Volcaniclastic aeolian dunes: Terrestrial examples and application to martian sands

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Abstract. On Earth, most aeolian dunes are quartz rich; others are composed of evaporites, carbonates, or clay/silt aggregates. Dunes made of wind-reworked volcaniclastic sediment comprise a less-commonly recognized fifth dune composition. Terrestrial volcaniclastic aeolian dunes are found in (1) arid to semi-arid volcanic regions and (2) coastal areas on volcanic islands. Their sediments can be formed by explosive volcanism or by erosion of lava flows and other lithified volcanic material. Commonly, these sediments have been transported by volcanic and/or fluvial processes before being reworked by wind. Their compositions range from mafic to sialic, depending on local volcanic sources. Volcaniclastic dunes, especially those of basaltic composition, may be the best compositional analog for aeolian dunes on Mars. Martian dunes are typically dark-hued and their sands may be derived from erosion of volcanic materials.

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Introduction

The prevalence of a particular mineral or rock type as a sand-forming material is generally dependent upon the abundance of the material in source rocks, the rate at which the material is formed, and the rate at which the material is destroyed by mechanical and chemical weathering processes. On the Earth, quartz comprises most of the silt, sand, and smaller pebbles, because it is abundant in continental crust and it is very resistant to mechanical and chemical weathering (e.g., Kuenen 1960, Krinsley and Smalley 1972). Most of the windblown, dune-forming sands on the Earth consist mainly of quartz (Bagnold 1941, McKee 1979, Pye and Tsoar 1990). Other dunes are composed of evaporite minerals such as gypsum (Jones 1953, McKee 1966), carbonates (McKee and Ward 1983), or clay aggregates (Coffey 1909, Huffman and Price 1949, Bowler 1973). Although there have been many published reviews and bibliographies concerning the nature of sand dunes found on the Earth and other planets (Warren 1969, Petrov 1976, McKee 1979, Smith 1982, Busche et al. 1984, Niessen et al. 1984, Lancaster 1988), very few papers have been noted, and only Pye and Tsoar (1990) have indicated that there is a fifth dune composition: those composed largely of volcaniclastic material. Dunes composed of aeolian reworked volcanic sands are of interest for three main reasons: (1) they may be compositional analogs to sands on the planet Mars, (2) they may aid in establishing criteria by which to distinguish cross-bedded sands formed by primary igneous processes (such as base surge) from secondary reworking by the action of wind (Smith and Katzman 1991), and (3) some volcaniclastic aeolian sands consists of grains of contrasting densities, allowing the study of grain sorting by composition in a natural environment. This paper reviews (a) the case for the occurrence of non-quartz, aeolian sands on Mars, and (b) the terrestrial occurrences of volcaniclastic dunes. The origins and compositions of the volcaniclastic sands and how they may relate to aeolian sands on Mars are also explored.

Martian Aeolian Sands

The Presence of Windblown Sand

There is considerable evidence for the widespread occurrence of sand on Mars (Peterfreund et al. 1981a, b). Evidence for sand comes from the images and the interpretation of thermal infrared data taken by orbiting spacecraft, and the observations obtained by the Viking landers.

Dunes are seen in a wide variety of locations on Mars (Figure 1), especially in the north polar region (Cutts et al. 1976, Tsoar et al. 1979), on the floors of some craters (Cutts and Smith 1973, Breed 1977, Thomas 1981, 1982, 1984), and in a variety of locations seen at very high resolution (Peterfreund 1981, Zimbelman 1987). Dune fields large enough to have been seen by spacecraft cover at least 0.5% of the martian surface (Peterfreund et al. 1981a). The morphology of martian dunes is similar enough to terrestrial dunes that they

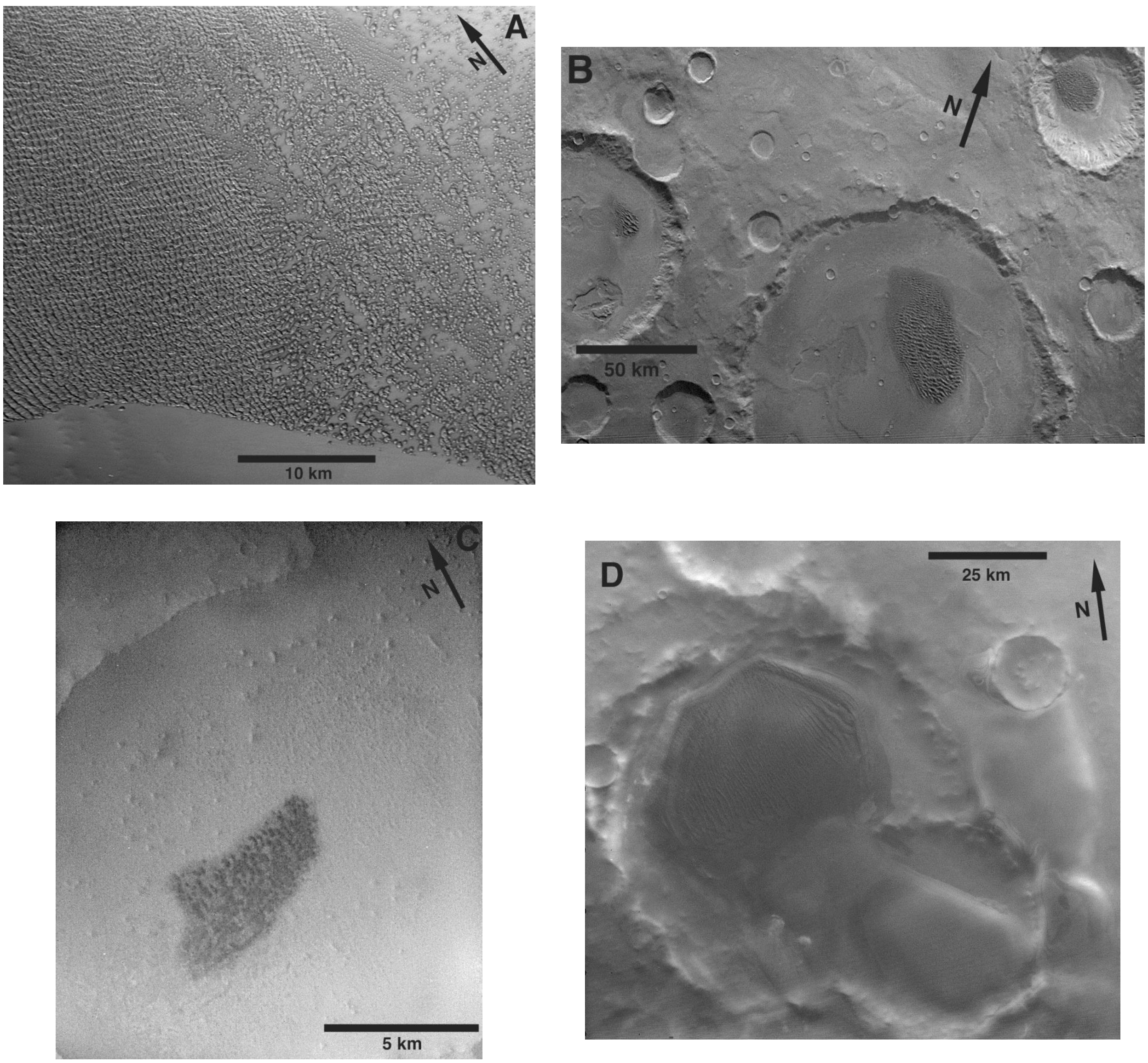

Figure 1. Examples of dune fields on Mars. **(a)** North circum-polar dunes (Viking image 524B23, center at 77.6°N, 50.9°W). **(b)** Three crescentic dune fields in the Hellespontus region (center left, lower right—Proctor Crater, upper right; Viking image 094A47, center at 47.2°S, 329.4°W). **(c)** Barchanoid dunes on the floor of a crater in the Oxia Palus region (Viking image 709A42, center at 1.9°N, 351.7°W). **(d)** Dune field in Richardson Crater, in the high southern latitude region (Viking image 516B64, center at 72.4°S, 180.2°W).

are considered to be composed mostly of sand-sized material (Cutts and Smith 1979, Breed et al. 1979).

Mid-infrared radiometric measurements obtained by the Viking Infrared Thermal Mapper (IRTM) have been used to determine thermal inertia, which under martian atmospheric pressure conditions is closely related to differences in the effective particle size of unconsolidated, granular materials (Neugebauer et al. 1971, Kieffer et al. 1973, 1977). Dunes in the Hellespontus region of Mars (40°S to 55°S, 315°W to 350°W) have thermal inertias that are consistent with particle sizes of 500 ± 100 µm (medium to coarse sand) (Edgett and Christensen 1991a). Dunes in the martian north polar region and dunes in central Syrtis Major also have thermal inertia values indicative of sand (Peterfreund 1981, Keegan et al. 1991).

Evidence for the presence of sand at the Viking lander sites has been a more controversial subject. Unfortunately, the smallest particle size that could be directly measured by passing material through screens within the lander was about 2 mm in diameter, thus the size distribution of particles smaller than 2 mm remains largely unknown (Shorthill et al. 1976). An analysis by Shorthill et al. (1976) inferred that there might have been as much as 60% materials smaller than fine sand (10–100 µm) and as much as 30% materials in the sand-size range (100–2000 µm) at the Viking 1 site. Analysis of the Viking lander images, however, led Mutch et al. (1976) to conclude that there was little, if any, sand-sized materials at the lander sites. Mutch et al. (1976) and Moore et al. (1977) favored the predominance of finer material (<< 100 µm) at the Viking lander sites, on the basis of (1) the occurrence of cohesive drifts, (2) the cohesion of material in the walls of trenches dug by the lander remote arm, and (3) lander camera observations of dust suspended in the atmosphere. Two explanations for the apparent lack of sand at the lander sites were offered: (a) sand had been transported out of the region, or (b) sand-sized particles may be rapidly abraded and broken down to smaller sizes in the martian wind regime (Mutch et al. 1976). High particle velocities and lack of cushioning by the thin (average 6.5 mbar) martian atmosphere, in combination with the possibility that the sand-forming minerals might be relatively less resistant varieties like feldspars and pyroxenes, are considered to be factor which would make it easier to reduce sand to smaller particle sizes through round, chipping, and splitting (Mutch et al. 1976, Sagan et al. 1977, Krinsley and Leach 1979, Krinsley et al. 1979, Krinsley and Greeley 1986).

Sharp and Malin (1984) re-examined the evidence for sand at the Viking lander sites. They noted the presence of possible granule ripples, which on Earth are known to be accumulations of 2–4 mm-sized particles moved by the impact of saltating sand (e.g., Sharp 1963). Sharp and Malin (1984) also noted that the drifts near Viking 1 have cross-bedded textures, which suggested to them that sand makes up the bulk of these features. Finally, aeolian deposits in the lee of cobbles at the lander sites were considered to consist of sand, because finer material should have gone into suspension and been carried farther away from the cobbles. The proposed sand materials were seen to be darker in color than surrounding fine materials (Sharp and Malin 1984), an observations that is consistent with the occurrence of dark materials also postulated to be sand comprising the north polar dunes, intracrater dunes, and possibly much of Syrtis Major.

Composition of Martian Sand

The composition of martian aeolian sand grains is presently unknown. Quartz is not likely to be common on Mars because there is little evidence that considerable amounts of martian magma became differentiated enough to form the silica-rich rocks from which quartz grains are derived (McGetchin and Smith 1978, Francis and Wood 1982). Regolith samples analyzed by the Viking landers had low K_2O and Al_2O_3 abundances, which indicates that the martian soils were derived from weathered mafic materials (Toulmin et al. 1977). Even 10% granitic material would have yielded significantly larger values for K_2O than were observed (Toulmin et al. 1977). Spectral reflectance data also suggest that mafic and ultramafic materials are prevalent on Mars, especially on the basis of the presence of the Fe^{2+} absorption band near 1 µm, which probably indicates unaltered pyroxenes and olivines (Singer et al. 1979). Broadband thermal infrared emissivities of dark, sandy deposits on martian crater floors show deeper silicate reststrahlen absorption features in the 11 µm than the 7 or 9 µm Viking IRTM channels, which also suggests a dominance of mafic or ultramafic material in martian sands (Edgett and Christensen 1991b, 1992). Most of the areas on Mars proposed to consist of windblown sand, such as the north polar dunes and various intracrater deposits, have low albedos which indicate the presence of dark-hued grains, such as basalt fragments. Indeed, most of the sand on Mars might consist of mafic and ultramafic lithic and mineral fragments (Smalley and Krinsley 1979, Baird and Clarke 1981). Many of these sands could be volcaniclastic in origin, as there are many volcanic regions on the planet (e.g., Greeley and Spudis 1981); Peterfreund (1985) also suggested that gabbroic intrusive rocks might produce large, durable, mafic sand grains.

Other sand compositions that could occur on Mars are: silt/clay aggregates, iron oxides such as limonite, and perhaps evaporite or carbonate materials. It has been proposed that some martian dune materials might be aggregates of silts and clay (Ksanfomaliti 1977, Greeley 1979, Krinsley and Leach 1981). Aeolian dunes composed of silt/clay aggregates are known to occur on the Earth; such grain aggregates are typically cemented by salts (Bowler 1973). Greeley (1979) suggested that similar aggregates could form on Mars, either by salt cementation or electrostatic cohesion (see Krinsley and Leach 1981). The possible martian aeolian properties of laboratory-produced clay aggregates (Saunders et al. 1986) were investigated by Saunders and Blewett (1987). Thomas and Veverka (1986) indicated that limonite sand would explain red–violet contrast reversals seen in Viking color images of dark aeolian deposits on the martian surface. However, while limonite has long been considered to be a possible component of martian surface materials (because of the prevalence of the oxidized iron (Fe^{3+}) which gives Mars its red color), its proposed occurrence has largely been a matter of controversy (e.g., van Tassel and Salisbury 1964, Adams 1968). Herkenhoff (1992) proposed that some Mars sands might contain or consist of heavy, fine-grained, oxide minerals such as magnetite or maghemite. Although no one has proposed martian aeolian sands composed of evaporites or carbonates, some investigators have suggested that carbonates might occur on Mars (see review in Blaney and McCord 1989); and Pollack et al. (1990) presented observations suggesting that windblown dust on Mars contains some amount of sulfates, carbonates, and hydrates.

Sources of Martian Sand

Sand on Mars could be produced by impact cratering, which would form lithic fragments and glass beads; by past fluvial action, for example, the large martian outflow channels (Baker 1982); by small-scale mechanical weathering due to thermal effects, frosts, or salts (Malin 1974); by pyroclastic eruption; and possibly by the aggregation of clay and silt particles. The original source for all of these particles would either have been martian highlands crustal material, any of the numerous suspected volcanic deposits (Greeley and Spudis 1981), or perhaps chemical sedimentary or evaporite deposits. Primary sand grains might be (a) feldspar, olivine, or pyroxene grains derived from ancient crust or as phenocrysts from lavas, (b) glassy particles from pyroclastic volcanism or impact explosions, (c) other fine-grained pyroclastic particles such as basaltic cinders, (d) epiclastic lithic fragments derived from the erosion of lava or welded pyroclastic deposits, or (e) possibly clay aggregates, evaporites, or carbonate materials.

Possible Examples of Martian Volcaniclastic Sand and Sources

Because martian sands are generally dark (low albedo) and may have spectral absorption features near 1 and 11 μm attributed to mafic mineralogies, most martian sands may be mafic or ultramafic in composition. Some of the sand might be derived from the fluvial erosion of lavas, as there are many examples of ancient fluvial systems (Baker 1982) and mafic lava flows (e.g., Greeley and Spudis 1981) on Mars. Some of the mafic sands might be pyroclastic in origin, although evidence for pyroclastic deposits on Mars is generally sparse. Some possible cinder cones have been identified (Wood 1979, Frey and Jarosewich 1982, Zimbelman and Edgett 1992), some volcanoes are thought to have significant pyroclastic deposits (Greeley and Spudis 1981, Mouginis-Mark et al. 1988, Greeley and Crown 1990), and some large, regional-scale ash deposits have been proposed to occur on Mars (Scott and Tanaka 1982, Moore 1990). One example of possible mafic volcaniclastic aeolian sand is a dark dune field on the floor of Valles Marineris; Geissler et al. (1990) examined visual reflectance data and suggest that these sands could be mafic glass eroded from dark deposits in Valles Marineris that have been interpreted as pyroclastic materials (Lucchitta 1987, 1990). Another potential site for volcaniclastic aeolian sands is the low albedo, sandy Syrtis Major region (Simpson et al. 1982), for which Soviet–French Phobos 2 near-infrared observations (ISM instrument) suggest the presence of augite-basaltic materials (Mustard et al. 1993).

Terrestrial Volcaniclastic Aeolian Dunes

The map in Figure 2 shows the location of volcaniclastic aeolian dunes discussed in this text. The dunes described here are composed of ≥ 50% volcaniclastic materials. There are many other locations where quartz or carbonate sand dunes have incorporated smaller amounts of locally-derived volcanic sands. Examples include quartz-rich dunes along the Snake River near St. Anthony and Bruneau, Idaho, which contain basalt fragments (e.g., 6–8% at Bruneau) (Koscielniak 1979, Murphy 1973, 1975); sand dunes of the Mexican Gran Desierto, directly adjacent to the Pinacate volcanic field (Arvidson and Mutch 1974); and carbonate dunes containing volcanic rock and mineral fragments found on islands in the Azores (Berthois 1953). The following sections provide general descriptions of the terrestrial volcaniclastic dune fields and their compositions. Such dunes occur in Iceland, western U.S.A., Hawaii, Peru, New Zealand, and several other volcanic regions. Table 1 presents a summary of volcaniclastic dune compositions for locations where such data are available.

Iceland

Iceland is a largely unvegetated, arid, volcanic island, and therefore seems a likely place to find volcaniclastic dunes. Nearly all of the Icelandic soils are composed of volcanic-aeolian sediments, most of which is loess (Preusser 1976). Sand dunes, however, are relatively rare, probably because the fluvial systems and strong winds effectively remove considerable amounts of loose sand from the island (Preusser 1976). Where sand dunes do occur, they are commonly dark or black in color due to mafic materials. They are typically small and low (~1–4 m high) (Iwan, 1937, Cailleux 1939, 1973, Sveinbjarnardóttir et al. 1982). Mafic aeolian sands and dunes have been noted in several locations, though none have been described in detail: (1) near the Skeidarájökull (glacier) (Bogacki 1970), (2) south of Lake Myvatn (Clarke 1970), (3) west of Vadalda (Malin and Eppler 1981), and (4) in the Merkurhraun near Gunnarshold, both in southwestern Iceland (Cailleux 1939). Finally, Ingólfsson (1982) reported that several small sand dunes composed of reworked tephra had formed on the young volcanic island of Surtsey by 1979. The Surtsey dunes were small and composed of coarse and very coarse sands derived from the adjacent tephra cone.

Central Washington, U.S.A.

Most of the dunes in the semi-arid central and southeastern parts of Washington consist mainly of quartz with minor basalt (Flint 1938, Lewis 1960,

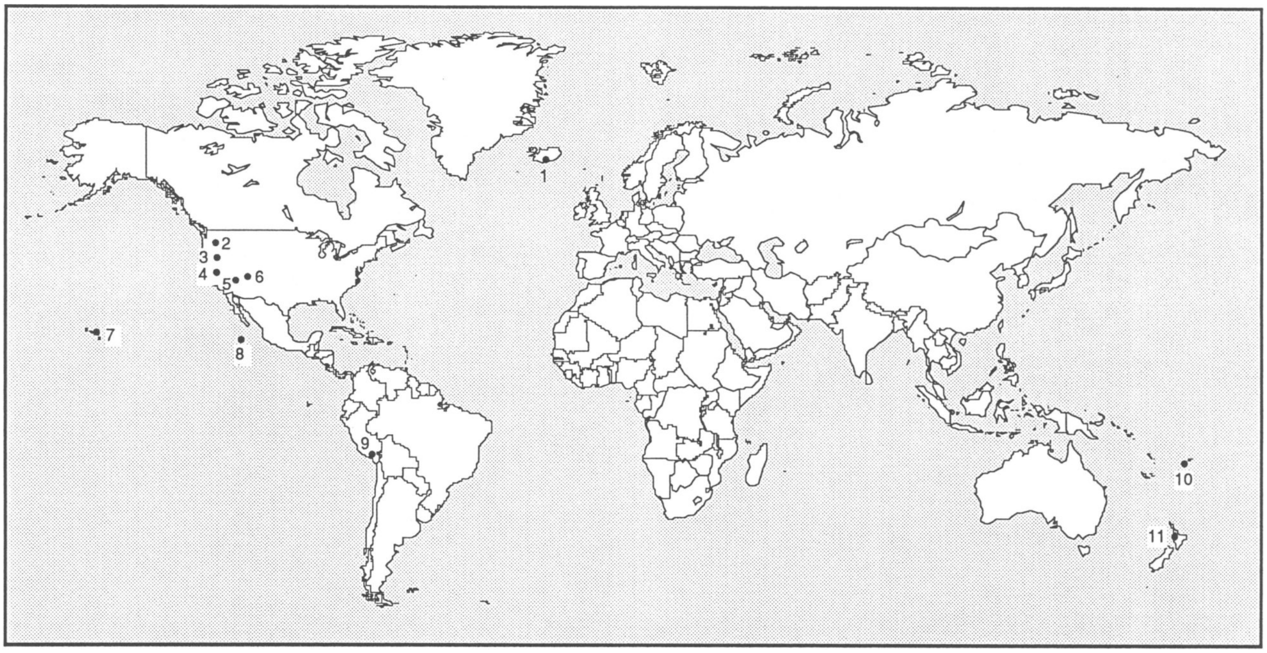

Figure 2. Locations of terrestrial volcaniclastic aeolian dunes. **(1)** Iceland. **(2)** Moses Lake, Washington. **(3)** Southeastern Oregon. **(4)** Mono Lake, California. **(5)** Central Arizona. **(6)** Great Sand Dunes National Monument, Colorado. **(7)** Hawaii. **(8)** Isla San Benedicto, Mexico. **(9)** Southern Peru. **(10)** Viti Levu, Fiji. **(11)** North Island, New Zealand.

Easterbrook and Rahm 1970, Grolier and Bingham 1971, 1978); however the largest dune field in the state is composed of 60% to 80% basalt fragments (Table 1; Petrone 1970). The basalt dunes are located in the Quincy Basin of central Washington, south of Moses Lake at the present site of the Potholes Reservoir (Figure 3). Basaltic sands are thought to have been deposited in the Quincy Basin during the floods which formed the channeled scablands in central Washington (e.g., Baker 1978), and these sediments were subsequently reworked by wind to form eastward-migrating, transverse, parabolic, and barchan dunes (Landes et al. 1912, Lewis 1960, Petrone 1970, Easterbrook and Rahm 1970, Grolier and Bingham 1971, 1978, Gulick 1990). Many of the largest dunes (10–20 m high) are presently active although the amount of vegetation cover has increased since the filling of the Potholes Reservoir in the early 1950s. In 1980, the dunes were covered with 2–4 cm of ash from the explosive eruption of Mount St. Helens, and since that time, dunes which remained active have lost this covering of ash, while inactive dunes have remained buried. The presently active dunes, mostly of parabolic form, have a layer of ash stratigraphically preserved within the dune at the location of the 1980 slip-face (Edgett et al. 1991).

Southeastern Oregon, U.S.A.

Figure 4 shows the distribution of Quaternary aeolian sands in the southeastern quarter of Oregon. The region is part of the North American Great Basin, and each basin in Figure 4 is the site of a former Pleistocene lake (Greenup 1941, Allison 1966). After the Pleistocene lakes dried up, some of the lake sediments in each basin were blown eastward to form duns (1–30 m high) (Cope 1889, Smith and Young 1926, Dole 1942, Allison 1941, 1966, 1979, 1982, Walker 1977, Mehringer and Wigand 1986). Their compositions are described as "ash, pumice, and rock-forming minerals, mostly alkalai feldspar and quarts" (Walker 1977). Weide (1974) noted that some of the dunes in Warner Valley are composed of silt/clay aggregates, rather than volcaniclastic sand grains. McDowell (1984) has also identified clay in the dune sands, and basalt fragments are present in some dunes (Edgett et al. 1991). The mineralogy of only one sample taken from the area known locally as the Shifting Sand Dunes (located in eastern Christmas Lake Valley) has been reported in the literature (Dole 1942). The dune sampled by Dole (see Table 1) contained nearly 50% feldspar and 50% pumice, with minor amounts of fossil material and other sediments from the dry Pleistocene Fort Rock Lake bed. Allison (1941, 1966, 1979, 1982) indicated the presence of elongate, vegetated dunes in Christmas Lake Valley and transverse dunes near Summer Lake that are composed almost entirely of pumice derived from the Crater Lake-forming explosion of Mt. Mazama about 6600 years ago.

Mono Lake, California, U.S.A.

Bailey (1989) mapped aeolian deposits near Mono Lake, California, and describes them as "dune and windblown deposits composed predominantly of sand, ash, and fine pumice clasts formed mainly by aeolian redeposition of ash and pumice lapilli." There is no other published information concerning these volcaniclastic sand deposits; recent field examination

Table 1. Generalized Average Compositions of Volcaniclastic Dunes[1]

Location	Quartz %	Feldspars		Basalt frag-ments %	Glass, Pumice %	Other rock frag-ments %	Amphiboles, pyroxenes, olivines, micas %	Opaque (metallic) %	Reference
		Plagio-clase %	K-spar %						
Moses Lake, Washington[2]	20–40	—	—	60–80	—	—	—	—	Petrone 1970
Shifting Sand Dunes, Fossil Lake, Oregon[3]	—	—	40	—	54	trace	3[4]	3[5]	Dole 1942[3]
Great Sand Dunes, Colorado	29	5.8	3.1	—	—	57.9[6]	3	1.2	Wiegand 1977
Ka'u Desert, Hawaii[7]	—	~5	—	~25	~50	—	~15[8]	~5	Gooding 1982
Arequipa region, Peru[9]	15	35	—	—	5	—	35	10[5]	Hastenrath 1967
Singatoka, Viti Levu, Fiji[10]	20	12	2	—	—	25	32	9[11]	Dickinson 1968
South Kaipara Barrier, North Island, New Zealand[12]	18–43	12–45	1–5	8–40[13]	—	—	—[13]	0–5[11]	Schofield 1970

[1]The figures presented here are the best available average % by volume.

[2]Values for the "leading tongue" at the eastern end of the dune field (see Figure 3).

[3]Sampled for only one dune (see Figure 4).

[4]Pyroxenes and trace biotite.

[5]Mostly magnetite.

[6]Mostly fragments of the San Juan volcanics, including some basalt.

[7]Generalized from Gooding (1982, Fig. 5); glass is basaltic.

[8]Mostly olivine.

[9]See Table 3 in Hastenrath (1967) and Table 4 in Finkel (1959).

[10]Averaged from Table 4 in Dickinson (1968).

[11]Mostly titanomagnetite.

[12]From Table 3 in Schofield (1970).

[13]Mafic minerals and basalt fragments grouped together.

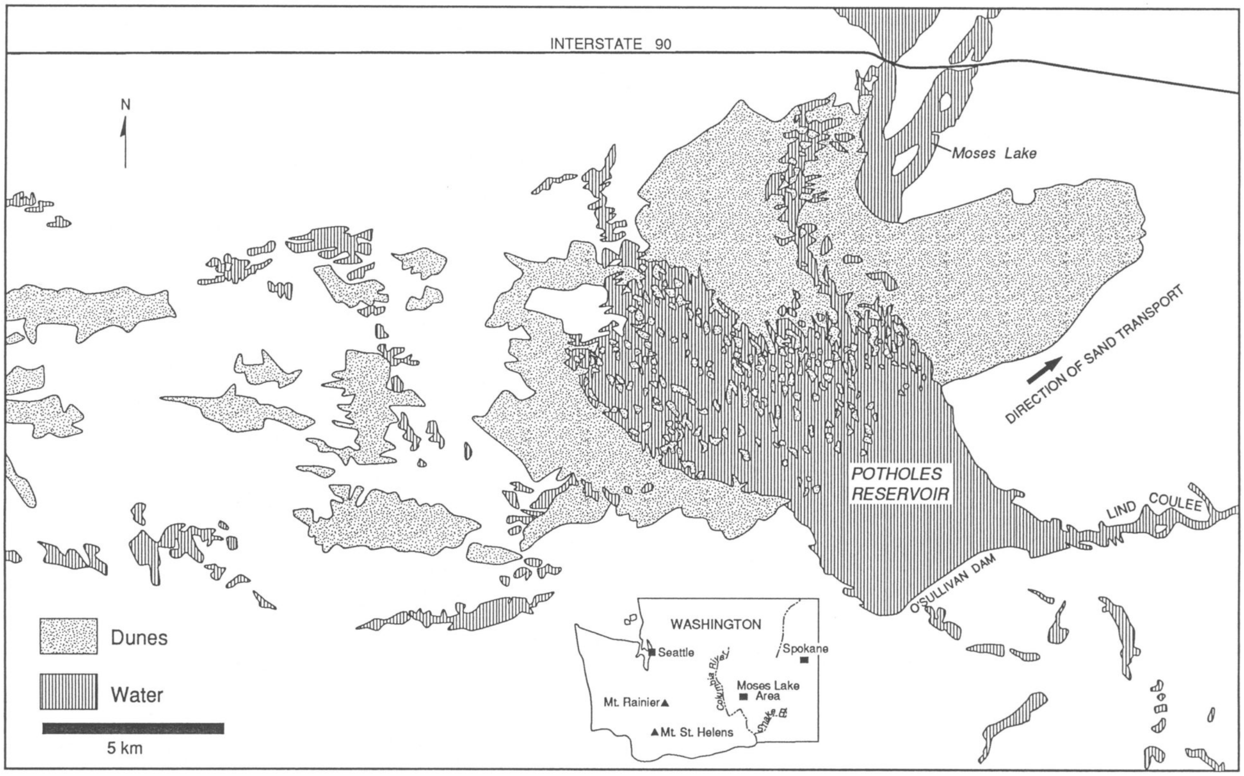

Figure 3. Sketch map of the Moses Lake dunes, central Washington, U.S.A. (based upon U. S. Geological Survey maps 1979, 1982, 1986a, 1986b).

has shown them to be largely vegetated. Most occur east and southeast of Mono Lake and consist mainly of obsidian and pumice (Edgett et al. 1991).

Central Arizona, U.S.A.

Black dunes composed largely of basaltic ash from the cinder cone field surrounding the San Francisco Peaks, near Flagstaff, Arizona, have been only briefly described in the literature (Colton 1937, Cooley et al. 1969, Moore et al. 1974, Moore and Wolfe 1976, Breed 1976, Breed et al. 1984). The sands are apparently derived from the aeolian redeposition of basaltic ash; the sands are blown eastward, and their particle size is thought to decrease with distance from the source. The dunes in this region are usually low, mound-like forms, or they are falling dunes occurring along eastward-facing cliffs, especially in the vicinities of the Wupatki National Monument and the Grand Falls of the Little Colorado River.

South-central Colorado, U.S.A.

The Great Sand Dunes National Monument is found in the San Louis Valley, adjacent to the Sangre de Cristo Mountains, northeast of Alamosa, Colorado. The dune field covers 104 km^2 and the dunes (mostly transverse) are up to 213 m high. These dunes have been described by various authors, especially Burford (1961), Johnson (1967), Wiegand (1977), and Andrews (1981). Wiegand (1977) found that the dune sands are composed of 51.7% volcanic rock fragments (Table 1) carried down from the San Juan volcanic field (to the west) by the Rio Grande.

Hawaii, U.S.A.

Numerous Hawaiian beaches are composed of basaltic sands, other beach sands are composed of calcareous fragments of organic origin (Moberly et al. 1965). Many dune sands on the Hawaiian Islands are also composed of carbonate materials (Wentworth 1925, Stearns 1940, 1946, 1947, 1970, MacDonald et al. 1983). However, there are also Recent and Pleistocene cross-bedded aeolian sands composed of dark, basaltic grains (Stearns 1940, 1946, 1947, Gooding 1982). Most of the dunes are low, poorly defined mounds (Stearns 1946, Gooding 1982). Most of the dunes are low, poorly defined mounds (Stearns 1946, Gooding 1982), while others may be barchanoid dunes up to 15 m high (Stearns 1946). Grain size and sorting characteristics were used by Gooding (1982) to establish that dunes in the Ka'u Desert (near Kilauea volcano, Island of Hawaii) are aeolian in origin. The Ka'u Desert dunes contain basaltic glass, olivine, lithic basalt fragments, magnetite, ilmenite, and plagioclase (Table 1; Gooding 1982), and are derived from the reworking of pyroclastic material.

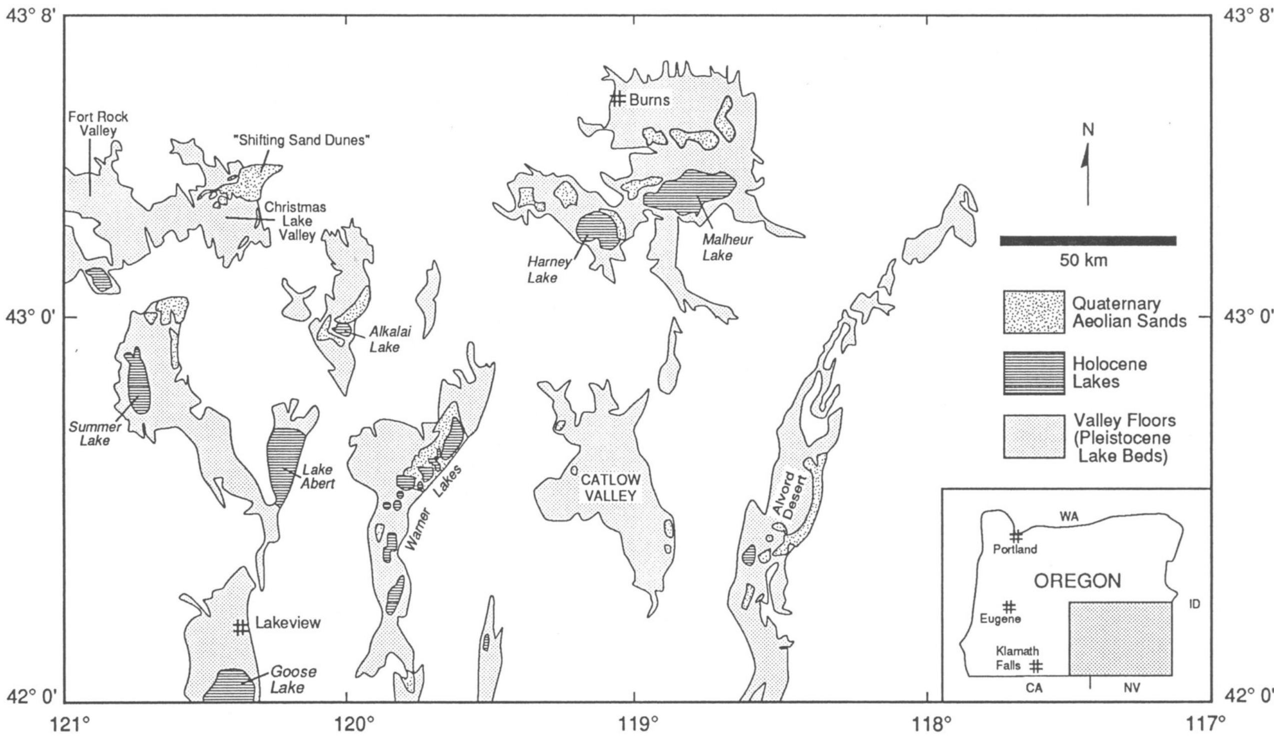

Figure 4. Sketch map of southeastern Oregon, showing the locations of basins (Pleistocene lake beds) and Quaternary aeolian sand deposits, including presently active dunes (after Walker 1977). The largest active dune field, known as the Shifting Sand Dunes, is located in eastern Christmas Lake Valley.

Isla San Benedicto, Mexico

Volcán Bárcena, on the island of San Benedicto (about 200 km due south of Mazatlán, Mexico) erupted in 1952. The eruption covered much of the island with trachytic tephra, most of which (80%) was smaller than 250 μm in diameter. Richards (1965), using aerial photographs, noted that by 1955 wind action had created barchan dunes composed of the 1952 tephra.

Southern Peru

A field of classic barchan dune forms is located south-southwest of Arequipa, Peru, in the Pampa de la Joya and Pampa de la Clemesi (Bailey 1906, Douglass 1909, Finkel 1959, Hastenrath 1967, Lettau and Lettau 1969). These barchans are generally 1–6 m high and dark gray in color. In order of abundance, their major mineralogic components are: plagioclase, hypersthene and horneblende, biotite, and volcanic glass (Table 1; Finkel 1959). These sands are thought be eroded from nearby volcanoes, the glass is probably pyroclastic in origin. Other dark-hued, volcaniclastic dunes and aeolian sands are found on and to the east of El Misti volcano located east of Arequipa (Bailey 1899, de Silva and Francis 1990).

Viti Levu, Fiji

Coastal dunes occur at the mouth of the Singatoka River on the Island of Viti Levu, Fiji (Dickinson 1968). These dunes are volcaniclastic in origin, some derived from primary volcanics, others from the erosion of Tertiary volcaniclastic rocks. The sand contains (in order of abundance) mafic and opaque minerals, volcanic rock fragments, plagioclase, and quartz (Table 1). Much of the sand was brought down to the coast by the Singatoka River.

North Island, New Zealand

Sand dunes, both active and older, vegetated forms, are common along the west coast of the North Island of New Zealand (e.g., von Hochstetter 1964, Cockayne 1909, 1911). The dunes along the west coast of the North Island range in composition from quartzose to mafic, and contain varying amounts of titanium- and iron-bearing minerals (Cockayne 1909, 1911, Hutton 1940, Fleming 1946, Nicholson and Fyfe 1958, Williams 1965). Much of the volcaniclastic material in these dunes is originally derived from the Taupo volcanic zone in central North Island and the Egmont volcanic region in southwest central North Island (Figure 5). For the purpose of discussion here, the North Island has been divided in to four regions (Figure 5), one of which occurs near the Taupo volcanic area itself (Area 4). From north to south, the three coastal dune areas are: (a) from Cape Reinga to the mouth of the Waikato River, (b) from the Waikato River to Cape Egmont, and (c) from Cape Egmont to Paekakariki.

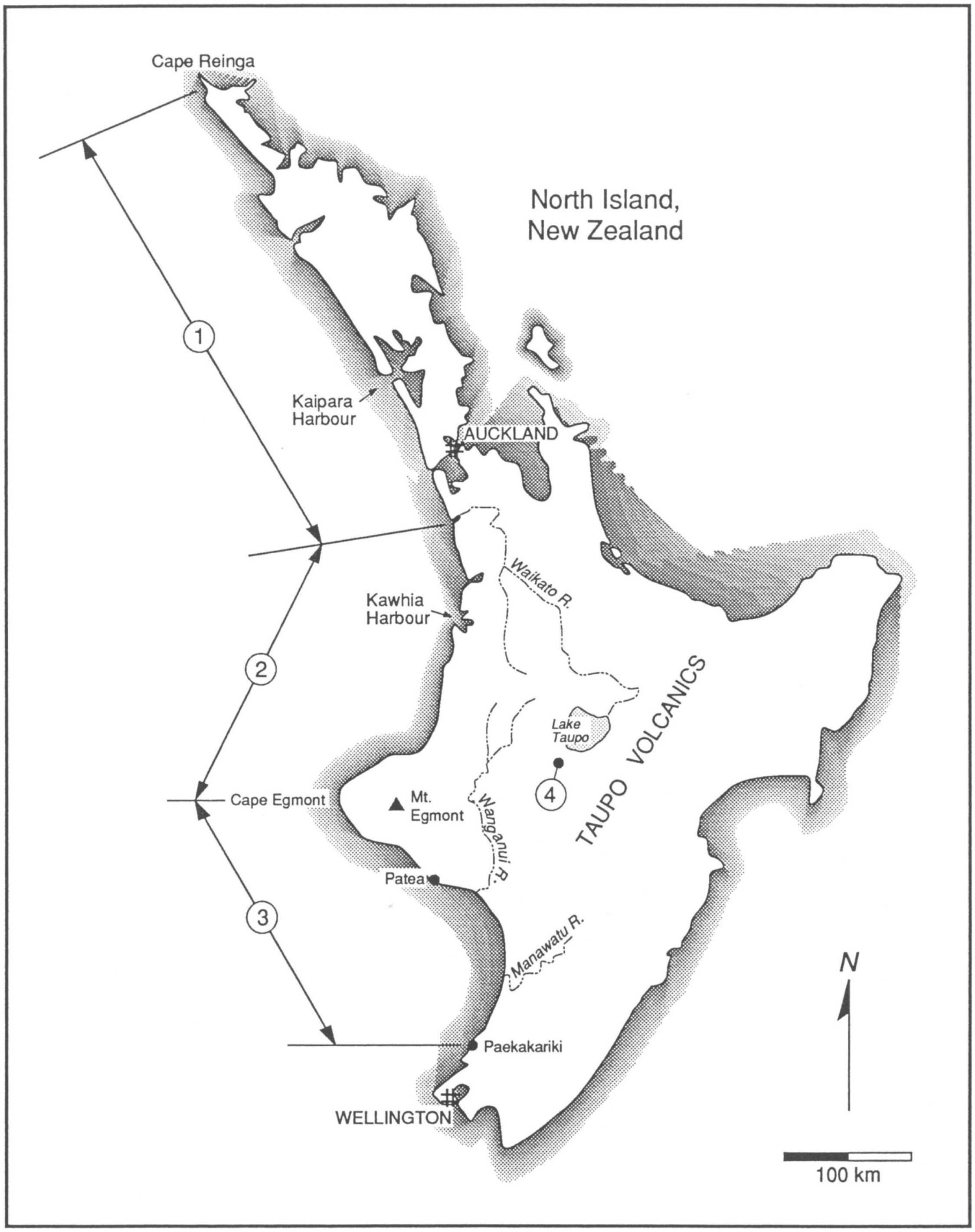

Figure 5. Map of North Island, New Zealand. The main volcaniclastic dune areas occur along the western coast in Regions 1, 2, and 3, and in the interior of the island at area 4. The locations labeled 1, 2, 3, and 4 are described in the text.

Area 1

Quaternary dune sands in this region vary in composition, size, and degree of activity. The dunes average 10 m in height, but commonly range up to 30 m, and some are as high as 275 m (Kear 1964). The primary dune-forming minerals along this coast are quartz and feldspar (both plagioclase and orthoclase), but there are several areas where non-quartz minerals dominate the dune deposits (Schofield 1970), including some dune areas with up to 80% (by weight) titanomagnetite (Nicholson and Fyfe 1958). The main areas of significant volcaniclastic dunes are (1) North Kaipara barrier, where one dune sampled by Schofield (1970) had 32% mafic materials, 24% plagioclase, and ~4% lithic fragments (on North Kaipara barrier, the mineralogy actually varies: 2 to 32% mafic minerals, 38 to 60% quartz, 26 to 52% feldspar); (2) South Kaipara barrier, where mafic mineral content varies from < 10% in the north to 80% in the south (Figure 6; Table 1; Schofield 1970); (3) at the mouth of Manukau River, where Schofield (1970) found 57% mafics and 29% plagioclase.

Area 2

Much of the beach and dune sand in this region (Figure 5) consists of dark "ironsands" which have been explored and mined for iron and titanium. Chappell (1970) and Pain (1976) examined the dunes near the Aotea and Kawhia harbors and noted that there are two Quaternary dune members: one, and older unit (deposited ~2000 years B.P.) and the other a recent, active unit (began ~800 years B.P.). Chappell (1970), Pain (1976), and Nicholson and Fyfe (1958) note abundant titanomagnetite resources in these sands, ranging from < 1 to 87% by weight ironsand. Cockayne (1909, 1911) noted that the dunes in this region are commonly black in color and contain varying proportions of hornblende and augite as well as magnetite. Most of the mafic materials are thought to have been derived from Mt. Egmont and the other Taranaki volcanoes to the south (e.g., Stokes and Nelson 1991).

Area 3

Dunes of the third region (Figure 5) occur in scattered locations from Cape Egmont to Patea, while dunes from Patea south to Paekakariki form a large unit known as the Manawatu Sand Country. The dunes between Cape Egmont and Patea contain mostly feldspars, hornblende, augite, and magnetite, and are mainly derived from the Egmont volcanoes (Cockayne 1909, 1911, Morgan and Gibson 1927). The Manawatu Sand Country has been studied by numerous people, usually in the context of soil stratigraphy, usage, and sand control (Te Punga 1953, Cowie 1957, 1963, 1968, Cowie and Smith 1958, Cowie et al. 1967, Saunders 1968, Holland 1983, Shepherd 1985, 1987). Cowie (1963) recognized four phases of dune-building in the Manawatu region, the oldest occurred 20,000 to 10,000 years B.P., the most recent began ~100 years ago. Fleming (1953) described the various dune morphologies of the region—the dunes are mostly transverse, parabolic, and barchans. The composition of the dunes generally varies from more mafic sands in the north to quartzo-feldspathic sands in the south (Te Punga 1953, Fleming 1953, Cowie and Smith 1958, Cowie et al. 1967). Dune sands in the north, near Patea, contain mostly magnetite, pyroxenes, plagioclase, and hornblende, with minor quartz; much of these sands are derived from the Mt. Egmont volcanic region (Hutton 1940, Fleming 1946, 1953). Dunes further south, especially between Wanganui and Paekakariki (Travers 1881), contain mostly quartz, feldspar, and carbonates, but some older dunes have 20 to 40% pumice and glass fragments thought to have come mainly from the Taupo eruption of ~200 A.D. (Fleming 1953, Cowie 1963).

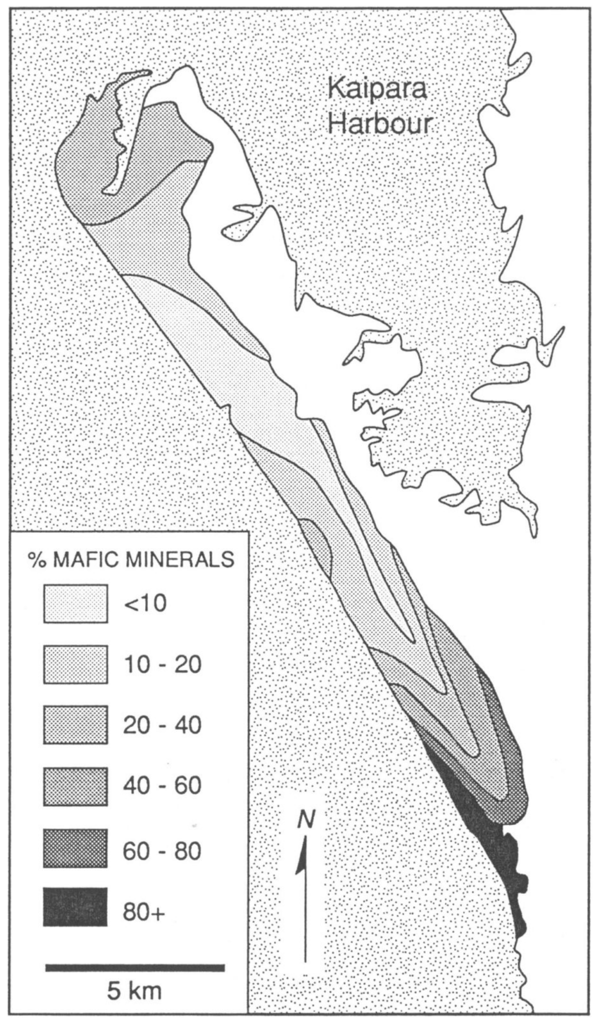

Figure 6. Contour map of mafic mineral content in the dune and beach sands on the south Kaipara Barrier, New Zealand (modified after Schofield 1970).

Area 4

Cockayne (1909, 1911) described three inland dune areas in New Zealand, two of which are quartz dune fields on the South Island, the third of which occurs in the Tongariro National Park on the North Island. Cockayne (1908) described the presence in Tongariro National Park of sand drifts and dunes (2–3 m high) composed of volcanic ash, including andesite fragments, pumice, and scoria.

Examples of Aeolian Cross-bedded Volcaniclastic Strata

In some volcanic regions, units of cross-bedded volcaniclastic sand have been described and interpreted as being aeolian drift and dune deposits. These are often, though not always, deposits of pyroclastic material, called "eolian tuffs" by Smith and Katzman (1991). Examples of these have been noted in a variety of locations, among them are Arizona, New Mexico, and Hawaii, U.S.A.; Iceland; Madiera, Chile; and New Zealand.

Mafic Examples

Dark, presumably mafic "fossil dunes" were seen in a sea-facing cliff at Hellnaskagi, in Reynishveffi, Iceland (Thorarinsson 1981); there are apparently to such cross-bedded units, the youngest is about 500 years old. Several basaltic aeolian sand units (Pleistocene and Recent) have been found on the narrow, eastern peninsula of the Atlantic island of Madeira (Goodfriend et al. 1991). Another dark, mafic, cross-bedded aeolian sand deposit was described by Chenoweth and Cooley (1960) near Cameron, Arizona. The Arizona "Pleistocene cinder dune" apparently formed from basaltic ash derived from the San Francisco volcanic field, much as similar dunes occur today (Colton 1937, Cooley et al. 1969). Still other dark, basaltic aeolian cross-bedded sand units (Pleistocene) have been recognized in the Hawaiian Islands. Some of the Hawaiian sands were noted by Stearns (1940, 1946, 1947), while Porter (1991) presented information about specific basaltic aeolian sand units on the Island of Hawaii.

Intermediate and Sialic Examples

In the volcanic Hopi Buttes of northeastern Arizona, White (1989, 1990) identified several aeolian reworked tephra units, some of which are interpreted to have been dunes which migrated across eroded maar crater surfaces. Smith and Katzman (1991) described pumaceous, cross-bedded aeolian deposits in the Peralta Tuff Member of the Bearhead Rhyolite in the Jemez Mountains (associated with the Valles Caldera) of New Mexico (also see Smith et al. 1991). Their work formulates criteria to distinguish aeolian reworked tuffs from primary cross-bedded (e.g., base surge) deposits. Bell (1991) described cross-bedded Mesozoic aeolian sandstones in northern Chile which contain considerable amounts (> 50%) of volcaniclastic feldspar and rock fragments. Finally, several volcaniclastic aeolian units have been described in New Zealand; two such areas on the North Island are noted here. One occurs in the Taupo volcanic zone; it is known as the Mokai Sand Formation (1–2 m thick) and is a cross-bedded, medium to coarse sand consisting of sub-rounded pumice and glass fragments (Vucetich and Pullar 1969, Self and Healy 1987). The other is the Katikara Formation on the Pouakai Ring Plain, north of the Egmont volcanic region; this formation, thought to be > 12,500 years old, consists of probable aeolian reworked tephra which forms mounds 3–10 m high (Neall 1975).

Discussion

Sources of Volcaniclastic Sand

The sands which comprise terrestrial volcaniclastic dunes come from several different types of sources. The sands are largely epiclastic or pyroclastic in origin, and some dune areas incorporate both types of particle. Epiclastic grains are those that have been eroded from a lithic material, such as a lava, while pyroclastic grains are formed by an explosion at a volcanic vent (Fisher 1961, 1966). Pyroclastic sands may be reworked directly from the original pyroclastic deposit, while both epiclastic and pyroclastic grains can be deposited in stream or lake beds, beaches, or glacial sediments.

Direct aeolian reworking of a pyroclastic deposit is a mechanism suggested for the pumaceous cross-bedded units in the Jemez Mountains, New Mexico, described by Smith and Katzman (1991). Aeolian reworking of pyroclastic material has also been suggested for the origin of some of the sands in the Arequipa dunes in Peru (Finkel 1959, Hastenrath 1967), and is probably the manner in which the ash dunes in central Arizona, the basaltic dunes in the Ka'u Desert, Hawaii, and the small dunes on Surtsey and Isla San Benedicto formed (Colton 1937, Richards 1965, Gooding 1982, Ingólfsson 1982). It is probable that runoff from small streams has also contributed to the erosion and sorting of pyroclastic sands which make up these dunes; for example, there are minor amounts of red-colored sand and silt derived from the underlying Moenkopi Formation included in several of the ash dunes in central Arizona (A. Levine, personal communication 1990).

Epiclastic and pyroclastic volcanic sands may be deposited by fluvial systems, either in lake beds, stream beds, or deltas. Pyroclastic material may also be deposited in these environments by airfall. Deltas may provide sand for beaches, and wave action may sort sands of varying densities. Many of the volcaniclastic duns described herein were formed by wind erosion of a dry lake or beach deposit. Glacial activity may also have eroded lavas to provide sands which were subsequently removed by wind and water.

Examples of wind-reworked volcaniclastic lake sediments include the dunes in southeastern Oregon, which are composed of pyroclastic and epiclastic grains deposited in Pleistocene lakes (Dole 1942, Allison 1966). Epiclastic basaltic sediments deposited in Quincy Basin by the Spokane floods form the source for the Moses Lake, Washington, dunes (Petrone 1970). River sediments are also a source of volcaniclastic dune sand, as in the case of the Great Sand Dunes of Colorado, which are formed largely of San Juan volcanic materials carried by the Rio Grande (Wiegand 1977). Finally, a combination of fluvial-deltaic deposition, ocean current transport, and wave action have acted together to form localized concentrations of volcaniclastic, mafic, titanomagnetite sands on the western coast of North Island, New Zealand, and some of these deposits have subsequently formed dunes (Fleming 1953, Schofield 1970).

Composition and Morphology of Aeolian Volcaniclastic Deposits

The mineralogic compositions of volcaniclastic sands depend on the type of volcanic materials present in a given region. The compositions include the entire range from sialic to mafic. Silicic pumice and glass sand dunes and aeolian deposits occur in southeastern Oregon; Mono Lake, California; Jemez Mountains, New Mexico; and pumaceous dunes in part of the Manawatu sands of New Zealand. Trachytic pyroclasts formed dunes on Isla San Benedicto (Richards 1965). A variety of compositions, ranging from rhyolitic to basaltic, occur in lithic fragments and minerals of the Great Sand Dunes, Colorado. Mainly andesitic volcaniclastic materials contributed to the coastal dunes in New Zealand, and primarily basaltic sands comprise the volcaniclastic dunes in central Washington, Iceland, Hawaii, and central Arizona. These compositional differences apparently have no effect on the overall dune morphology. Basaltic dunes in central Washington exhibit typical transverse, barchan, and parabolic morphologies (Petrone 1970), and the mafic dunes in Peru are considered to be among the classic barchan forms (Finkel 1959, Hastenrath 1967).

Composition differences might, however, have some effect on the ripples which form on the surface of the dunes. Particles with different densities have different threshold velocities (Bagnold 1941, Iversen et al. 1976), and different transport rates (Willetts 1983, Williams 1987). Density differences might affect the manner in which the sands are sorted across a ripple surface; such is commonly seen in quartzose sands where fine magnetite grains become concentrated along ripple crests (Evans 1962, Sharp 1963). Grain sorting according to density was also noted in granule mega-ripples seen in Iceland and at the Mono Craters, California, where dark obsidian grains formed the crests and light-colored, low-density pumice was concentrated between the crests (Greeley and Peterfreund 1981, Peterfreund 1982). It was also noticed that the low-density pumice was coarser-grained than the obsidian (Greeley and Peterfreund 1981); whereas it is more common that coarser grained particles are concentrated along the ripple crests when the grains are of nearly uniform density (Sharp 1963). Wind tunnel simulations of condition on the planet Venus were seen to produce concentrations of heavy minerals on the stoss and crest of aeolian bedforms (Greeley et al. 1991). Sorting by sand density and composition is also considered to have occurred at the scale of an entire dune field in the Arequipa, Peru, region, where the abundance of dark, heavy minerals decreases downwind (Finkel 1959, Hastenrath 1967).

Volcaniclastic aeolian sands allow a unique opportunity to study the effects of particle density and density–particle size relationships in a natural setting, because there are many volcaniclastic dune areas where grains have very different densities (e.g., pumice and magnetite in the same sand). Particle density effects have not been studied in great detail for aeolian sediments (Willetts 1983); particle density effects have, however, been studied for volcaniclastic sediments in a fluvial regime (Smith and Smith 1985). Smith and Smith (1985) noted that sorting characteristics of volcaniclastic sediments may not be diagnostic of depositional environment, because different density (or specific gravity) particles may be in hydraulic equilibrium at different particle diameters (Rittenhouse 1943, Chepil and Woodruff 1963).

Implications for Mars

The aeolian sands of Mars are generally thought to be composed of non-quartz minerals, the dominant materials are probably mafic minerals and lithic fragments. Plagioclase in also a likely aeolian sand, as are opaque minerals such as magnetite and ilmenite. Many of the volcaniclastic aeolian sands on Earth provide a compositional analog to the possible martian sands. It is found that, contrary to common perceptions, many of the minerals that are easily chemically altered (such as plagioclase, olivine, and pyroxene) under terrestrial surface conditions, can be found forming dunes in a variety of environments, most of which are relatively arid at the present time. In addition, compositionally immature aeolian sandstones may survive many millions of years in some locations, such as the Mesozoic deposits in northern Chile (Bell 1991).

Some of the dark-hued aeolian materials on Mars, might, in fact, have a volcanic origin. This idea was first suggested by McLaughlin (1954, 1956a, 1956b), before spacecraft had visited the planet. Because volcanic landforms are relatively common on Mars (Greeley and Spudis 1981), they are a likely source for basaltic (and other compositions) sands. Geissler et al. (1990) have suggested that dunes on the floor of the Valles Marineris consist of basaltic glass pyroclasts. Simpson et al. (1982) proposed that dune material adjacent to the caldera in the center of Syrtis Major may be composed of aeolian–reworked basaltic ash and cinders.

Morphologically, the dunes on Mars are not significantly different from those on Earth (Cutts and Smith 1973, Breed 1977, Breed et al. 1979); and, as noted above, the morphologies of terrestrial volcaniclastic dunes are not significantly different from quartz, carbonate, evaporite, or clay dunes. Compositionally, however, the volcaniclastic dunes serve as the best analog to martian dunes. Field and remote sensing studies concerning the manner in which the different volcanogenic minerals and fragments are distributed in a dune field or sorted by the wind at the small bedform scale might provide further insight into the nature of the martian dunes.

Further Research

This report represents a first attempt to locate and synthesize the literature describing examples of volcaniclastic aeolian dunes. The authors have only field–checked the dunes occurring within the continental U.S.A. (Edgett et al. 1991). Further field studies of the dunes in Washington and Oregon, U.S.A., are in progress. The authors would appreciate correspondence with readers having knowledge of other examples of volcaniclastic dunes. We anticipate updating our review, perhaps after thermal infrared spectra or surface samples of the dunes on Mars have been obtained. Thus, maps, photographs, corrections or clarifications regarding our present review, and copies of papers about volcaniclastic dunes and their compositions would be especially welcome and appreciated.

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