, 1984].

The average and maximum rift length increases with summit plateau area. Rift lengths in large seamounts and
small guyots increase slowly with changes in pressure differential $P$ (about 0.2 km/MPa) (Figure 21), and are all relatively uniform in length. ($P$ is the pressure differential between the base of the magma column and the hydrostatic pressure at the same level without the edifice [Vogt and Smoot, 1984]). The rate of magma injection into these rift zones is low, and the shallow magma chambers are probably not well developed. High effective viscosity results in a high $P$ gradient (5 MPa/km). At about 100 MPa, maximum rift length increases rapidly with an increase in $P$ (about 3 km/MPa), but there is a great deal of range in the lengths of the rifts. This may indicate frequent injections and shallow magma chambers from which magma flows easily along flank rift zone conduits. The effective viscosity is low, allowing magma transport under low $P$ gradients (0.3 MPa/km) [Vogt and Smoot, 1984].

Regional tectonic stresses are important in controlling the initial geometry of the volcanic conduits. Eruption rate, distance from the vent, magma viscosity and pre-existing topography are all major controls on the form lava flows may take such as pillow lavas, lobate tubes, sheet flows, ponded lavas or hyaloclastites. As the volcano grows and becomes more sizeable, gravitational stresses within the edifice become important in controlling the morphology [Fornari et al., 1987].

Near axis seamounts show a preferential origin near transforms, fracture zones and overlapping spreading
FIGURE 21. Length of flank rift zones plotted against P, the pressure difference between base of magma column (2.9 g/cm$^3$) below summit caldera, and hydrostatic pressure at the same level outside volcanic edifice [from Vogt and Smoot, 1984]. Numbers indicate guyots and unnumbered circles are seamounts of the Geisha chain. Data suggest 1) rifts are very short or absent on small seamounts, 2) rift length increases slowly (0.2 km/MPa) up to about 100 MPa (1 kbar), after which maximum rift length increases rapidly (about 3 km/MPa) with increasing P.
centers [Menard, 1969; Lonsdale, 1985; Fornari et al., 1984; Batiza et al., 1989a]. It is possible that at these locations, where the crust is fractured more than usual, there are more magma conduits available.

Different flow morphologies (pillows, sheet flows, hyaloclastites, etc.) are a function of magma chemistry, locus of flow source on the seamount (flanks or summit), rate of extrusion, and depth of the seamount. The upper slopes of seamount flanks probably represent the angle of repose for assemblages of pillow lavas, talus and coarse pyroclastic debris [Vogt and Smoot, 1984]. However, steeper slopes may occur. Walker [1973] showed that the length of lava flows depends on effusive rate and that volcano slope is inversely related to effusive rate. Small, slowly erupted submarine flows would produce a greater proportion of fragmented debris which can accumulate at higher angles of repose. If the average eruption produces a small amount of lava and small flows which cool quickly and do not flow great distances down the flanks, slopes will be steeper. The eruption of silica-rich lavas which are more viscous and will not flow far will increase slope angle. Eruptions which produce more pyroclastic material rather than lava flows also form steep slopes. Such eruptions occur generally shallower than 500 m, at a level where exsolving volatiles cause fragmentation of the magma [Fisher and Schmincke, 1984]. Other factors contributing to steeper slopes which are not primary
volcanic processes, include increased slumping during and after volcanism, slumping due to inflation and seismicity and coral reef growth. The upper slopes may have steepened due to coral growth and the accumulation of cemented calcareous debris following volcanism. Lower flank slopes of mature volcanoes are affected more by sedimentary processes rather than primary volcanic processes (discussed in a later section). However, Lonsdale and Spiess (1979) suggest that the lower slopes of young seamounts are lava flow aprons rather than sediment aprons found around emerged volcanoes.

Many seamounts or guyots have flat tops and/or step-like flat terraces and steep scarps. These are explained as results of wave erosion [Menard, 1964].

VERTICAL MOVEMENT

Sclater et al. (1971) explained the sinking of oceanic lithosphere due to thermal contraction with age. Volcanoes sitting on the oceanic lithosphere would be expected to sink along with it as it subsides. However, drilling on atolls and volcanic islands exhibit significant subsidence [Ladd and Schlanger, 1960] which cannot be explained by simple subsidence as the lithosphere thickens and cools. In other instances, uplift of volcanic platforms is seen. The study of such vertical movements is useful in determining properties of the oceanic lithosphere. The
possible causes of such motions are thermal and mechanical in nature.

Many atolls stand at higher elevations than expected for the age of the seafloor upon which they sit. Often, they are located near active or recently active volcanoes. Eustatic sea level highs are unable to explain the differences in elevations of atolls of the same age, or the extent of such a rise in sea level (in some instances as much as 70 m [McNutt and Menard, 1978]). This led McNutt and Menard [1978] to propose tectonic uplift due to deformation of the lithosphere by volcanic loading.

The lithosphere can respond to loads like a thin elastic plate overlying a fluid. Loads on the plate cause bending and displace the underlying fluid. Buoyancy forces from the displaced fluid are the actual supports of the feature producing the load, while bending of the lithosphere distributes the weight over an area greater than the feature itself. This flexure creates a depression near the edifice and a rise or arch further seaward (Figure 22) [Walcott, 1970]. Flexural studies are very important in determining mechanical properties of the lithosphere, such as: elastic thickness and flexural rigidity. Elastic thickness is the maximum thickness of the lithosphere allowable to account for a certain amount of flexure, assuming elastic plate behavior. Elastic thickness can be mathematically determined by modelling to fit a certain flexural profile, or can be estimated using gravity and
FIGURE 22. Bathymetric and free air gravity profiles across the Hawaiian ridge near Oahu showing Hawaiian swell, moat and arch [from Watts, 1976].
bathymetry data [Watts, 1978]. Flexural rigidity is a measure of the stiffness of the lithosphere and determines the wavelength and amplitude of flexure [Watts, 1978]. Flexural rigidity is determined by the elastic thickness.

Watts [1978] established that elastic thickness depends on crustal age. Elastic thickness is fixed at the time of loading and does not change significantly with time [Watts et al., 1980]. Small elastic thicknesses are associated with loads on young lithosphere while large thicknesses are calculated for older lithosphere. As oceanic lithosphere ages, it thickens and cools, therefore responding more rigidly to surface loads. By comparing computed gravity anomaly profiles for different elastic thicknesses with observed profiles to determine the actual elastic thickness (Figure 23), Watts et al. [1980] determined the age of the lithosphere at time of loading. In this manner, it is possible to estimate whether an island or seamount was formed on young lithosphere (near the ridge crest) or old (off the ridge) (Figure 24). Estimates of the distribution of volcanism can then be made (i.e., for the Pacific plate since the Jurassic (Figure 25) [Watts et al. 1980]).

McNutt and Menard [1978] discuss how growth of a young volcano flexes the sea floor and affects the apparent sea level on nearby atolls. Figure 26 shows atolls sinking with normally subsiding oceanic crust. A growing volcano flexes the lithosphere, uplifting the atoll on the arch while lowering the one in the moat. This explains how some
FIGURE 23. Comparison of observed free-air gravity anomaly profiles with computed profiles based on the elastic plate model and assumed values of the elastic thickness $T_e = 5 \text{ km}$ and $25 \text{ km}$ [from Watts et al., 1980].

a) Line Islands ridge at $1.7^\circ \text{ N}, 157^\circ \text{ W}$ and Mid-Pacific mountains at $18.5^\circ \text{ N}, 179.5^\circ \text{ W}$. The observed profiles can best be explained by the computed profile of $T_e = 5 \text{ km}$. A ridge crest origin is concluded for these. b) Makarov Guyot at $29.6^\circ \text{ N}, 153.5^\circ \text{ E}$ and Louisville ridge at $39.3^\circ \text{ S}$ and $167.5^\circ \text{ W}$. The observed profiles are best explained by the computed profile of $T_e = 25 \text{ km}$. Therefore, an off-ridge origin is deduced. The numbers at the right of each computed profile are the r.m.s. differences between observed and calculated gravity anomalies and the number below each bathymetric profile is the distance in nautical miles along a ship track, used to locate the feature.
FIGURE 24. Schematic model [from Watts and Ribe, 1984] illustrating a) that seamounts formed on or near a mid-ocean ridge crest are associated with low elastic thicknesses (Te), while seamounts formed off ridge have high values. b) As the ridge axis migrates and the lithosphere ages, values of Te are not expected to change appreciably with time. The value of Te provides information on the tectonic setting of seamounts.
FIGURE 25. Maps from Watts et al. [1980] a) showing the distribution of volcanoes formed on young (A) versus old (B) lithosphere based on gravity and lithospheric flexure studies, and b) summarizing the inferred distribution of volcanism on the Pacific plate during the interval 90-120 Ma. Triangles represent features formed at the ridge crest; circles show features formed off-ridge.
FIGURE 26. Model of apparent sea level change on coral atolls caused by volcanic loading on an elastic lithosphere [from McNutt and Menard, 1978]. The nodal point, 1, depends on flexural rigidity.
atolls in an area are uplifted while most in the same area are not. The distance of the nodal point, indicated on the figure, from the volcanic load depends upon the flexural rigidity. Atoll uplift, therefore, can be used to determine the flexural rigidity of the oceanic lithosphere. However, unreliable measurements might arise if sea level was higher during previous interglacial periods or there was rapid erosion after emergence [McNutt and Menard, 1978].

Excessive subsidence of atolls and seamounts has been documented. Proposed causes include: 1) isostatic adjustments to the weight of the volcano or coral reef cap, 2) viscoelastic flexure of underlying lithosphere, and 3) subsidence following volcano formation on anomalously shallow regions, or swells, of the sea floor [Detrick and Crough, 1978]. Although isostatic adjustments would explain the high subsidence rates observed and may account for some of the subsidence [Watts and Cochran, 1974], a volcano should be completely compensated almost as soon as it is built, assuming time scales of isostatic adjustments are the same as those for the continents [Detrick and Crough, 1978]. Viscoelastic flexure involves the lithosphere responding to loading by viscous flow. Watts [1978] showed that this mechanism could not explain gravity and bathymetry data from the Hawaiian-Emperor seamount chain, and presumably, similar seamount chains. Detrick and Crough [1978] proposed that volcanoes form atop swells
high, wide, anomalously shallow regions often elongate in form [Crough, 1983]. As the swells return to normal depths, the volcanoes subside. Topographic swells have been found to be a major morphological feature of the sea floor. Detrick and Crough's [1978] suggestion becomes even more plausible upon discovering that these swells are quite often coincident with volcanic edifices. In fact, almost every oceanic volcano sits atop a topographic swell [Crough, 1983].

The origin of such swells is still uncertain. If the Hawaiian Swell is taken as a typical example of such a feature, any proposed mechanism must be able to explain the following observations [Von Herzen et al., 1989]: 1) rapid uplift of the swell, 2) the gradual disappearance of the swell after passing over the hotspot, 3) the lack of a notable heat flow anomaly over the older part of the swell, and 4) the apparent compensation of the swell at depths $> 50-100$ km (within or below the lower part of the thermal plate) [Crough, 1978].

Previous evidence has led to suggestions of a thermal origin involving substantial reheating of the lower part of the lithosphere. The compensation depth obtained by Crough [1978] was consistent with the amount of lithospheric thinning needed to explain the initial uplift and gradual subsidence of the swell using a simple reheating model [Detrick and Crough, 1978]. The apparent gradual increase in heat flow over the older part of the swell [Von Herzen
et al., 1982] was also consistent with reheating of the lower part of the lithosphere. However, Von Herzen et al. [1989] have discovered that the magnitude of the heat flow anomaly is much less than originally thought. This severely hampers proposed mechanisms of origin of swells which invoke high temperature material close to the surface. The subsidence rates are explained adequately but the absence of high heat flux is not. Von Herzen et al. [1989] suggest that dynamic support of swells by plume-related convection offers the best explanation for the observations discussed earlier. In such a model, the upper part of the lithosphere behaves rigidly, acting as a "conducting lid", while the lower portion can deform ductily and take part in the flow [Parsons and McKenzie, 1978]. A rising mantle plume can penetrate and replace lower lithosphere and underlying asthenosphere which has been cooled conductively with hotter, bouyant mantle material [McKenzie et al., 1980; Parsons and Daly, 1983]. The swell is supported by the uplift from dynamic convection and the thermal expansion of the overlying conducting lid [Von Herzen et al., 1989].

The convective model described above can account for the general shape of the swell but not the shallow depth of compensation, nor the elastic plate thickness. However, coupled with a low-viscosity zone in the upper mantle to simulate a partial melt, Robinson and Parsons [1988] showed that the convective model can explain the observations.
The low-viscosity zone would allow lateral flow at the top of the rising mantle plume to entrain the lower lithosphere material and to sweep it away. This process could explain the rapid uplift of the swell. The shallow compensation depth of the swell can be explained by mantle convection in the presence of a low-viscosity zone since effective temperature changes would be concentrated near the base of the conducting lid [Robinson et al., 1987]. Also, the buffering effect of partial melting on mantle temperatures may explain the low heat flux over the swell. Any excess heat may be released through volcanism and crustal thickening over the central portion of the swell [Von Herzen et al., 1989].

SEDIMENTATION AND SEDIMENT TRANSPORT

Sedimentation is an important component in the morphologic development of seamounts. On older lithosphere sediment can completely bury smaller volcanoes. In less extreme cases, sedimentation can modify the apparent slopes on seamounts and mask the features on their summits. Sediment tends to make slopes less steep and will bury interesting features like summit craters or satellite cones. Sedimentation can even affect heat flow readings [Von Herzen et al., 1989].

A variety of sediment types are found on seamounts, including volcanogenic, lithogenic, hydrothermal, biogenic
and authigenic. Variations in deposition and removal of these materials are the result of differences in volcanism, biological productivity, current action, bioturbation and hydrothermal activity, which can lithify sediment in situ to form crusts [Levin and Nittrouer, 1987; Stanley and Taylor, 1977].

Levin and Nittrouer [1987] discuss how sediment character (texture, composition, color) is influenced by environment. Latitude, water depth and age affect sediment supply and removal through changes in biologic productivity, carbonate dissolution and hydrothermal precipitate production. In high latitudes with areas of high productivity, the supply of foraminifer sand-sized particles to the sea floor increases. As water depth increases, carbonate dissolution generally increases; therefore, the removal of foram sand-sized particles from the sea floor also increases. Young seamounts that are active or were recently active commonly have more hydrothermal and volcanic sediment exposed. Older seamounts have greater sediment cover and lack surface hydrothermal precipitates.

Different settings within seamounts affect sediment accumulation (Figure 27) [Levin and Nittrouer, 1987]. Sediment supply to seamounts occurs mostly as pelagic fallout; hence, the interiors of calderas and pit craters, areas of weak current activity, are effective sediment traps. The coarser deposits are found on topographic highs
FIGURE 27. Schematic diagram of topographic settings and generalized sediment types on a typical young seamount in the eastern Pacific Ocean [from Levin and Nittrouer, 1987].
(summits) where current activity winnows out finer materials. Finer sediments are found at the base of seamounts where carbonate dissolution is greatest, provided the carbonate compensation depth is exceeded, and winnowed sediments accumulate.

Physical and biological activity cause smaller scale variations in sediment texture and distribution [Levin and Nittrouer, 1987]. Bedforms such as ripples concentrate different grain sizes within different parts of their waveform. Organisms redistribute sediments by bioturbation or selective use of certain sediments in the formation of tests. Faunal abundances are greatest in calderas, pit craters and on summit benches while the lowest are at the base and in hydrothermal fields of seamounts [Levin and Nittrouer, 1987].

Rates of sedimentation vary in different parts of the ocean. Therefore, the effect of sedimentation on morphology will depend on the location of the seamount. Rates are lowest (\(< 1 \text{ cm/1000 yrs}\)) in the deepest parts of the oceans. Intermediate rates (1-3 cm/1000 yrs) are found in the shallower oceans, on mid-ocean ridges and in high latitudes where there is high biogenic productivity [Kennett, 1982]. The highest rates (10-30 cm/1000 yrs) are found at continental margins, inland seas, abyssal plains and in the North Atlantic.

Sea level influences sedimentation rates. Rates are higher in the deep sea during times of low sea level
During low sea level stands, continental shelves are exposed, so the eroded materials from the continents are flushed to the deep sea. During high stands, these materials would be deposited on the continental shelf. Davies and Worsley [1981] also pointed out that during high sea level, there is more biogenic precipitation on the continental shelves, starving the ocean basins.

Sediment can be transported downslope or across slope. Cross-slope movement is accomplished mainly by currents. Downslope movement is accomplished mainly by gravity related flow processes including debris flows and creep.

Stanley and Taylor [1977] studied sediment transport processes on the twin-peaked Gilliss Seamount in the North Atlantic. Their observations are similar to those from other seamounts. Detailed bathymetry of the Gilliss Seamount shows a step-like topography formed by steep scarps alternating with ledges. Lonsdale et al. [1972], as discussed previously, interpreted similar terraces on the Horizon Guyot as a result of primary volcanic processes. The sediment on the ledges is driven by currents, as indicated by ripples. As the traction mode nears the edge of the ledge, spill-over occurs. The material is then transferred down the steep scarps by gravity-driven processes (Figure 28). At the base of the scarp, the sediment is once more open to entrainment as part of the traction carpet. The sequence is repeated until the
sediment reaches the base of the seamount. At this point, gravity flow is no longer an important factor, but bottom currents continue to modify the accumulating sediment. This progressive downslope movement explains the lack of significant sediment accumulation on the upper flanks of seamounts.

The lower flank slopes of seamounts consist mainly of sediment aprons. These aprons, or cones, or fans, surrounding the base are generally small, but increase greatly in area and volume once the guyot state is reached. Menard [1956] stated that these aprons were probably composed of a mixture of volcaniclastics, shallow water bioclastics and pelagic sediment transported from the upper slopes of the seamounts by slumps and debris flows. The crenulations typically seen in bathymetric contours of the lower slopes of ocean volcanoes represent slump fronts and channels carved by debris flows (Figure 29).

In order for large aprons to develop, it is necessary for the volcanic vent to approach the sea surface. More clastic debris can be formed and rock density decreases because vesicularity increases due to lower confining pressure. This begins at about 1 km depth and accelerates as the vent approaches a few 100 m of the surface [Vogt and Smoot, 1984]. The importance of producing more clastic debris is obvious. Decreased rock density is helpful in producing large aprons because material can be entrained
FIGURE 29. Detailed bathymetry of Makarov Guyot (100 fm = 183 m contour interval) [from Vogt and Smoot, 1984]. Arrows show paths of some major debris flows.
for longer periods and can be deposited farther from the base of the volcano.

One effect of large aprons is to alter heat flow. Rapidly deposited cold sediments can lead to a reduction in the observed surface heat flux, while radioactive heat produced by the sediments increases it. Hutchison [1985] discussed the affects of sedimentation rate, type and depth of burial on heat flux observed. High sedimentation rates reduce the flux. The response of the surface flux is rapid, but after deposition stops, the system gradually recovers. Highly conductive sediments such as salts, evaporites and sandstone affect the heat flux the least, while low conductivity, low porosity materials like limestone affect it the most (Figure 30). Thermal diffusivity increases with depth of burial due to sediment compaction.

CURRENT ACTION

Ocean currents affect a variety of processes which, in turn, affect the morphology of seamounts. Surface currents are intimately involved with sedimentation rates, coral growth and formation of ocean bottom currents. The circulation patterns of surface waters and bottom and intermediate waters differ greatly. Surface waters have narrow latitudinal ranges very similar to climatic belts. Surface currents are primarily wind-driven. Deeper
FIGURE 30. Comparison of the alteration to the surface heat flow due to the deposition of salt, shale, sandstone and limestone [from Hutchison, 1985]. Each curve accounts for 10 km of deposited material.
circulation patterns originate from surface processes, then spread widely through the water column [Warren, 1971]. The dominant process involved in these deeper currents is gravity-driven thermohaline circulation.

Although precise directions, longevity and intensities of ocean bottom currents are not well known, they are known to exist. Swallow [1971] discovered deep-sea currents of 5-10 cm/sec that completely change direction within the space of a month. Intensities can be inferred by observing the following: the extent of winnowing of sediment, current modification features (ripples, sediment scour, etc.), and the presence of organisms [Stanley and Taylor, 1977]. Current direction can be determined in some cases, such as from asymmetric ripples and current tails.

Deep ocean current velocities are periodic. Semidiurnal and monthly tidal variations have been recorded (Figure 31) [Kennett, 1982]. The effect of tidal energy is greatest in shallow seas and on continental margins. In times of lower sea level when continental shelves are exposed, tidal energy is a greater factor in the deeper ocean. Tidal currents are known with velocities up to 17 cm/sec just above the sea floor on seamounts [Kennett, 1982].

Currents have been shown to be affected by seamount topography. Submarine topography can change current direction and velocity. Flow is accelerated over the peaks of seamounts [Genin et al., 1986]. The faster currents not only are capable of entraining more sediment, but also
FIGURE 31. Sample plot of current meter data to show tidal periodicities in the current [from Lonsdale et al., 1972].
enhance biogenic production and coral growth (Figure 32), at depths where growth is possible, by providing more nutrients. Coral growth will be discussed in more detail in a later section. This current acceleration is another explanation for the tendency to have a thinner sediment cover on seamount summits than on the flanks. In addition, it affects the type of sediment and grain size found [Levin and Nittrouer, 1987]. More biogenic sediments are found in areas of higher nutrient content due to the greater productivity. Where currents are strong enough to winnow fine sediments (over seamount summits) coarser lag deposits are found.

Where currents are forced to divert around seamounts, erosion and deposition of sediments is affected. More erosion will occur where the current first encounters the obstruction, while deposition will occur on the lee side. This will affect the appearance of the volcano.

EROSION AND MASS WASTING

Erosion is a powerful influence on the morphology of any geologic landform. Seamounts are no exception. Whether near the ocean surface or deeply submerged, erosion can affect the shape of any edifice. On near surface seamounts, erosion by ocean surface currents, wave action, bioturbation, mass wasting and in some cases, subaerial erosion may occur. On deeper seamounts, erosion is
FIGURE 32. Bathymetry, densities of *Stichopathes* and sediment cover across two satellite peaks on Jasper Seamount [from Genin et al., 1986]. Each dot indicates the animal density or sediment cover in a single photograph. Note that animal densities follow the bathymetric line on the sharp peak (C) whereas lower densities are found near the center of the wide Peak (A) than near its edges. Animal densities are relatively higher on small topographic prominences. Note the sharp increase in sediment cover in the depression near peak C.
accomplished by bottom currents, dissolution of sediment, bioturbation and mass wasting.

Malahoff et al. [1982], suggested that slumping is typical on seamounts and guyots. Indeed, major landsliding is apparently an important factor throughout the growth of oceanic volcanoes, continuing even after dormancy [Moore et al., 1989]. Some of the largest landslides occur near the end of the shield-building stage when volcanoes are active and vigorously extending their shorelines. However, during the entire shield-building stage, intermittent mass wasting events probably occur, but the evidence is obscured by incorporation of the landslide debris into the edifice as it grows [Moore et al., 1989].

Many slope failures on Hawaiian volcanoes display movement perpendicular to the primary rift zones of the volcano, or head on the sides of long, shallow submarine ridges which are most probably submerged rift zones [Moore et al., 1989]. There is a cause-effect paradox at work here. Rift zones form along zones of tension at the head of landslides, but landsliding can also be triggered by the lateral motion caused by magma injection into the rift zones [Moore et al., 1989].

Continued mass wasting after the cessation of volcanic activity, could provide an alternative explanation to flank rift zones (discussed in a previous section) for the starfish shape of some volcanoes. The pattern of radial arms could represent unslumped residual ridges, the remains
of a more extensive volcano (Figures 7, 29). On the other hand, the mass wasting may simply remove debris from between the rifts, made more resistant by cores of dikes. This would expose flank rift zones previously embedded in the edifice.

Erosion is responsible for such striking morphological features as flat-topped guyots. Guyots are seamounts that were truncated by erosion at the ocean surface [Menard, 1964]. Ideally, guyots are perfectly flat-topped with no summit plateau relief. Achieving such an ideal state requires shoreline erosion rapid enough to negate thermal or other subsidence. In addition, there can be no differences in the erodibility in the rocks of the edifice; otherwise, resistant knobs and shallow depressions may be formed [Vogt and Smoot, 1984].

However, the summits of most guyots are not ideal. Generally, the observed summit relief increases with increasing summit area and a decrease in thermal age and erosion rate. The summit area versus summit relief relation reflects the importance of shoreline erosion. On such near surface seamounts, erosion can truncate smaller seamounts faster than large ones [Grigg and Epp, 1989]. (This is discussed in more detail in the next section.)

The importance of erosion cannot be denied. Menard [1983] discussed the role of erosion in isostasy. He stated that the principal factors which influence vertical movement of volcanic islands are: thermal subsidence of
the underlying crust (at a rate proportional to the square root of time), eustatic sea level changes, and isostasy. Isostatic effects result from compensation for thermal subsidence, eustatic fluctuations, loading by coral growth (minimal due to the small density contrast between coral and water) and unloading by erosion. According to Menard [1983], the most important isostatic effect is erosion, which is greatly influenced by the presence or absence of barrier reefs (to be discussed in the next section).

On near surface seamounts, there are two main types of erosion - wave and river. The relative importance of these types of erosion depends upon the radius of the island. The rate of wave erosion varies with perimeter while the rate of river erosion varies with area. Smaller islands tend to be eroded more by waves, while large ones are more eroded by streams [Menard, 1983]. This is reinforced by the influence of height on rainfall. The higher the mountains, the higher the percentage of precipitation and the greater the amount of erosion. Windward sides of high islands can be eroded up to 30-40 x faster than the lee side. This can be seen on the island of Kauai where once the windward side was eroded, rainfall centered on the lee side and deep canyons were cut [Menard, 1983].

Subsidence combined with wave and river erosion affects morphology and distribution of off-lying volcanic islands. In cases where thermal subsidence is slowed by erosion but not balanced, wave erosion may form a series of narrow
terraces separated by cliffs; or a relatively smooth shelf that is steeper near the shelf break than in the interior (Figure 33) [Menard, 1983]. On large tropical volcanoes where river erosion has carved out ridges and spires, subsidence causes the narrow saddles between spires to disappear and leave small outlying islands.

Erosion is affected by the rate of flow of bottom currents and the corrosiveness of the bottom water to biogenic sediments [Kennett, 1982]. Current activity can lead to oversteepening of sediments on seamount flanks which, in turn, may lead to slumping [Levin and Nittrouer, 1987]. Erosion by currents can also form moats (Figure 34) or accentuate those formed around seamounts as a result of lithospheric flexure. The longer the biogenic sediment is exposed to the bottom water with no deposition occurring, the greater the possible dissolution of the sediment.

REEF CAP GROWTH

Coral reef caps form on some seamounts and serve as modifications to the morphology. They often lend a flat-topped look to seamounts which are termed guyots (Figure 35). They can influence the erosional history of oceanic volcanoes: a reefless volcano will erode faster than one with a protective reef. Reefs may trap erosional debris, thus affecting subsidence due to loading, which in turn
FIGURE 33. Profiles of eroded basement summits of four small guyots in the Gulf of Alaska compared with theoretical shapes produced by shelf widening at 1 km/my and thermal subsidence at different equivalent thermal ages [from Menard, 1983]. Note the smooth shelves are steeper toward the shelf break than in the interior.
FIGURE 34. Schematic diagrams showing the formation of moats by bottom currents [from Davies and Laughton, 1972].
FIGURE 35. Morphology of dated oceanic volcanoes with barrier reefs [from Menard, 1983] showing the typical flat topped appearance of guyots. Short dashed lines indicate relief interpolated into profile, and long dashed lines represent assumed subaerial and probable relief buried under reefs. Vertical exaggeration, 5 x.
affects the erosional/depositional history. Reefs are important not only because they contribute to shaping seamounts, but they can also be used to study sea level changes and tectonic processes (subsidence and uplift).

The maximum upward growth of coral reefs is about 10 mm/yr [Grigg and Epp, 1989]. This rate applies in areas of optimum growth for corals, at depths of 5-10 m and at low latitudes below the Darwin Point. The Darwin Point, presently at about 29° [Grigg, 1982], is the latitude at which reefs cannot grow upward fast enough to keep pace with subsidence of the edifice. It has varied from 27°-31° over the past 20 my [Rotondo, 1980]. Below a critical depth, coral reefs are unable to maintain a net positive rate of upward growth. Growth rates decline very quickly below the critical depth. For example, at 30 m below sea level, growth rates are 15%-40% of their maximum [Grigg and Epp, 1989]. If growth is unable to keep up with subsidence, the reef will drown. The depth at which drowning will occur is between 30 and 40 m. It could be less depending on variations in light, temperature, sedimentation, turbidity and erosion. Some corals can survive at greater depths, but cannot build to the surface.

Genin et al. [1986] observed that the abundance of coral varied with seamount topography. For instance, on narrow peaks, corals are more abundant on the crest; on wide peaks, they are more abundant at the edge of the crest; and on knobs and pinnacles, abundance increases. It was
suggested that the cause was accelerated currents over the peaks. This provides higher fluxes of nutrients to the corals and benefits their growth (Figure 32). Indeed, modern corals are known to depend upon breaking surf, up to a limit, thus preferring the windward edge of an island [Vogt and Smoot, 1984].

In fact, current acceleration, up to a limit, over topographic features may increase local populations of all suspension-feeding organisms. In addition, topographically induced upwelling may contribute to particle flux on the benthic layer. The eastern tropical Pacific is characterized by a shallow mixed layer and a pronounced oxygen minimum. Low plankton abundances are associated with oxygen minimum zones in the water column. Such zones, if sufficiently broad vertically, serve as effective barriers against plankton. The consumption of particulate matter in these regions is reduced, thus increasing the supply of undegraded organic material reaching the sea bed [Wishner et al., 1990]. Sharp benthic zonations may result due to the enhancement of the food supply and the exclusion of hypoxia-sensitive organisms, as seen on Volcano 7 [Wishner et al., 1990].

The subsidence of an island depends on its size and the presence of reefs [Menard, 1983]. Islands with barrier reefs sink as though on thermally youthful crust, regardless of actual age (Figure 36). Isostatic uplift is eliminated because the barrier reefs prevent wave erosion
FIGURE 36. Comparison of subsidence of barrier reef islands, reefless islands and reefless guyots [from Menard, 1983]. Presumably, guyots remained at sea level, like reefless islands, until truncated and then sank like barrier reef islands of the same age. The 1 and 25 my lines indicate subsidence expected for oceanic crust with these thermal ages.
and capture stream erosion products. Islands with no reefs do not sink until they are truncated by erosion. Apparently, thermal subsidence is compensated by isostatic uplift as a result of erosion.

The morphology of islands is affected by the presence or absence of reefs [Menard, 1983]. The shelves surrounding islands are mostly erosional. As volcanic islands grow, cliffs may form at the same time. But sea cliffs are not found on older islands with barrier reefs. The reef protects the island from further wave erosion while streams erode any existing cliffs. On reefless islands, there are cliffs, sea stacks and the shelf is very rocky.

Embayments form when valleys submerge. These are conspicuous on islands with barrier reefs, but rare or minor on reefless islands [Menard, 1983]. On younger islands with vigorous stream erosion, deltas may fill some bays. On older islands, this is not the case since there is not an adequate supply of sediment.

Reefless islands tend to remain at sea level until almost totally truncated. This is why guyot platforms are smooth, indicating they were stable long enough to be eroded away. The surrounding seafloor continues to sink as it cools while the island is isostatically uplifted. Once the island has been leveled, it subsides along with the seafloor at an equivalent thermal age [Menard, 1983].

The longer a reefless island remains at sea level, the wider its insular shelf will become, assuming late shield-
building or post-erosional stage volcanism does not flow over and reset the width/age relation [Menard, 1983]. Also, the shelf will be wider on the weather side of the island than on the lee side because of greater erosion.

Grigg and Epp [1989] suggest that summit area of the volcanoes upon which reefs form, may be a factor in the success or failure of atolls. Islands with smaller summit areas are eroded to a greater extent during lower sea level stands than those with larger summits. Since such islands are truncated to greater depths, as sea level rises, the growth of corals may not be able to keep pace with the rate of sea level rise. Larger summit areas are not eroded as quickly so can support growing reefs during sea level changes. There are indeed paired atoll-guyots observed, such as Bikini-Sylvania (Figure 37) [Grigg and Epp, 1989]. Bikini is an atoll with a living coral reef, while the smaller and deeper Sylvania is a guyot with a drowned reef.

Obviously, once a coral reef is drowned, it can no longer continue to influence the morphology of the seamount. Unless, of course, there is tectonic uplift of the edifice or sea level falls and the reef is able to reestablish itself.

Coral reefs have been known to come back after having been exposed and killed during low sea levels. During periods of low sea level, subaerial erosion works on the exposed reefs and forms karst features. This lends an uneven, craggy appearance to the summit morphology, as the
FIGURE 37. Perspective diagram of Bikini Atoll and adjoining Sylvania Guyot [from Emery et al., 1954].

The reef on Bikini Atoll is still living while that on Sylvania Guyot has been drowned.
mid-Pacific Mountains where steep sawtooth profiles are seen [Winterer and Metzler, 1984]. Perhaps this is yet another explanation for Menard's [1964] wave cut platforms.
CHAPTER 3
SEAMOUNT 6 TEXTURAL ANALYSIS

INTRODUCTION

In addition to being able to accurately map structural features of the seafloor, sidescan sonar systems such as SeaMARC II can distinguish areas of differing acoustic backscatter strength. Energy backscattered from any surface outcropping on the seafloor is a function of variables including 1) the composition and roughness of the surface, 2) the slope angle of the reflecting surface and 3) acoustic look angle of the outgoing acoustic energy.

If the important controls on backscatter strength of the seafloor can be quantified, sidescan sonar images, such as produced by SeaMARC II, will come much closer to becoming geologic and lithologic maps of the seafloor. This is important because 100% optical/sample coverage is not possible, even for small areas. To this end, I have undertaken the following study.

The tools used in this study are the acoustic mapping systems SeaMARC II and SeaBeam. My technique was to choose a feature set consisting of statistics measured from backscatter data to and attempt to correlate members of this feature set with observations of geology from bottom photographs. The location chosen for this study is a well-studied seamount in the eastern Pacific - Seamount 6.
Seamounts are ideal for this study since they offer variable slopes and variable geology within a small area. Seamount 6 has been mapped with SeaBeam, has overlapping SeaMARC II coverage, abundant towed camera and submersible photographs, and has been well sampled.

Provided in this chapter, is a brief introduction to the data acquisition system, location of study area, texture features used to describe remotely sensed data and classification methods followed by an application of the analytical techniques to Seamount 6.

SeaMARC II

Remotely sensed data is an important tool in the study of the seafloor. 100% photographic/sample coverage is not possible; therefore, remote acoustic mapping systems are utilized to provide wider swaths of data than produced by one pass of a towed camera, for instance. Acoustic systems are used in seafloor mapping since electromagnetic radiation does not penetrate the water column. The following section is a brief overview of the tools and techniques of seafloor remote sensing used in my study. Specifically, I provide an introduction to SeaMARC II, discussing physical description, data acquisition, how images and bathymetry are created, corrections applied to data and data analysis.
Systems used in the study of the seafloor are multi-beam and bathymetric sidescan sonars. Multi-beam echo sounders, such as SeaBeam, yield high resolution (about 10 m vertical and 100 m horizontal) bathymetric maps. For a complete system description, see Renard and Allenou [1979]. Bathymetric sidescan sonars, such as SeaMARC II, yield high resolution (10 m) acoustic imagery, plus swath bathymetry with 50 m vertical and 100 m horizontal resolution. As SeaMARC II imagery gives high resolution detail, we hope to use this as an analog to aerial and satellite remotely sensed imagery to aid in the classification of the seafloor.

A brief description of SeaMARC II follows. A complete system description can be found in Blackinton et al. [1983]. The SeaMARC II seafloor mapping system (Figure 38) consists of two subsystems. The first produces digital sidescan sonar images by recording acoustic backscatter amplitudes versus time. The second measures the direction of the acoustical returns versus time and converts the angles to bathymetry [Blackinton et al., 1983]. The unit is towed at about 100 m beneath the sea surface for deep ocean surveys to avoid surface sound speed variations and to enhance platform stability. Normal survey speeds are 8-9 kts. The sidescan swath width is limited by hardware to 10 km, while the width of the bathymetry swath is limited to 60° on either side of the towfish.
FIGURE 38. Schematic of the SeaMARC II acquisition system [from Bergersen, 1991]. Acoustic backscatter amplitudes are referenced according to arrival angle ($\theta$) and radial distance ($R$).

FIGURE 39. Schematic of the acoustic "footprint" of a sidescan sonar system [from Johnson and Helferty, 1990]. The grey area represents the intersection of the beam pattern with the ocean floor. A typical beam width for such a system is $2^\circ$ along-track.
Beam forming is accomplished by two parallel, inclined sonar arrays on either side of the SeaMARC II towfish. The port arrays operate at a frequency of 11 kHz while the starboard arrays are 12 kHz in order to reduce interference of returning signals from either side. The transducer arrays produce a fan-shaped beam about 2° wide along track (Figure 39). Each ping, or outgoing pulse, (in water > 1000 m deep) images a swath 10 km wide across track. The result is a sampling of a bowtie-shaped piece of the seafloor, centered at nadir.

Sidescan sonar images are created as two-dimensional displays of pixels. The image pixels are acquired from the bowtie-shaped swath by dividing the returning signal into increasing intervals of time. For each ping, 2048 pixels, each 5 m across and each with an associated intensity which graphically represent how sound interacts with the seafloor, are generated. Each digitized pixel is represented as an 8-bit word, limiting the range of image intensity values to 256. The middle 80 pixels are discarded since they are generated much too rapidly to be resolved. The result is an image with 984 pixels per ping per side of the ship track.

The intensity value assigned to each pixel depends upon how the acoustic signal interacts with the seafloor. The acoustic return received at the towfish is a combination of backscattered (diffracted) and specularly reflected (as by small mirrors) sound from the ocean bottom (Figure 40).
FIGURE 40. Possible results of an outgoing acoustic signal interacting with the ocean floor [from Johnson and Helferty, 1990]. The energy from the signal can either be reflected away or backscattered to the receiver.
Rough surfaces both scatter and reflect, while smooth surfaces mainly reflect specularly. Thus, except when the angle of incidence is 0, smooth surfaces will appear less reflective. SeaMARC II plots strong reflectors as dark, so rough surfaces, which tend to return more signal to the towfish, will appear dark on a SeaMARC sidescan image and be associated with higher intensity values. By the same token, smooth surfaces appear lighter on a SeaMARC sidescan image because less energy is returned to the towfish.

SeaMARC II bathymetric information is collected in the same manner as the sidescan image data, but is processed differently via a phase difference algorithm. The pairs of transducers on each side of the towfish operate in parallel on transmission. Upon reception, each row in the pair is sampled separately. Any signal with an incidence angle off normal to the transducer face will have a different phase lag at each row. The depression angle ($\theta$) of the reflector (Figure 41) is calculated from this phase lag [Hussong and Blackinton, 1983]. The slant range (SR) (Figure 42) to the reflector is calculated using round trip travel time of the signal and an assumed sound speed of 1500 m/s. Conversion of SR and $\theta$ for each reflector into across-track distance and depth produces values which can be contoured, creating a bathymetric map.

Processing corrections applied to SeaMARC II image data are of two types - radiometric and geometric. Radiometric corrections refer to those that remove all contributions to
FIGURE 41. Schematic of SeaMARC II phase difference measurement technique for determining the acoustic angle to the target reflector. Using precise arrival times and this calculated angle $\theta$, range and depth pairs can be calculated across-track. [From Blackinton et al., 1983].
FIGURE 42. Diagram showing how regional slope can cause the target reflector to be displaced from its actual position, due to the flat bottom assumption. Looking upslope, the 2-way travel time tricks the system into "thinking" the target is closer to nadir than it really is. Conversely, a target downslope will be displaced farther from the ship track than in actuality. This is called the layover effect and can be corrected for in order to move the targets to their proper positions [from Reed and Hussong, 1989]. SR is slant range and C is speed of sound in water. See text for details.
the sidescan data which are not indicative of actual changes in the acoustic character of the seafloor [Reed and Hussong, 1989]. Geometric corrections position features as close as possible to their actual locations. Radiometric corrections I focus on are time varying gain, angle varying gain and amplitude gain; the geometric correction most important to this study is layover correction.

Two of the radiometric corrections are applied automatically to the SeaMARC II sidescan data during acquisition. One is time varying gain, used to correct for spreading and attenuation of the acoustic signal [Bergersen, 1991]. The other is angle varying gain which corrects for variations in the transmitted beam pattern and the change in backscattered signal intensity observed at different look angles [Reed and Hussong, 1989]. A third type of amplitude gain is a scaling factor applied by an operator in response to the image amplitude of the data being recorded. These preprocessing corrections attempt to minimize image distortion, but are often inaccurate for the seafloor being surveyed. Post-acquisition processing is needed to eliminate these system and operator induced signals. However, in order to avoid loss of any pertinent acoustic information due to further processing and filtering which would remove the aforementioned distortions, we used a minimally processed sidescan record (Figure 43) in this study for extracting texture information.
FIGURE 43. This raw SeaMARC II sidescan image shows sample locations from the preliminary study.
Before spatial analyses of sidescan sonar images can be attempted, the pixels must be correctly positioned, so a layover correction is applied. The layover correction is a post-acquisition processing technique which corrects the across-track distance at which pixels have been placed. During acquisition, most sonars assume the bottom is flat, at a mean depth given by the nadir signal return. If the depth across-track is different from the nadir depth, calculations of across-track distance will result in the pixels being misplaced. As seen in Figure 42, if the seafloor actually slopes, the calculated slant range (SR) using the time of travel to the target reflector (B in Figure 42) and back to the towfish results in displacement of the reflector downslope, and the pixels associated with the reflector are plotted too close in across-track distance. Conversely, a point targeted in the downhill direction (A in Figure 42) is plotted too far from nadir because of the flat bottom assumption. For the purposes of accurate sample locations for our study, layover correction was applied to the sidescan using a computer program which adjusted the across-track position of pixels composing the sidescan, to SeaMARC II bathymetry (Figure 44) which had been re navigated and coregistered to SeaBeam bathymetry (Figure 45). Using the SeaMARC II bathymetry as a "map", the program mosaics the sidescan image over the bathymetry so that topographic features are positioned close to their actual plan view locations (Figure 46).
FIGURE 44. Fully corrected SeaMARC II bathymetry of Seamount 6. Contours are at 50 m intervals. Black signifies areas of no information.
FIGURE 45. Fully corrected SeaBeam bathymetry of Seamount 6 showing plotted digitized ANGUS tracks. Contours are at 100 m intervals.
FIGURE 46a. Layover-corrected, renavigated SeaMARC II sidescan with digitized ANGUS tracks superimposed. Displays both the top and bottom coregistered sidescan swaths covering our survey area. ALVIN tracks were not digitized due to transponder problems.
Figure 46b. Shows just the top swath of the survey (covered in 46a).
Data analysis can become more complicated than alluded to previously, i.e., rough surfaces appear as strong reflectors and smooth surfaces as weak reflectors. For instance, the generalization stated above is highly dependent upon regional slope; the inverse (rough surface = weak reflector and smooth surface, strong) is possible. Also, significant (up to several meters) acoustic penetration of sediments occurs at frequencies of 12 kHz or lower [Gardner et al., 1991]. The depth of this penetration depends on the frequency and angle of incidence of the signal and the physical properties of the sediment. Sonar images may, therefore, map lithologic variations obscured by a thin coating of sediment. Indeed, Mayer [1979] and Tyce et al. [1980] have demonstrated interference by shallow subbottom reflections. Another possible complication is volume reverberation. Once the sound penetrates below the water/sediment interface, it interacts with a volume of sediments and is reradiated in all directions, including back to the sidescan receivers [Stanton, 1984; Jackson et al., 1986]. The return signal would therefore be stronger than normally expected from sediment and the image would appear darker.

With this rudimentary knowledge of SeaMARC II data acquisition and analysis in hand, I now present a brief description of the area of seafloor that was surveyed for this study.
SETTING

Seamounts are ideal to the current study because they possess variable geology, variable slopes and cover only a small area. The type of geology present is highly correlated with regional slope and, therefore, with location on the seamount (i.e., base, flank, summit). Seamount 6 was selected due to an abundance of data from a variety of surveys: SeaMARC II, SeaBeam, ANGUS towed camera and ALVIN submersible. What follows is a short tour of Seamount 6, including a brief introduction, an overview of types of lithologies found at the base, flanks and summit regions of the edifice, as well as a summary of the volcanic history of the volcano. Detailed descriptions and analyses are found in Batiza et al. [1989] and Smith and Batiza [1989].

Seamount 6 is a small linear seamount group consisting of three coalesced edifices aligned parallel to the relative motion of the Cocos plate [Batiza and Vanko, 1983] (Figure 47). Located about 75 km south of the East O'Gorman Fracture Zone, it sits on crust of ~3.0 Ma (Figure 48). The middle edifice of this group (6C) is the largest (52 km³, ~1300 m high), while the east and west volcanoes (6E and 6W) are 21 km³, 750 m high and ~22 km³, 420 m high, respectively [Batiza et al., 1989]. Volcano 6E has an arcuate summit bench and a summit crater (750 m wide) which
FIGURE 47. Bathymetric map of Seamount 6 showing the ALVIN and ANGUS tracks [Batiza et al., 1989]. Contours are in hundreds of meters.
FIGURE 48. Location map of Seamount 6 [Batiza et al., 1989]. Magnetic lineations after Klitgord and Mammerickx [1982].
is open to the west. The summit of 6C is gently sloping and terraced. Batiza et al. [1989] interpret it as a filled caldera. A prominent flank rift zone on the north side of 6C trends at 340°, subparallel to the trend of the East Pacific Rise. The lower flanks of 6C are gentle and contain many small volcanic cones, up to 100 m high. Seamount 6 was surveyed by 5 ALVIN dives and 5 ANGUS camera runs (Figure 47). The geology determined from examination of bottom photographs acquired along these traverses are shown in Figures 49, 50 and 51.

The cone and lava field south of 6C, surveyed by ANGUS 222 and ALVIN 1390 (Figure 51), is composed mainly of pillow lava which forms haystacks and outcrops of bulbous and tubular pillows [Batiza et al., 1989]. Most of these outcrops, even on steep slopes, are evenly covered with dark fine-grained sediment. The individual rocks have coatings of manganese up to 2 cm thick and are weathered. Very steep slopes, which are probably fault-controlled, have in situ tubular pillows and may, in addition, possess aprons of pillow talus. Similar deposits characterize the deeper (below 2500 m) flanks elsewhere on 6C and 6E, but are almost completely covered with sediment and have greater amounts of talus [Batiza et al., 1989]. The steeper outer flanks of 6C and 6E (2500 m to 2300 m) consist of pillow tubes elongated downslope and, for the most part, are free of sediment (Figure 52). On gentler slopes, there are scattered outcrops of bulbous pillows and
FIGURE 49a. From Batiza et al. [1989], Seamount 6E geology along ANGUS and ALVIN tracks.
FIGURE 49b. From Batiza et al. [1989], profiles of the dive tracks from Figure 49a showing geology (V.E. 4X).
FIGURE 50a. Geology of Seamount 6C along ANGUS and ALVIN tracks [Batiza et al., 1989]. Symbols are the same as those of Figure 49.
FIGURE 50b. Profiles (V.E. 4X) along the dive tracks from Figure 50a [Batiza et al., 1989]. Symbols are the same.
FIGURE 51. Geology of the cone/lava field on the flanks of Seamount 6 along the ALVIN and ANGUS tracks, with profiles (V.E. 4X) [Batiza et al., 1989]. Same symbols as Figure 49.
FIGURE 52. Photo showing elongate and bulbous pillows on the south flank of Seamount 6C. ANGUS 223, 0918Z [from Batiza et al., 1989].
minor talus deposits. Between 6C and 6E, lies a saddle which is sedimented and has about 50% or less rock outcrop. Lava haystacks and small pillow ridges characterize the slopes of 6C near the saddle. The slopes of 6E in this vicinity consist mainly of pahoehoe (Figure 53) and large lobate pillows with scattered talus outcrops [Batiza et al., 1989].

Shallower than 2200 m, the summit of 6C consists mostly of hyaloclastite (Figure 54), sheet flows, submarine pahoehoe and lobate pillows. Hyaloclastite is a name used to describe many different rocks consisting mostly of basalt glass fragments. Deposits vary in shape and size and the rocks themselves often have different textures and types of glass shards, presumably depending upon magma type and mode of formation [Smith and Batiza, 1989]. On Seamount 6, most of the hyaloclastites are of hydrovolcanic origin, but some show evidence of being sedimentary in origin. The deposits are sheet-like with variable surface textures. The sheets thin at the distal ends and are often broken up into slabby fragments. Hyaloclastites may be rubbly nearer inferred vents and overlie sheeted deposits [Smith and Batiza, 1989]. Above 2000 m, the summit deposits of hyaloclastite, sheet flow, submarine pahoehoe and lobate pillow appear very young and have only 2-3 mm of manganese coating [Batiza et al., 1989]. More than ~50% outcrop is present in the summit area. These outcrops are separated by sediment ponds which become rippled at depths
FIGURE 53. Photo showing ropey pahoehoe on the gentle slopes of Seamount 6C near the saddle with 6E. ANGUS 220, 1044Z [from Batiza et al., 1989].
FIGURE 54. Photo showing tabular hyaloclastite and rubbly lava with rippled sediment on the summit terrace of Seamount 6C. ANGUS 223, 1017Z [from Batiza et al., 1989].
< ~2000 m. The summit area of 6E consists mostly of pillow flows with only rare pahoehoe and lobate pillows. A thick covering of rippled sediment blankets the summit of 6E. On both 6C and 6E, the general direction of sediment transport is to the east [Batiza et al., 1989], as shown by directional indicators such as ripples and scour marks. On 6C and 6E, sediment transport is usually downslope; however, there are some locations on 6E where sediment ripples indicate transport up gentle slopes.

The caldera walls of 6E are mostly covered with pillow talus. The wall is vertical in places, but elsewhere on gentler slopes, there are scattered outcrops of bulbous pillows which appear in situ [Batiza et al., 1989]. Talus, covered and mixed with sediment, is present at the base of the caldera wall. Sediment almost entirely covers the floor of the caldera. Pahoehoe, lobate pillows and deposits of honeycomb-textured lava that resembles spatter compose the rare outcrops [Batiza et al., 1989]. The rocks on the summit of 6E appear older than those on the summit of 6C. On 6E, rocks generally have thick (mostly 2-3 cm, up to 7 cm) coats of manganese and are weathered. Based on outcrop appearance, thickness of manganese coats and sediment thickness, Batiza et al. [1989], proposed that 6W, 6E and 6C formed about the same time. Pahoehoe preceded caldera collapse on 6E, followed by pillow eruptions at the summit bench and elsewhere. At the time of formation of the 6E caldera, 6C may also have had a caldera. The
filling of the 6C caldera may have occurred at the same
time as post caldera eruptions on 6E. However, summit
eruptions on 6C apparently continued for some time after
volcanic activity ceased on 6E.

TEXTURE FEATURES

In order to describe the sidescan data used in this
study, some knowledge of how images are interpreted is
necessary. Humans use three features to interpret
pictorial information [Haralick et al., 1973], including
spectral, textural and contextual. While spectral features
describe the average tonal variations in different bands of
the visible and infrared portions of the electromagnetic
spectrum, textural features contain information about the
spatial distribution of tonal variations within a band.
Contextual features provide information derived from the
surroundings of the area being analyzed. Context, texture
and tone are always present in an image, but there are
times when one property may dominate. Texture and tone are
most important in single-spectral band image areas such as
sidescan images. Tone describes the varying shades of grey
of pixels in an image. Texture pertains to the spatial
distribution of the grey tones composing an image. Texture
is an intrinsic property of all surfaces, containing
important information about the structural arrangement of
surfaces. It is a strongly stationary property independent
of illumination and carries useful information for discrimination purposes. Provided below is a discussion of the relative effectiveness of second-order textural statistics versus first-order statistics, as well as an introduction to the grey level co-occurrence matrix method of texture characterization by which the second-order statistics used in this study were estimated.

Multispectral remote sensing images such as Landsat consist of a scene imaged simultaneously in several bands of the electromagnetic spectrum [Schowengerdt, 1983]. For single-spectral images such as those produced by single-frequency side-looking sonars and radars, first-order statistics like mean intensity are not effective in classifying imagery. Different intensity values from a uniform reflector can be caused by several factors, including random noise, system changes and imaging geometry. Image speckle or random image noise can be caused by ambient noise, electronic noise and water column transmission anomalies [Reed, 1987]. It takes the form of isolated pixels with intensities uncorrelated to their neighbors. Gain changes applied during acquisition of data, required by variations in signal strength due to changes in water depth or reflectivity of a lithologically uniform surface, destroy any 1:1 relationship there might have been between intensity and target strength [Reed, 1987]. And finally, variations in backscatter strength occur along with changes in angle of incidence of the
outgoing acoustic signal, which are not completely rectified by angle varying gain corrections. Therefore, looking at texture, an abstract second-order statistic feature which relies on the spatial relationships between tonal variations and is independent of image intensity (though it is inter-related), is more useful in classifying single spectral imagery such as sidescan data.

Subjectively, texture recognition is easy for the human observer, but difficult to quantify in the objective sense. Of the many proposed methods for extracting statistical texture signatures for image analysis [Pratt, 1978; Laws, 1980; Haralick et al., 1973], we use the grey level difference matrix method which is closely related to the grey level co-occurrence matrix method of Haralick et al. [1973]. The use of this type of texture characterization is justified by results of experiments in human texture perception which indicate that second-order probabilities of the form measured by co-occurrence matrices are important in texture discrimination [Julesz, 1962; Julesz et al., 1973; Gagalowicz, 1980]. What follows is a description of the grey level co-occurrence matrix method of Haralick et al. [1973].

The grey level co-occurrence matrix method works under the assumption that texture information in an image is contained in the overall or average spatial relationship which the grey tones in the image have to one another [Haralick et al., 1973]. A set of grey-tone spatial-
dependence probability-distribution matrices are created for a given image block. These matrices express the relative frequencies $P_{ij}$ with which two adjacent pixels, separated by distance $d$, occur on the image, one having grey tone $i$, the other $j$ [Haralick et al., 1973]. Moments evaluated from these joint probability matrices provide quantitative estimates of image texture.

Digital images are stored in the computer as a 2 dimensional array of resolution cells, or pixels. Each pixel is assigned some grey tone value $N_g$, $N$ being the number of grey levels possible. Each grey level co-occurrence matrix is a square matrix with dimension $N$. The entries in this matrix, $S(i,j;\theta,d)$, are the number of times a pixel of intensity $i$ has a neighboring pixel of intensity $j$ in the direction $\theta$, at a distance of $d$. As an example, say the image is represented by

\[
\begin{array}{cccc}
1 & 1 & 2 & 3 \\
1 & 2 & 2 & 3 \\
2 & 3 & 3 & 4 \\
1 & 2 & 4 & 4 \\
1 & 1 & 3 & 3 \\
\end{array}
\]

In the horizontal direction ($\theta = 0^\circ$), at a lag or distance of 1, the grey level co-occurrence matrix is

\[
\begin{array}{cccc}
1 & 2 & 3 & 4 \\
1: 4 & 3 & 1 & 0 \\
2: 3 & 2 & 3 & 1 \\
3: 1 & 3 & 8 & 2 \\
4: 0 & 1 & 2 & 6 \\
\end{array}
\]

[Reed, 1987]. In the case of our grey level difference matrices, the entries in the matrices express the
probability that two pixels separated by distance \( d \), in the direction \( \theta \), will differ in intensity by \( X \) amount. The elements of the matrix are normalized by dividing each entry by the total number of possible entries for that direction \( \theta \) and distance \( d \). The matrices are evaluated at \( \theta = 0^\circ, 45^\circ, 90^\circ \) and \( 135^\circ \) (Figure 55). To avoid any angular dependence upon the texture signature, Haralick et al. [1973] suggested taking the average and the range of the statistics which are extracted at each angle \( \theta \).

Haralick et al. [1973] discussed 14 texture feature vectors which can be extracted from the grey level co-occurrence matrices. As many of these features are strongly correlated to each other, we use only 6 plus isotropy, devised by Reed and Hussong [1989]. The texture features we examine include: contrast, measuring image contrast; inverse difference moment, measuring local homogeneity of the image; entropy, a measure of image complexity or how structured the image is; energy, measuring image busyness or coarseness; mean; correlation, a measure of linearity of the image; and isotropy, measuring the rotational invariance of the image. Appendix A contains formulas and explanations for each of the texture features mentioned.

A texture feature vector, composed of the seven second-order statistics listed above (contrast, entropy, etc.) is used to define each sample (i.e., numbered boxes in Figure 43). A sample is a group of pixels from a SeaMARC II
FIGURE 55. Diagram showing the angles of $\theta$ at which the grey level co-occurrence matrices are evaluated [from Haralick et al., 1973]. Pixels 1 and 5 are the nearest neighbors of the * pixel at $0^\circ$ (horizontal). Pixels 4 and 8 are nearest neighbors at $45^\circ$. 3 and 7 are nearest neighbors at $90^\circ$, and 2 and 6 at $135^\circ$. This information is purely spatial and has nothing to do with grey levels.
sidescan image chosen as a representative area of a specific lithology. The following section describes how the feature vectors are used to classify samples into distinct groups.

CLASSIFICATION

The goal of classification is to attempt to assign the samples, each defined by a texture feature vector (composed of the seven second-order statistics), into distinct classes. This can be achieved through unsupervised or supervised classification schemes. An unsupervised classification is done by clustering. The feature vectors from each sample are given to a clustering routine which determines natural groupings of the data. In a supervised classification, samples are assigned to classes according to supplementary knowledge, such as that obtained of lithology from the examination of bottom photographs. In this way, classes are defined by representative, homogeneous samples for each class (in this study: pillow, sediment, hyaloclastite, sheet flow and talus).

The clustering algorithm used in this study for an unsupervised classification is Hartigan's [1975] K-means algorithm. The algorithm is initialized by assuming there are 1, 2,...K classes, K specified by the operator. K number of initial mean vectors are chosen either randomly,
assigned by the operator or defined by the data. This study allows the data to define the initial mean vectors. This is achieved by initially assigning each sample area to a class (1, 2, ..., K), done one of two ways, either cyclically or by weighted means. This is specified in a parameter file which is read by the clustering routine. The cyclic method will assign the first K samples to classes 1 through K, then the next K samples to classes 1 through K, and so on. The method of weighted means sums all the components of the feature vector for each sample, sorts them from highest to lowest and assigns the first X (total sample number/K) samples to class 1, the next X samples to class 2, and so on to class K. The clustering routine then calculates the means of the feature vectors for each of the assigned K classes. This mean vector for each K defines the centroid of that K class. The routine assigns each sample to that cluster from whose centroid its feature vector is least distant. Once all the samples are assigned to a cluster, the cluster means are recomputed, and the samples reassigned on a minimum distance basis. The process continues until there is no significant change in cluster assignment from one iteration to the next. In this way, each sample is classified, according to its second-order statistics, to the group to whose members it is most similar.

A supervised classification can be done using groundtruth information (bottom photographs) which we have
from ANGUS towed camera and ALVIN submersible dives. I assigned each sample area to a specific lithologic class (pillow, talus, hyaloclastite, sediment or sheet flow) based on my observations of the bottom photographs. The average feature vector and standard deviation for each class can then be calculated from the representative samples.

**METHODS**

Now that we have the necessary background information in hand, I discuss in this section the methods applied in this study. Specifically, I describe the preparation of SeaBeam and SeaMARC II data sets for analysis, the process involved in the selection of sample areas and how sample boxes were "picked" via a computer program designed to calculate texture statistics. I also include a brief review of preliminary results from the application of our technique which leads into a discussion of our final results.

Accurate location of sample areas for analysis required the coregistration of SeaBeam and SeaMARC II data. The logged SeaBeam data from Scripps contained shipboard navigation, which upon plotting showed significant offsets. To correct this, software was developed to play back the SeaBeam mosaic which was adjusted, by hand, to fit the bathymetric map from Batiza et al. [1989] (Figure 47),
assumed to be correctly navigated and adjusted. SeaMARC II navigation was readjusted by optimizing the fit, by hand, of SeaMARC II bathymetry (Figure 44) with the readjusted SeaBeam mosaic in Figure 45. SeaMARC II navigation was then replotted and merged with SeaBeam navigation (Figure 56).

The ANGUS camera tracks and ALVIN dive tracks had to be adjusted to the new coregistered data. These transponder-navigated tracks were previously plotted on SeaBeam bathymetry (Figure 47). In order to adjust the tracks, time ticks were added at 5 minute intervals and then digitized. There had been transponder problems on the ALVIN dives, so only the ANGUS tracks were digitized and plotted on the re-navigated, layover-corrected SeaMARC II sidescan (Figure 46). Figure 45 shows the dive tracks plotted on the mosaicked SeaBeam bathymetry. This allows for more accurate placement of sample boxes.

Despite all our careful efforts in coregistering and renavigator SeaBeam and SeaMARC II data, navigational uncertainties may still exist. These uncertainties may be of key importance and, at least, should be kept in mind. It is possible that due to navigational uncertainty, a patch of seafloor we see in the bottom photographs, does not correspond precisely with a patch of seafloor chosen to sample on the SeaMARC II sidescan image. Misplacement of sample boxes may also arise in the translation of the sample box locations from the layover-corrected sidescan
FIGURE 56. Renavigated, coregistered SeaMARC II and SeaBeam bathymetry at 50 m contour intervals. Black signifies areas of no information.
(Figure 46) to the less processed sidescan record actually used for the textural analysis (Figure 57).

With the SeaMARC II sidescan data prepared, a short feasibility study on the ability of SeaMARC II to distinguish among various lithologies was begun. In order to proceed, an assessment of groundtruth data (bottom photographs) was necessary in choosing viable sample locations. I meticulously examined the films and photographs from 5 ANGUS camera tows and 5 ALVIN dives and evaluated them for possible sample area locations. Geologic maps (derived from observations of these same films by Batiza et al. [1989]) along each track (Figures 49, 50, 51) provided a useful guide to pinpointing specific areas for examination. Sixteen sample locations were initially chosen (Figure 43), four of each in the following lithologies: sediment, pillow basalt, hyaloclastite and talus. Criteria for choosing these areas were: homogeneity of lithology, suitably extensive regions, similar bottom slope gradients for each lithologic group and similar look direction, or azimuth at which the area is viewed. "Suitably extensive regions" were determined by assuming a sample box size of approximately 200 pixels (in SeaMARC data), a size sufficiently large to generate useful statistics yet small enough to prevent blurring of any possible texture signature. This size box corresponds to about a 200 m² area. Locations along the dive tracks with lithologies extending this distance were chosen. Slope
FIGURE 57. Minimally processed SeaMARC II sidescan showing the locations of the refined set of 112 sample boxes. a) The first 72 samples boxes from the bottom swath of the survey are shown here. b) and c) display the samples from the top swath of the survey, viewed from a different look direction. Great pains were taken for the accurate placement of boxes and a uniform size of about 160-200 pixels. Descriptions, first and second-order statistics for each box are shown in Table 1.
FIGURE 57b. See caption for Figure 57.
FIGURE 57c. See caption for Figure 57.
gradient measurements were obtained from Figure 58, a slope map derived from the composite SeaBeam/SeaMARC II bathymetric data. Sample boxes were picked using a textural analysis program, developed by Tom Reed at the Hawaii Institute of Geophysics, called "get_tex". This program displays the minimally-processed SeaMARC II sidescan image (mentioned in the SeaMARC II section) and allows boxes to be picked using the mouse. Second-order image statistics for the area within each sample box are then generated by the grey level co-occurrence matrix method discussed in the Texture Features section.

Plots of the means and ranges of each lithology (Figure 59) showed that our estimates of image texture from SeaMARC II are apparently capable of discriminating among several seafloor lithologies. This preliminary study determined that SeaMARC II could distinguish differences in regions that camera data portrayed as pillow basalt, hyaloclastite and sediment/talus. The acoustic similarity between sediment and talus could be due to thin layers of sediment overlying talus. Acoustic penetration of the thin sediment layer would return the characteristics of talus. Another possible explanation is that at the scale of 12 cm (the wavelength of insonification), talus presenting facets 1 to 10 m in major dimension would appear equally as rough/smooth as flat-lying sediments. Signal amplitude from the talus would be greater, but the coherence might be the same as that of a pond of sediment. As texture,
FIGURE 58. Digital slope magnitude map of Seamount 6 derived from the composite SeaBeam/SeaMARC II bathymetric data. Subtle color changes are at 2° intervals of slope angle.
FIGURE 59. Plots of means and ranges of second-order statistics generated from the sample boxes in Figure 43. Plotted values for pillows = the mean of boxes 9, 10, 11, 12; for sediment = mean of boxes 1, 2, 3, 4; for hyaloclastite = mean of boxes 5, 6, 7, 8; for talus = mean of boxes 13, 14, 15, 16. Note that second-order statistics apparently can clearly distinguish between some lithologies, such as pillows and hyaloclastite.
expressed by our grey level co-occurrence matrix features, is largely independent of image amplitude, these two lithologies might be difficult to separate texturally.

The next step was to test the system with more variable sample areas. 142 locations were chosen, varying in lithology, bottom roughness, bottom slope gradient, and look angle. Due to a turn in our SeaMARC II survey data (Figure 46a), a small portion of Seamount 6 is viewed from two different look directions. 50 of these 142 samples are duplicate locations viewed from a different look direction (look angle will also be different, as this is calculated from water depth and distance from nadir). Roughness was estimated from the films and photographs according to a scale I devised based on geology (Appendix B). Descriptive variables along with first and second-order image statistics, generated by the updated "ss_get" program (which operates the same as "get_tex" but has some added features and was developed by Tom Reed at the Hawaii Institute of Geophysics), for each sample area were evaluated by using clustering routines based on Hartigan's [1975] K-means algorithm. Initial attempts at clustering the data were not terribly encouraging. Distinct classes did not emerge from the data.

Upon further review of the ALVIN and ANGUS films and photographs, I eliminated several sample locations as being ambiguous in lithology, having too low a percentage of the intended lithology, or being too limited in lateral extent.
In addition, it was determined that some sample locations were mislocated. The locations of the refined set of 112 sample boxes are shown in Figure 57. Descriptive variables and first and second-order statistics for each box are summarized in Table 1. Details about each entry in Table 1 are in Appendix C. The next section is a discussion of the results obtained in endeavors at classification of the data from this set of samples.

RESULTS AND DISCUSSION

Supervised and unsupervised classifications of the data from the refined set of 112 samples were attempted. Unsupervised classification took the form of several clustering scenarios changing certain variables. Variables in these permutations include the method of assigning initial centroids (discussed previously in the Classification section), number of samples and number of classes. For a supervised classification, ALVIN and ANGUS dive films and photographs were used to assign each sample area to a specific lithologic class, as discussed in the Classification section.

A supervised classification of the 112 samples resulted in the plots shown in Figure 60. Stars in the plots are the means of the second-order statistics while the error bars represent standard deviation. As can be seen, no
TABLE 1. Sample location descriptors. Long=longitude, Lat=latitude, Mean I=mean intensity, MAD=mean absolute deviation, Std=standard deviation, Var=variance, Skew=skewness, Kurt=kurtosis, Rough=roughness, IDM=inverse difference moment, Corr=correlation. Each entry is explained in Appendix C.
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**TABLE 1. (Continued) Sample location descriptors.**
FIGURE 60. Plots of second-order texture statistics for a supervised classification of all 112 samples. PIL is pillow, SED is sediment, TAL is talus, HYL is hyaloclastite, and SHT is sheet flow. *'s show the mean for each class and |'s are standard deviation. Notice that no distinct groups separate out.
distinct groups are evident. Eliminating samples 73-112, representing a different look direction, produced the plots in Figure 61. Apparently, look direction doesn't matter, the results are the same: the means of the 5 classes determined by bottom photos are not statistically distinct.

Since the supervised classification failed to produce distinct groups, clustering was attempted to determine if the data would form natural groupings. Following is a recap of the results from several clustering attempts. Initially, all 112 samples were clustered using both initial centroid assignment methods and specifying 5-10 classes. Figure 62 shows results for 5 classes using the cyclic pattern method of initial centroid assignment. Note that for the most part, the standard deviations overlap one another. However, there is some separation. Class 5 stands out in contrast and mean. Class 1 stands out in mean, energy and entropy. Class 5 consists of 7 members and has a composition of 43% pillow, 29% talus, 14% sheet flow and 14% sediment. Class 1 has 16 members: 44% hyaloclastite, 44% sediment and 12% pillow. The remaining 3 classes have 27 members or more and are variable in composition.

Figure 63 shows the results from the same set of samples using the mean weighted method of initial cluster assignment. Cluster membership changes to a certain extent, but patterns are similar. Class 5 is a tighter cluster here, consisting of 2 talus samples. It is
FIGURE 61. Plots of second-order texture statistics for a supervised classification of just the first 72 samples. Symbols and classes are the same as in Figure 60.
FIGURE 62. Plots of second-order texture statistics for clustering done with 112 samples, cyclic pattern method of initial centroid assignment and 5 classes specified. Symbols are the same as in Figure 60.
FIGURE 63. Plots of second-order texture statistics for clustering done with 112 samples, mean weighted method and 5 classes specified. Symbols are the same as in Figure 60.
distinct in contrast, isotropy, correlation and mean. Class 4, with 9 members (44\% pillow, 33\% sediment, 12\% talus and 11\% sheet flow), stands out in contrast and mean. Class 1 in this clustering is essentially the same as class 1 in the previous clustering: 44\% sediment, 38\% hyaloclastite, 18\% pillow and has 18 members. This class stands out in mean, energy, entropy and IDM. The other 2 classes have 41 and 42 members and are essentially random in composition.

Though there are slight differences in the way these 2 methods of initial centroid assignment cluster, the trends we see are basically the same. Significantly pure groups of the 5 lithologies do not cluster naturally. Groups consist mainly of mixes of a) sediment and hyaloclastite and b) pillow and talus. Many sediment samples cluster in groups composed mainly of pillow plus talus. These particular samples are mimicking pillow/talus and are not being recognized as sediment, probably due to one or both of the reasons discussed earlier in the Methods section.

As more and more classes are added, cluster membership changes in specifics, but basic trends remain the same. Sediment and hyaloclastite continue to cluster together as does pillow and talus. Purity of classes does improve, because the large mixed classes are being split into smaller ones. Eventually, every cluster could be pure, but consisting only of a single member each, if 112 classes are specified. When the results from greater number of classes
are plotted, however, their separability does not significantly improve. As shown in Figure 64, clustered using 8 classes and the cyclic pattern method of initial centroid assignment, class 5 is the only one that stands reasonably well on its own. It is composed of equal parts pillow and talus (33% each) and equal parts sheet flow and sediment (17% each). Looking at the members of the other clusters, we have class 1 with 70% hyaloclastite, 20% sediment and 10% pillow, and class 7 with 66% sediment, 11% hyaloclastite and 13% pillow. The plots of classes 1 and 7 do not completely separate from each other, but viewed as one they are distinct in mean, energy, entropy and IDM.

Similarities are seen using the mean weighted method of initial centroid assignment with 8 classes specified (Figure 65). Classes 6 and 8 stand out in contrast, isotropy, correlation and mean. Class 6 is 50% pillow, 25% sheet flow and 25% sediment and has 4 members. Class 8 has 2 talus members. Class 1 is 50% hyaloclastite and 36% sediment (14 members); while class 7 has 15 members composed of 66% sediment and 20% hyaloclastite. Class 1 is distinguishable in every statistic, overlapping class 7 only in energy and IDM.

No pure clusters of more than 2 members emerge until 20 classes are specified using the cyclic pattern method of initial centroid assignment. Class 18 in Figure 66 is made up of 3 hyaloclastite members. Class 4 is 68% hyaloclastite and 16% sediment. These classes, while not
FIGURE 64. Plots of second-order texture statistics for clustering done with 112 samples, cyclic pattern method and 8 classes specified. Symbols are the same as in Figure 60.
FIGURE 65. Plots of second-order texture statistics for clustering done with 112 samples, the mean weighted method of initial centroid assignment and 8 classes specified. Symbols are the same as in Figure 60.
FIGURE 66. Plots of second-order texture statistics for clustering done with 112 samples, the cyclic pattern method and 20 classes specified. Symbols are the same as in Figure 60.
distinguishable individually (as class 18 in contrast and mean), are distinct collectively, except in IDM.

Attempting to ascertain if samples 73-112 (from the second look direction) were clouding the statistics, I eliminated them from the clustering routine. The remaining 72 samples were put through the same treatment as before. The only significant difference this produced was to make the clusters smaller. Clustering trends remain the same, creating sediment/hyaloclastite and pillow/talus groups. As an example, see Figure 67, representing the cyclic pattern method of initial centroid assignment with 5 classes specified. In Figure 67, class 2 is distinct in contrast, isotropy, correlation and mean. It is made up of 45% hyaloclastite, 36% sediment and the rest pillow. It is most closely related to class 1, consisting of 60% sediment, 20% hyaloclastite and 20% pillow. Classes 3, 4 and 5 are predominantly pillow/talus mixes.

Next, I went back to the 112 samples and eliminated all that were not composed of at least 80% of the intended geology in an attempt to ascertain if "cleaner" samples would improve the purity of the clusters. This completely removed the sheet flow samples. The remaining 97 samples (4 lithologies) were clustered into 5 classes, thus allowing for a "misfit" class. Assuming that the largest cluster with a potpourri of geology is the aforementioned misfit cluster, it would be class 4 with 36 members (Figure 68). Class 1 is equal amounts of pillow and talus with
FIGURE 67. Plots of second-order texture statistics for clustering done with 72 samples, cyclic pattern method and 5 classes specified. Symbols are the same as in Figure 60.
FIGURE 68. Plots of second-order texture statistics for clustering done with 97 samples representing 80% pure lithology, the cyclic pattern method and 5 classes specified. Symbols are the same as in Figure 60.
sediment. Class 2 is predominantly hyaloclastite (54%) with 38% sediment. Class 3, like class 1, has an equal mix of pillow and talus again with sediment. Class 5 is mostly sediment (56%) with 22% hyaloclastite.

Class 2, mainly hyaloclastite, is the only one to stand out (Figure 68) in entropy, mean and contrast. It overlaps class 5, which is predominantly sediment, in the other statistics. Classes 1 and 3 have essentially the same composition, but are distinct from each other in the plots of their statistics.

The same pattern is seen until 9 classes are specified. At this point, a cluster of 86% sediment emerges having 7 members. Unfortunately, its plot is not distinct (class 7 in Figure 69). It overlaps most consistently with class 1 (70% hyaloclastite and 20% sediment).

At 15 classes, a cluster of 3 hyaloclastites (class 6 in Figure 70) and one of 4 sediments (class 10) are the first pure clusters to emerge. The hyaloclastite group (class 6) can be distinguished from the sediment class (10) in all statistics except correlation. Only the hyaloclastite group is distinct from all others, in contrast and mean.

This leads to the conclusion that hyaloclastite and sediment, as well as pillow and talus, are texturally similar. Eventually, with enough classes specified, some hyaloclastite and sediment samples do separate into different clusters, but are not necessarily distinguishable
FIGURE 69. Plots of second-order texture statistics for clustering done with 97 samples representing 80% pure lithology, the cyclic pattern method and 9 classes specified. Symbols are the same as in Figure 60.
FIGURE 70. Plots of second-order texture statistics for clustering done with 97 samples representing 80% pure lithology, the cyclic pattern method and 15 classes specified. Symbols are the same as in Figure 60.
from each other. However, only hyaloclastite is distinct from all other groups, in some cases.

Not to lose sight of our original purpose, to assess the affect of lithology, roughness, slope angle and look angle on texture signature, I performed the following experiment. First, since the measure of roughness, in this study, is determined by a scale based on the lithology rather than a quantitative measurement, it can be discarded from this discussion. I grouped the 97 samples (80% pure) into the two look directions (Figure 71). These 2 groups were separated into 4 sub-groups of slope angles: 0-10°, 10-20°, 20-30°, 30-40°. Each of these 4 groups was further subdivided into look angle ranges: 0-10°, 10-20°, 20-30°, 30-40°, 40-50°, 50-60°, 60-70°. Lithologies in each of the look angle groups is rather arbitrary. However, slope angles of 0-20° are predominantly sediment and hyaloclastite (71% combined), while angles of 20-40° are mostly pillow and talus (70% combined).

This scheme produced 19 groups with at least 1 member apiece for the look direction on the bottom swath of the survey, and 18 groups for the look direction represented by the top swath. The means and standard deviation for each statistic are shown in Figure 72 and 73, respectively. No distinct separation is in evidence.

One further grouping scheme (Figure 74) has the 97 80% pure samples again divided first into the 2 look directions. These 2 groups were subdivided into the 4
FIGURE 71. Diagram showing a grouping scheme employed to assess the effects of look direction, look angle, slope angle on the classification of the data. The samples were divided into the two look directions. These two groups were then divided into one of 4 slope angle ranges. Each of these groups was further subdivided into one of 7 look angle ranges.
FIGURE 72. Plots of second-order texture statistics showing the results of the grouping scheme in Figure 71. These plots display data from the look direction represented by the southern swath of the SeaMARC II survey (Figure 57a). Symbols are the same as in Figure 60.
FIGURE 73. Plots of second-order texture statistics showing the results of the grouping scheme in Figure 71. These plots display data from the look direction represented by the northern swath of the SeaMARC II survey (Figures 57b and c). Symbols are the same as in Figure 60.
FIGURE 74. Diagram showing a second grouping scheme similar to that in Figure 71. This scheme divides the samples into the two look directions. These two groups are then divided into one of the 4 lithologic groups, then further subdivided into one of the 4 slope angle ranges.
lithologies: pillow, talus, hyaloclastite and sediment. These 4 groups are further subdivided into 4 slope angle ranges: 0-10°, 10-20°, 20-30°, 30-40°. Look direction 1 with 62 samples has 14 subsets (Figure 75), while look direction 2 with 35 samples has 12 (Figure 76). Again, no distinct groups emerge. Using this method of analysis, the conclusion is that look direction, look angle and slope angle have no recognizable affect on the second-order texture statistics.

A question arose at the conclusion of this study which challenged the value of our results. Particularly, do the texture statistics derived from the SeaMARC II sidescan image tell us anything different from what the eye can detect from the same image? The results of this study show that mixed groups of hyaloclastite/sediment and pillow/talus are distinguishable by SeaMARC II based on estimates of texture statistics. By simply looking at a SeaMARC II sidescan image, large patches of rock and sediment can be distinguished from each other. As explained in the SeaMARC II section, rough surfaces, such as those composed of rock, generally appear as dark areas on a SeaMARC sidescan image; while smooth surfaces, such as sediment, appear light.

In order to answer the question stated above, I examined each sample box on the SeaMARC II sidescan (Figure 57) from the set of 97 80% pure samples. I assembled the samples for each lithology which were located in large, homogeneous
FIGURE 75. Plots of second-order texture statistics showing the results from the grouping scheme in Figure 74. These plots display data from the look direction represented by the southern swath of the SeaMARC II survey (Figure 57a). Symbols are the same as in Figure 60. Note some groups were empty, so no plot exists.
FIGURE 76. Plots of second-order texture statistics showing the results from the grouping scheme in Figure 74. These plots display data from the look direction represented by the top swath of the SeaMARC II survey (Figures 57b and c). Symbols are the same as in Figure 60.
patches, according to my observations. Since mean intensity is really what the eye detects in a homogeneous area, I plotted the mean and standard deviation of the mean intensities for the samples in each lithologic group (pillow, sediment, talus, hyaloclastite) (Figure 77). As shown in Figure 77, the means for each lithologic group are unique; however, the standard deviations for each group overlap a great deal. Therefore, mean intensity cannot, by itself, define distinct groups of lithology, whereas some second-order texture statistics can. It is interesting to note in Figure 77 the similarity in means and standard deviations for hyaloclastite/sediment and pillow/talus. These were the two mixed groups that could, largely, be separated by their texture statistics.

CONCLUSIONS

In conclusion, SeaMARC II can distinguish hyaloclastite/sediment from pillow/talus acoustically. Second-order statistics for each of these mixed groups are distinct regardless of look direction, look angle and slope angle. Hyaloclastite is the only lithology that is completely distinct from the other lithologies in this study.

Several possible explanations for the lack of any striking distinctions between classes exist. The first is
FIGURE 77. Plot of mean ('s) and standard deviation (bars) of mean intensity for each lithologic group, using the set of 97 80% pure samples. HYL is hyaloclastite; TAL is talus; SED is sediment and PIL is pillow. Note that the means are unique, but standard deviations overlap a great deal. The first-order statistic of mean intensity does not help in distinguishing the lithologic groups.
that the feature vectors (the second-order statistics) utilized in this study are not sufficient to capture the information in the data. Secondly, we may have missed our intended geology and are not sampling what we think we are sampling. A third explanation is that the lithologies are not, in fact, acoustically distinct. The fact that hyaloclastite/sediment and pillow/talus classification combinations are common seems to indicate that this is the case, in as far as hyaloclastite and pillow lava are not acoustically distinct from sediment and talus, respectively; yet the combinations are distinct from each other. And finally, the ground truth leaves much to be desired, in that the bottom photographs provide such a limited view of what the geology is to either side of the dive track that they are not sufficient to estimate the homogeneity of a patch of seafloor.
The texture feature vectors used in this study are extracted from a grey level difference co-occurrence matrix, as explained in the text (Texture Features), according to the following formulas. These equations are as taken from Reed and Hussong [1989] unless otherwise indicated.

**CONTRAST**

This feature measures image contrast. Image contrast is related to the range of grey levels in an image. The greater this range is, the greater the contrast [Schowengerdt, 1983]. It is calculated by

\[
\text{CONTRAST} = \sum_{n=0}^{N_g-1} n^2 \left( \sum_{i=1}^{N_g} \sum_{j=1}^{N_g} S(i,j) / R \right)_{|i-j|=n}
\]

where \( N_g \) is the number of distinct grey levels in the image and \( R \) is the number of neighboring pixel pairs. An example of how contrast varies between two simple images can be seen in Figure 78.

**INVERSE DIFFERENCE MOMENT (IDM)**

This feature measures local homogeneity. In a homogeneous image (box 15, Figure 57a), there are few
FIGURE 78. Figure showing two different land-use category images and a comparison of some of their texture features: ASM (angular second moment) or the same as Energy, Contrast and Correlation [from Haralick et al., 1973]. a) An image of grassland, representing a heterogeneous image with a wide range of grey levels. b) An image of a body of water, representing a homogenous image.

<table>
<thead>
<tr>
<th>Angle</th>
<th>ASM</th>
<th>Contrast</th>
<th>Correlation</th>
<th>Grassland</th>
<th>ASM</th>
<th>Contrast</th>
<th>Correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°</td>
<td>.0128</td>
<td>3.048</td>
<td>.8075</td>
<td></td>
<td>.1016</td>
<td>2.153</td>
<td>.7254</td>
</tr>
<tr>
<td>45°</td>
<td>.0080</td>
<td>4.011</td>
<td>.6366</td>
<td></td>
<td>.0771</td>
<td>3.057</td>
<td>.4768</td>
</tr>
<tr>
<td>90°</td>
<td>.0077</td>
<td>4.014</td>
<td>.5987</td>
<td></td>
<td>.0762</td>
<td>3.113</td>
<td>.4646</td>
</tr>
<tr>
<td>135°</td>
<td>.0064</td>
<td>4.709</td>
<td>.4610</td>
<td></td>
<td>.0741</td>
<td>3.129</td>
<td>.4650</td>
</tr>
<tr>
<td>Avg.</td>
<td>.0087</td>
<td>3.945</td>
<td>.6259</td>
<td></td>
<td>.0822</td>
<td>2.863</td>
<td>.5327</td>
</tr>
</tbody>
</table>
dominant grey level transitions. As a result, the co-
ocurrence matrix for such an image will have small
entries. Therefore, IDM values for homogeneous areas will
be larger than in non-homogeneous areas. Compare IDM
values in Table 1 for boxes 15 (0.2180) and 11 (0.0585), a
heterogeneous area, in Figure 57a. The calculation is

$$\text{Ng} = \sum_{i=1}^{N_g} \sum_{j=1}^{N_g} \frac{S(i,j)}{((1 + (i-j)^2) \times R}$$

where Ng is the number of distinct grey levels in the image
and R is the number of neighboring pixel pairs.

**ENTROPY**

Entropy is a measure of image structure and complexity.
This value is largest when all the pixel pairs occur with
equal frequency, as they would in a perfectly random image.
The value decreases as image structure increases [Reed,
1987]. It is calculated by the expression

$$\text{Entropy} = -\sum_{i=1}^{N_g} \sum_{j=1}^{N_g} \frac{S(i,j)}{R} \times \log \left( \frac{S(i,j)}{R} \right)$$

where Ng is the number of distinct grey levels and R is the
number of neighboring pixel pairs.
ENERGY

This is a measure of image energy or coarseness. The coarser an image, or more heterogeneous it is, the smaller the energy value will be. In Figure 78b, there are not many dominant grey level transitions, so the co-occurrence matrix will have few large magnitude entries. In Figure 78a, the matrix will have a large number of small entries. Since the energy feature is a square of the sums of the entries, the energy value will be smaller for coarser images [Haralick et al., 1973]. Energy is

\[
\text{Energy} = \sum_{i=1}^{N_g} \sum_{j=1}^{N_g} \left( \frac{S(i,j)}{R} \right)^2
\]

MEAN

This simply is the sum average of the entries in the grey level difference co-occurrence matrix. Values are higher for images with greater differences in grey levels. Mean is

\[
\text{Mean} = \sum_{i=0}^{N_g-1} i \times \left( \frac{S(i,j)}{R} \right)
\]

where \( N_g \) is the number of grey levels and \( R \) is the number of neighboring pixel pairs.
CORRELATION

A measure of linearity of an image and is highest in images with linear dependencies, as shown in Figure 78. The formula is

\[
\frac{\sum_{i=1}^{\text{Ng}} \sum_{j=1}^{\text{Ng}} (i,j) S(i,j)}{R - \mu_x \mu_y \frac{\sigma_x \sigma_y}{}}
\]

where \( \text{Ng} \) is the number of distinct grey levels, \( R \) is the number of neighboring pixel pairs and \( \mu_x, \mu_y, \sigma_x, \sigma_y \) are the means and standard deviations for row and column sums, respectively.

ISOTROPY

This is a measure of the isotropy of the image; isotropy being directional uniformity of properties. This is calculated as the sum of the differences of the grey level difference co-occurrence matrices for angles of \( \theta = 0^\circ \) and \( 90^\circ \) [Reed and Hussong, 1989] (Figure 55).
APPENDIX B. Keys to codes and scales

KEY TO VOLCANO CODE

The volcano code consists of 3 characters, only one of which is a non-zero number. The placement of the non-zero indicates the region of the Seamount 6 where the sample area is located.

```
  0 1 0
  6W (west) 6C (central) 6E (east)
```

The non-zero numbers refer to where the sample area is located on the specified edifice:

- 0 not present
- 1 base
- 2 flank
- 3 near summit
- 4 summit

For example, box 1 (Figure 57a) is located on the flank of Seamount 6C, the central edifice. Therefore, its volcano code is 020. Similarly, box 5 has the volcano code 004 (Table 1), placing it on the summit of Seamount 6E, the easternmost edifice.
KEY TO THE GEOLOGY DESCRIPTORS

Each geology descriptor consists of 10 characters, or 5 pairs of characters. Each character pair refers to a particular lithology, as shown below.

```
00|00|00|00|00
pillow sediment talus hyaloclastite sheet flow
```

The first number of each character pair refers to the percentage of the specified lithology present: 0=not present, 1=1-10%, 2=11-20%, 3=21-30%, 4=31-40%, 5=41-50%, 6=51-60%, 7=61-70%, 8=71-80%, 9=81-90%, A=91-100%.

The second number of the pair provides a more detailed description of that particular lithology:

<table>
<thead>
<tr>
<th>PILLOW</th>
<th>SEDIMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 not present</td>
<td>0 not present</td>
</tr>
<tr>
<td>1 well sorted</td>
<td>1 rippled</td>
</tr>
<tr>
<td>2 moderately well-sorted</td>
<td>2 mounded, pocked/cratered</td>
</tr>
<tr>
<td>3 poorly sorted</td>
<td>3 bioturbated</td>
</tr>
<tr>
<td>4 encrusted</td>
<td>4 lag deposits/cobbles</td>
</tr>
<tr>
<td>5 slightly dusted</td>
<td>5 3 and 4</td>
</tr>
<tr>
<td>6 1 and 4</td>
<td>6 2 and 3</td>
</tr>
<tr>
<td>7 2 and 4</td>
<td>7 1 and 2</td>
</tr>
<tr>
<td>8 3 and 4</td>
<td>8 1 and 3</td>
</tr>
<tr>
<td>9 1 and 5</td>
<td>9 1 and 4</td>
</tr>
</tbody>
</table>
A 2 and 5
B 3 and 5

A 1 and 5
B 1 and 6
C 1, 2, 3, 4
D 2 and 4
E 2 and 5
F as coating/trace amount

TALUS
same descriptors as for pillows

HYALOCLASTITE
0 not present
1 slabby/crusty
2 nodular/hackly
3 1 with some nodes/rubbly
4 1 and 2 mix

SHEET FLOW
0 not present
1 slabby
2 nodular/hackly
3 ropey/hummocky
4 lobes
5 1 and 2
6 1 and 4
7 2 and 4

As an example, the code 8221000000 describes a sample location with 71-80% moderately well-sorted pillows and 11-20% rippled sediment. Similarly, the descriptor 00A9000000
refers to an area of 100% sediment which is rippled and has lag deposits and cobbles.

KEY TO ROUGHNESS SCALE
This scale is an attempt to quantify roughness. It is based not on quantifying the roughness of the terrane by, for example, digitizing photos, but rather on an estimate of relative roughness assuming a scale. It is based on the lithologies observed in the bottom photographs. Roughness decreases down-scale.

10 very rough, large-looking pillow/talus
9 mixed pillow, talus (not well-sorted), sediment (small percentage)
8 well-sorted talus
7 well-sorted pillows
6 egg-like pillow nodules
5 pahoehoe/sheet flow
4 hyaloclastite crust w/larger scale features
3 sediment w/ripples, mounded sediment, hyaloclastite crust
2 sediment w/lag deposits, worm holes
1 smooth sediment
APPENDIX C. Key to Table 1

Sample # corresponds to the numbered boxes on the SeaMARC II images in Figure 57.
Location (key'd to dive) is encoded in order to locate the sample area along the dive track. The first number (1 or 2) refers to an ALVIN dive or an ANGUS camera track, respectively. The second number (after the first decimal point) is the dive number. The third number (after the second decimal point) is a range, indicating the times along the dive tracks (to the best estimation) corresponding to the sample areas.
Nadir Long and Nadir Lat refer to the longitude and latitude of the sample areas, located along the nadir of the SeaMARC II swaths. These values are used to calculate actual longitude and latitude.
Distance From Nadir is the distance measured, perpendicular to the SeaMARC II ship track, from the nadir to the center of the sample area. (This was measured directly from the SeaMARC II images using a Gerber Scale.) This value is also used in the calculation of the actual longitude and latitude of the sample areas.
Ship Heading indicates the azimuth of the ship's course during the SeaMARC II survey. This value is used to calculate the actual longitude and latitude.
Longitude is calculated from the expression:

\[ |((D \times \sin(C+90))/180300)| + A \] if the sample area is on the starboard side of the ship track, and

\[ A - |((D \times \sin(C-90))/180300)| \] if on the port side.

Where A is nadir long, D is distance from nadir, C is ship heading and 180300 is a conversion factor for longitude (m/deg).

Latitude is calculated from the expression:

\[ |((D \times \cos(C+90))/111120)| + B \] if on the starboard side, and

\[ B - |((D \times \cos(C-90))/111120)| \] if on the port side.

Where the notation is the same as above and B is nadir lat, and 111120 is a conversion factor for latitude (m/deg).

Volcano is encoded in order to indicate precisely where on Seamount 6 each sample area is located. The encoding is explained in the Key to Volcano Code section of Appendix B.

Geology is an encoded description of lithologies present in the sample areas. The encoding is explained in the Key to Geology Descriptors section of Appendix B.

Box Size shows the size, in pixels, of each sample area. Efforts were taken to keep box size as equal as possible.

Mean I is the mean intensity statistic computed from the grey level difference matrices generated by the textural analysis computer program, ss_get.
MAD is the mean absolute deviation from the same matrix.
STD is the standard deviation.
VAR is the variance.
SKEW is the skewness.
KURT is the kurtosis.
Slope Angle is the best estimation of the range of bathymetric slope within each sample area, as taken from the slope map in Figure 58.
Look Angle is the angle, perpendicular to the ship track, from nadir out to the center of the sample area. It was calculated according to the expression: \( \tan^{-1}\text{(distance from nadir/water depth)} \).
Rough attempts to quantify the roughness of the sample area according to a scale which is explained in the Key to Roughness Scale section of Appendix B.
Water Depth for each sample area was determined from the SeaBeam bathymetric map.
Contrast, IDM (Inverse Difference Moment), Entropy, Energy, Mean, Corr (Correlation), and Isotropy are the seven textural variables generated by the ss_get program using the grey level differences. These are explained in Appendix A.
Same as Box indicates which sample areas have a corresponding pair which was viewed at a different look direction.
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