INTERACTION OF SAND TRANSPORT WITH TOPOGRAPHY AND LOCAL WINDS IN THE NORTHERN PERUVIAN COASTAL DESERT

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Summary

The boundary-layer winds of the Peruvian coastal desert are remarkable for their constancy in direction and magnitude, as well as their aridity. Near the coastline winds nearly parallel the coast, but they are deflected upslope further inland. Sand deposits are ubiquitous and occur at three intergrading scales. Transverse barchanoid dunes and fixed dunes created by interaction of wind with bedrock topography occur at the smallest, micro scale. At the meso scale, larger topographic obstacles funnel the boundary layer winds and sand in transport around mountains and through cols. Fixed dunes may form before such obstacles due to blocking, and long sand tails often form downwind from the mountain in the zone of flow reconvergence. The sand tails often extend many times the height of the obstacle, and they grow due to self-reinforcing interactions of wind with the topography, possibly including induced thermal convection. At the largest, macro scale, the upslope component of upvalley anabatic winds has formed winding linear ridges of sand locally exeeding 500 m in thickness and 20 km in length along broad divides perpendicular to the coastline. Accentuation of the topography by sand deposition reinforces the upslope winds.

1. Introduction

The coastal zone of Peru and northern Chile is notable for its extreme aridity, amenable temperatures, consistent winds, and widespread eolian landforms. The sand deposits of the coastal desert interest geomorphologists because they have been little modified by vegetation or fluvial processes and because they have been formed by nearly unidirectional winds. Furthermore, unusual eolian landforms have been created by the local winds that result from interaction of the rugged topography with the trade wind circulation. Meteorologists have long been intrigued by the extraordinary coastal climate of the western South American coastline, but they have not realized the potential of the eolian landforms to provide detailed information concerning near-surface wind characteristics for the strong afternoon winds. The present paper discusses the interaction of sand transport and deposition with topography and local winds for about 225 km of the northern Peruvian coastal desert lying between Trujillo (8° 10' S) and Culebras (10° S).

2. Topography

The coastline of this portion of northern Peru trends NNW (Fig. 1). The topography is rugged, rising to peaks as high as 5000 m within 75 km of the coast. The main cordillera, with peaks reaching 6000 m, lies about 30 km further inland, separated from the coastal range by a valley paralleling the coastline. The topography within 15 km of the coastline is varied, with isolated peaks and ranges with relative relief locally exceeding 1000 m surrounded by rolling plains generally less than 600 m in elevation. These plains are generally marine terraces or

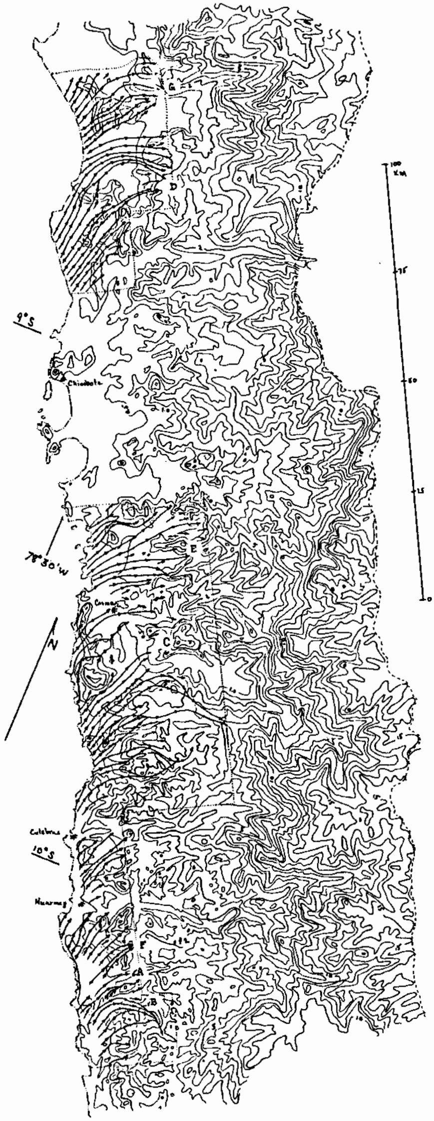


Figure 1. Map of the Peruvian coastline between Trujillo and Huarmey showing paths of net sand transport. Contours in feet x1000, with 500 ft contour interval below 2000 ft and 1000 ft interval above 2000 ft. Dashed lines show crest of first Andean summits (Cordillera Negra). Dotted lines show limits of areas mapped from aerial photography. Part of southern mapped area is shown in Figure 2, and the northern mapped area corresponds to Figure 3.

lowlands graded to former sea level stands (Grolier, et al., 1974). Bedrock knobs are locally exposed, but much of the area is covered by smooth or rolling alluvial and eolian blankets. The eolian mantle locally may exceed 200 m in thickness. Superimposed upon this topography are migrating dunes and topographically-controlled, generally linear sand deposits that range up to 150 meters in thickness. Spectacular linear mountains of sand locally exceeding 500 m in thickness and 20 km in length extend generally perpendicular to the coastline; these are generally positioned on interfluves.

3. Climate and Vegetation

This summary of climate and vegetation is based largely on descriptions of Prohaska (1973), Grolier, et al. (1974), Lettau (1978a,c), and Lettau and Costa (1978). The coastal desert is one of extreme aridity; in many places rainfall exceeds 1 mm only every decade or two. Most rainfall is limited to infrequent years of abnormal climate, the el Nino. However, between about 100 and 800 m above sea level winter stratus clouds intersecting slopes lead to drizzle and condensation that support lichens and other leathery plants called "lomas". Although some lomas forms locally on sand deposits, most of the eolian landforms are unaffected.

The most prominant feature of the coastal climate is a pronounced maritime inversion. The base of the inversion generally lies between 600 to 800 m in winter and about 200 to 500 m in summer, and the inversion ranges from about 1000 m thick in winter to about 200 m thick in summer. The winter inversion is stronger than the summer inversion. Stratus clouds below the inversion are nearly universal in winter, but occur primarily at night and morning during the summer. Although the inversion rises slightly inland, particularly where winds are funneled orographically, mountains higher than 1000 m project through the maritime layer, especially in summer. The most significant effect of the maritime inversion is to restrict the vertical extent of the trade winds and to limit convective penetration.

The boundary layer trade wind circulation along the Peruvian coast is remarkable for its consistency in direction and speed throughout the year. Mean 4 pm winds at Puerto Chicama (7 42' S) during southern summer (Dec, Jan, & Feb) average 8.2 knots, and winter (Jun Jul, & Aug) winds are only 2% less (Prohaska, 1973). The seasonal range of 7 pm winds at Lima (12 S) is slightly greater, with summer winds 13% greater than winter (Prohaska, 1973). Wind direction is likewise consistent, with a difference of only 3 between winter and summer 7 pm wind directions at Lima. Winds undergo a pronounced diurnal variation, with strong southerly winds during the afternoon and light winds at night and early morning. The afternoon winds are SSE to S, and thus have a slightly onshore component. The afternoon winds, at least during the summer, are generally strong enough to initiate sand transport at light to moderate rates. Radiosonde observations at Lima indicate that winds in the zone 1 - 3 km (above the marine inversion) winds are northwesterly, particularly during summer. Above 3 to 4 km winds are easterly, resulting from subsiding foehn circulation over the Andes from the Amazon basin. Again, this upper circulation pattern is most pronounced in summer, and results in precipitation and vegetative cover above 1500 to 2000 m, but no precipitation near the coastline because of adiabatic warming of the subsiding air. The northwesterly intermediate winds partially compensate the low-level southerly winds and partially the upper-level easterlies.

Lettau (1978c) suggests the coastal southerlies are a low-level jet oriented along the coastline within the maritime layer. A secondary, Hadley-cell counterclockwise (facing downwind) circulation creates a slight onshore surface component, updrafts inshore, downdrafts offshore and an oceanward return circulation in the upper boundary layer. Lettau suggests that the jet flow and cell extend inland on the order of 60 to 100 km. The jet flow is a thermal-tidal circulation induced by diurnal surface heating and cooling of the land relative to the ocean balanced by coriolis acelleration acting on a generally southerly airflow. Near-coastal upwelling of cold water induced by the flow is important in the thermal balance of the jet, and accounts for the aridity of the coastline.

At high elevations (>2.5 km) and at distances far from the coast surface winds are classic mountain-valley winds with daytime anabatic (westerly) and nighttime katabatic (easterly) winds (Lettau and Costa, 1978). Howell (1954) reports typical daytime upvalley and nightime downvalley winds in the Chicama Valley (just north of the study area), which is oriented perpendicular to the coastline. Secondary daytime upslope winds and nightime katabatic winds are superimposed upon the up- and down-valley circulation.

At locations intermediate between the coast and the higher mountains the wind pattern is more complicated. Along the southern coast of Peru at a latitude of about 16.7° S a broad, seaward-sloping plateau (the Pampa de La Joya) extends at elevations of 1000 to 1500 m from about 10 to 55 km from the coastline. Located on this plateau is the much-studied La Joya barchan dune field (see Lettau, 1978a for a summary bibiography). Nighttime winds are katabatic southeasterlies, but anabatic upslope winds are weak and limited to early-morning hours. The nighttime katabatic winds, although fairly strong, transport little sand due to low shear velocities resulting from the nocturnal inversion (Lettau and Lettau, 1978). The strong, barchan-moving afternoon winds are south-southeasterly, similar in orientation to the near-coast winds and essentially parallel to the contours and coastline (Lettau and Costa, 1978).

By contrast, the northern Peruvian coastline studied here is more compressed, since the 5000 m crestline lies only 75 km inland as compared to 105 km in southern Peru. The abrupt coastal escarpment and high-level plateau are absent along the northern coastline, replaced with discontinuous lowlands near sea-level separated by isolated coastal mountains (Fig. 1).

Since sand transport occurs primarily during summer afternoons, transport directions inferred from eolian landforms (see below) reflect surface winds at these times. Maps of sand transport direction for this area (Figs. 1 to 3) indicate a fairly gradual transition from southerly, slightly onshore winds near the coast to northwesterly to westerly winds within 10 to 15 km inland (as also noted by Craig and Psuty, 1962, Gay, 1962, and Grolier et al., 1974). This veering of afternoon surface winds results partly from transition from the coastal maritime

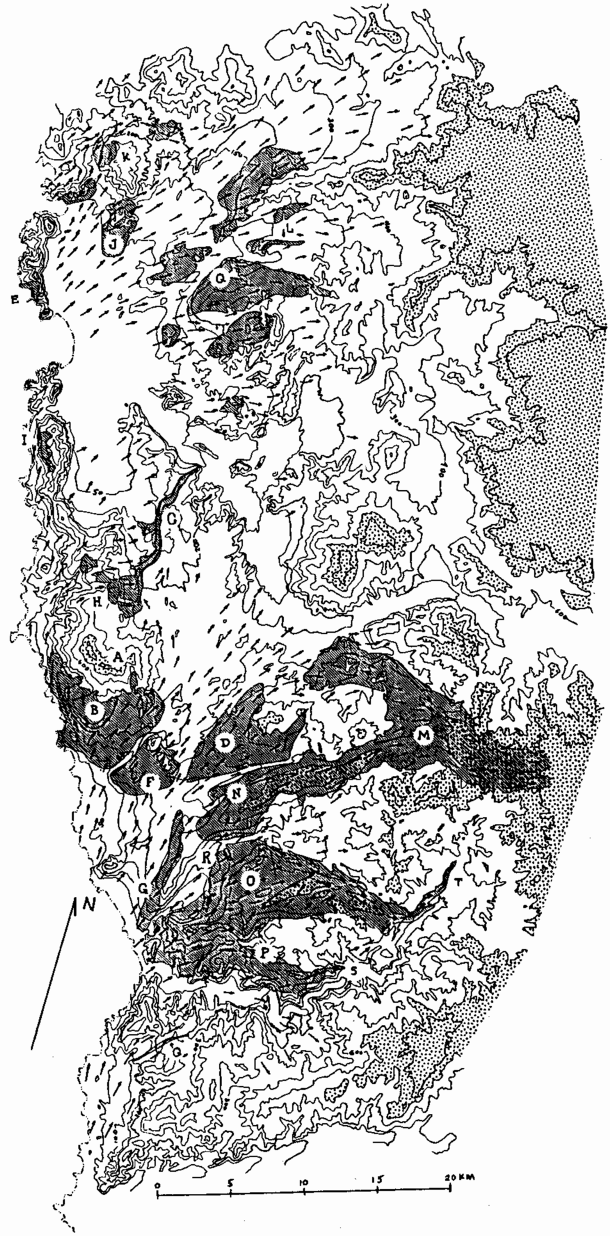


Figure 2. Major sand deposits and sand transport directions (arrows) for part of the Peruvian coastline shown in Figure 1. Ruled areas are major sand accumulations (generally thicker than 20 m). Contours at 50, 100, 150, 200, 300, 400, 600, 800, and 1000 m, with elevations above 800 m shown in dot pattern.

alongshore jet to anabatic flow inland and partly from penetration of the surface above the maritime layer into the overlying northwesterlies. Thus along the northern Peruvian coast the southerly coastal maritime jet is narrower than indicated by Lettau's (1978c) model and more restricted than further south along the Peruvian coastline.

4. Dune Forms

The coastal dunes of Peru offer a unique opportunity to study eolian processes due to the combination of amenable temperatures, hyper-aridity, easy accessability (along the Pan American highway), nearly daily eolian activity, and unidirectional winds. In addition, this area affords the opportunity to study the interaction of topography, meteorology, and sand transport at a variety of scales from individual barchan dunes to mountains of sand tens of kilometers in length. The information used in this study comes from a field expedition in October, 1971, supplemented by topographic maps at a scale of 1:100,000 and vertical aerial photographs at an approximate scale of 1:50,000. The limits of aerial photography available to the author are shown by dashed lines on Figure 1. The mapped area includes most of the eolian landforms along the coastal zone in the area shown in Figure 1 with the exception that some eolian deposits extend further inland than the mapped area in the northern portions of this figure and in Figure 3.

The stereo aerial photographs were used to prepare detailed maps of eolian landforms at a scale of 1:50,000. Among the features mapped included actively migrating, mostly barchanoid dunes, sand sheets without obvious dunes, fixed

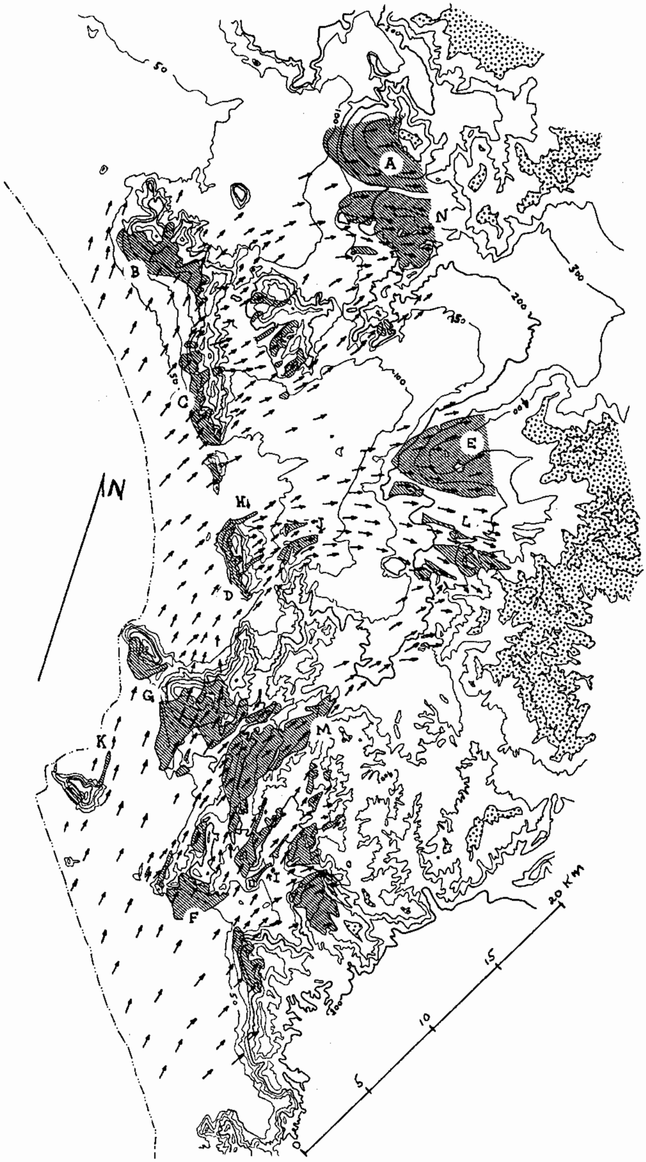


Figure 3. Major sand deposits and sand transport directions for the northern part of the Peruvian coastline shown in Figure 1. Explanation as for Figure 2 except that areas above 600 m in elevation shown in dot pattern. Note that sand deposits extend northeastward beyond the limit of available aerial photography near locations "A" and "E".

(climbing and falling) dunes, and the mountainous dune massifs. In addition, bedrock and alluvial flats were also mapped. Figures 2 and 3 are abstracted from the original maps to show major sand deposits and the trend of sand movement inferred from dune orientation, sand streaks, fixed dunes extending from topographic obstacles, and sand massifs. Individual migrating dunes and smaller fixed dunes are not shown in these figures. Due to the nearly unidirectional winds, transport directions inferred from the aerial photographs and shown in Figures 1 to 3 are thought to be within a few degrees of true net transport directions at the majority of locations. The mapped transport directions probably correspond to average afternoon winds, particularly those in the summertime. However, observed winds at any time would not be as consistent as the mapped tranport paths due to turbulent fluctuations in direction and velocity and slight systematic shifts in direction during the afternoon hours of maximum velocity. In particular, areas in the wake of topographic obstacles generally experience highly unsteady flow due to large-scale eddies.

As can be seen from these figures, a massive volume of sand is involved in these eolian deposits. The sand derives ultimately from erosional debris from the Andean mountains and more proximally from alluvial and beach sediments, active and relict. Since the net transport direction is mountainward and since some of the sand is delivered directly into major perennial stream valleys draining Andean headwaters, a portion of the sand is recycled through the eolian transport system. The mountainous sand deposits involve such vast accumulations of sand that the length of time required to form them is minimally several thousand years, although some may be much older and in a near-equilibrium state balancing sand gain and loss. The present hyperarid environment has probably existed for at least 5000 years, although the Pleistocene may have experienced periods of relatively moister conditions (Costa, 1978). Local valleys and washes along the coastline are commonly buried by modern eolian sediments with little evidence of fluvial reworking. The smaller fixed dunes probably have existed in near-equilibrium for as long as the present atmospheric circulation pattern has existed. Individual migrating barchans, on the other hand, have lifetimes of tens to hundreds of years, although the dune fields may be much older with loss of dunes due to dissipation, collision with obstacles, or migration into active alluvial valleys balanced with formation of new dunes near the sand source.

The interaction of sand transport with topography and local winds produces several types of eolian deposits associated with different types of interaction. The eolian landforms can be subdivided into three intergrading classes of increasing spatial scale correlated with the size of the landform relative to the maritime boundary layer. The smallest ("micro-scale") deposits are deeply submerged in the boundary layer and are similar to small eolian landforms elsewhere in the world in nearly unidirectional wind regimes. Intermediate, "meso-scale" deposits involve topographic features penetrating, or nearly penetrating, the maritime layer in which topographic channeling of winds and possible thermal convection become important. Finally, the mountainous, "macro-scale" sand massifs originate through circulation patterns involved with mountain-valley wind systems. Although the terms micro, meso, and macro are used here for convenience, from the meteorological perspective all of the atmospheric circulation patterns would be classified as "local" winds of micro and possibly meso scale.

4a. Micro-scale Eolian Landforms

Eolian deposits of micro-scale involve both mobile dune forms and fixed dunes associated with topographic obstacles. Micro-scale sand deposits are generally less than 20 m thick, although they may rest on thicker sand deposits and may be associated with topographic features up to 200 or 300 m in height. The mobile dunes consist of barchans where present as isolated dunes and as complex assemblages of barchanoid transverse dunes where the dunes are more tightly packed (Fig. 4). Longitudinal dunes unassociated with topographic obstacles are conspicuously absent. This reinforces the suggestion by Howard (1978) and others that barchan and barchanoid transverse dunes are the preferred dune form in unidirectional wind regimes. Micro-scale longitudinal dunes seem to require either bi- or multi-modal wind regimes (McKee and Tibbitts, 1964; Fryberger, 1979) or the presence of a stabilizing vegetational cover, as in the Simpson Desert of Australia and the southwestern United States. Peruvian barchans have been utilized in several studies of dune migration due to their abundance, accessability, and present-day activity as well as an adequate historical record (e.g., Finkel, 1959; Gay, 1962; Hastenrath, 1967; Lettau and Lettau, 1969, 1978). Among the insights offered by these studies are the inverse relation between dune height and migration rate,

Figure 4. Topographically-funneled sand deposits northeastward of locations "J" and "K" in Figure 2. Fixed lee dunes of micro to meso scale with superimposed migrating barchanoid dunes extend downwind from bedrock ridges (arrows) oriented at an angle to oncoming wind. Photograph looks upwind. Note isolated barchans on right side. Lee dunes similar to those shown in Figure 6c. Photograph from Grolier, et al., 1974.

the remarkable persistance of form and size during migration, the important contribution of dune translation (bulk flow or overturning) to net sand drift, and the predictability of migration rate using wind speed measurements. Grolier _et al._ (1974) note that the unequal migration rate of large and small barchans leads to collisions in crowded dune fields with the result that smaller dunes may be incorporated into larger dunes and that larger dunes may become distorted and possibly broken up by the effect on the windfield of a smaller dune approaching from upwind.

Isolated barchans in the study area commonly originate from calving from the end of a topographically-fixed lee dune (Fig. 5a). In some areas, such as the _La Joya_ barchan field of southern Peru, a calving source for the dunes is not evident, and they may arise "spontaneously", possibly at sites of distubance of the windfield by small irregularities on the desert floor. Transverse dune fields occur where large quantities of sand are in transport, generally resulting from topographically-induced sand flow convergence of meso or macro scale. In fact, meso- and macro-scale accumulations of sand generally have superimposed transverse dune fields of micro scale (Fig. 5b).

The most common micro eolian landform in the rugged coastal topography of northern Peru are non-migrating sand accumulations associated with topographic obstacles, generally bedrock hills. Such deposits have a rich and confusing terminology in the literature, being called "climbing" and "falling" dunes (despite their immobility), "lee" or "trailing" dunes if extending downwind from a topographic obstacle (although to the authors's

a

b

Figure 5. Sand tails of micro (a) and meso (b) scale. (a) Narrow, smooth lee dune extending downwind from small hill in foreground. Note barchans calving from end of sand tail; (b) Lee dune at location "I" in Figure 3 extending downwind from hill in background. Note superimposed transverse dunes.

knowledge "windward" or "leading" dunes have not been distinguished), "sand tails", "sand drifts", and "longitudinal dunes". The author prefers to restrict "longitudinal" to apply to dunes oriented downwind in the absence of a topographic control. The term "fixed dune" is used here for all such topographically controlled sand accumulations of micro and meso scale, with the caveat that fixed dunes involve active migration of sand over their surface as opposed to vegetatively "stabilized" dunes.

Most of the micro scale fixed dunes in the study area are probably close to an equilibrium balance in which the dune form and size is roughly constant through time and the dune surface is one of transport without net erosion or deposition. If a dune form is fixed, then the transport rate cannot change downwind except to balance convergence or divergence of the sand flow (as implied by Equation 1 of Howard, 1978, for a zero erosion rate). Such an equilibrium form is stable because enlargement of the fixed dune by deposition increases wind speeds over the surface due to greater exposure, resulting in compensating erosion. Similar but inverse reasoning applies to erosion.

Fixed dunes associated with topographic obstacles occur in locations where sand transporting capacity over the original topography becomes supersaturated. This can occur either due to downwind decrease in velocity or due to convergence of sand, or both, but will not occur until transport rates reach saturation. A variety of topographic forms can create such conditions. The simplest is a symmetrical smooth hill with maximum gradients less than about 20 degrees (Figs. 6a, 7). Blocking effects cause a

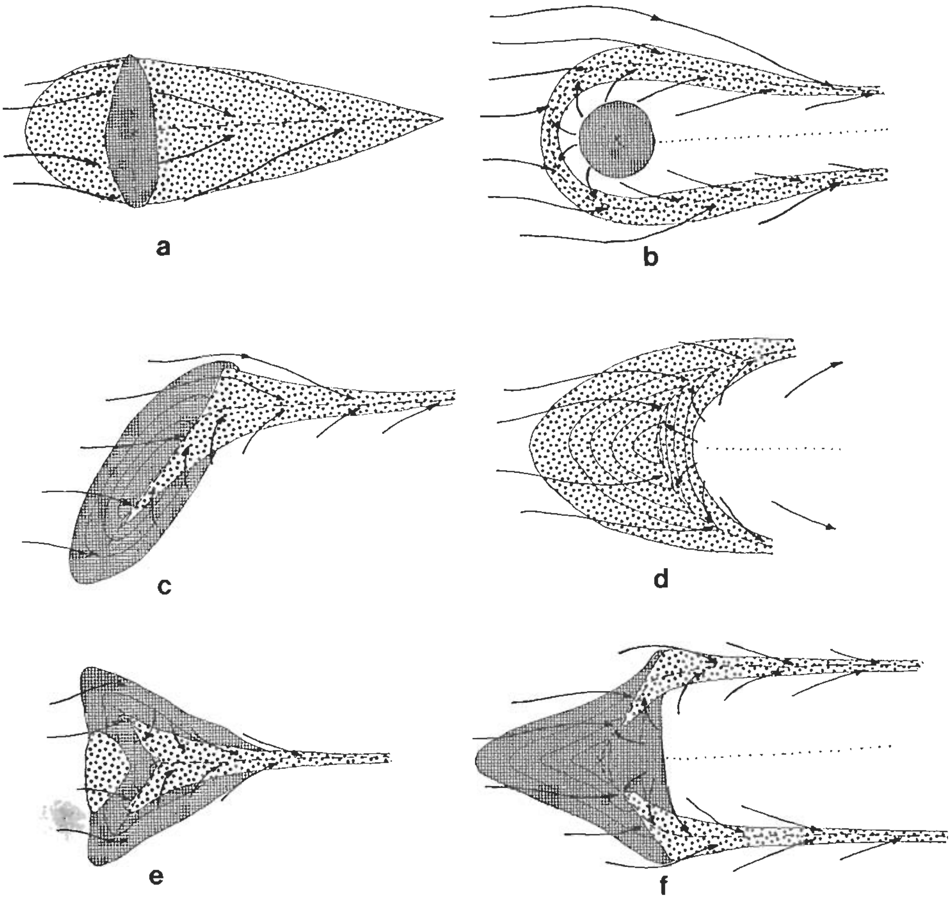
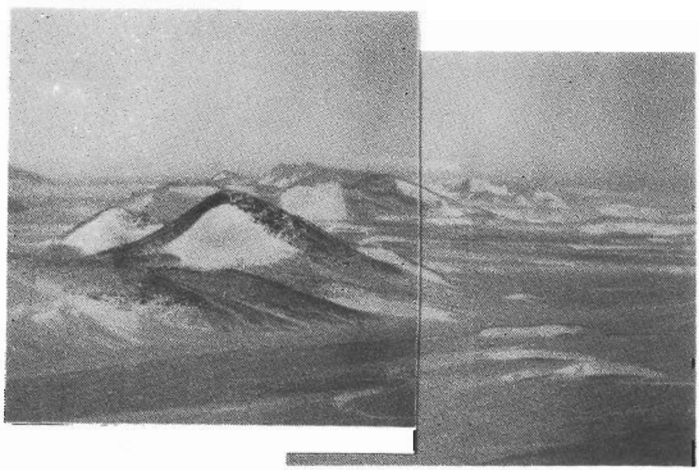


Figure 6. Schematic illustrations of micro-scale fixed dunes. Bedrock obstacles shown in cross-hatched pattern, and sand deposits in dot pattern. Contours on bedrock shown in (c), (e), and (f), and on barchan dune shown in (d). Arrows show surface wind directions. Dashed lines on sand show zones of convergent flow and/or flow separation. Dotted lines show zone of flow divergence.

low velocity zone near the leading edge of the hill, creating a streamlined sand fillet if the oncoming sandflow is sufficiently close to saturation. Note that this is a zone of flow divergence, which partially compensates for the lower velocity and generally results in a relatively small deposit, if any. The zone immediately behind the hill also experiences low velocities near the surface, compounded by convergent flow. Flow separation may or may not occur in this zone. A streamlined sand fillet forms which is generally higher and longer than the upwind deposit and which may occur even if there is no leading-edge deposition. Convergent winds created by wake vorticity encourage downwind extension of the lee dune. The lee dune generally gradually diminishes in size and dies out within a distance equivalent to several times the relative relief of the obstacle due to gradual decay of the vorticity (Fig. 7), although barchans may locally calve from the end of the dune (Fig. 5a). The downwind extension of the lee dune is encouraged by the sideslopes (which may reach the angle of repose), because net sandflow is diverted systematically downslope from the wind direction (Howard, 1977). That is, sand transport that on level surfaces would be directed strongly toward the center of the rear of the topographic obstacle by convergent winds is given a downwind component by the topographic gradient ("downwind" here refers to the direction of the undeflected oncoming wind). This mechanism acts in addition to the increased exposure to the wind of a higher sand body to limit the height of a lee dune.

Another simple form of fixed dune occurs when an elongated low hill (maximum gradients of about 20 degrees) is oriented with

a



b

Figure 7. Short fixed dunes of type shown in Figure 6a extending downwind from small bedrock hills. Wind moves from left to right in (a) and toward observer in (b). Area located south of Huarmey (Fig.1).

its short axis at an angle to the oncoming wind (Figs. 4, 6c, 8). The leading-edge fillet is generally absent, but a prominant sand tail extends downwind from from the trailing end of the hill. The diversion of the sandflow towards the downwind end of the hill occurs both because the flow upwind of the hill is imparted a component parallel to the long axis of the hill and because a large, asymmetrical vortex is generated behind the hill (Fig. 6c) which shepherds sand along the crest of the hill and downwind along the trailing fixed dune. A prominant sand "shadow" swept free of sand commonly occurs downwind of the central portions of such hills (Fig. 4). Several parallel elongated hills have created a considerable lateral migration and concentration of sand transport in the area shown in Figure 4. The sand diversion created by such topographic forms can be utilized to protect engineering works. A barchan dune (Fig. 6d) is essentially two mirror-image hills of the above type merged at their upwind side. Sand is diverted to flow from both wings of such a barchan, whereas a sand shadow extends downwind from the central portions of the dune. Figure 6e,f shows two slightly more complex topographic forms common to the study area that exhibit fixed dunes which embody the principles discussed above.

A topographic obstacle with a steep leading edge creates a cell of reverse circulation (the "horseshoe vortex") surrounding the base of the obstacle (Fig. 6b). Sand is deposited in the zone of convergence as well as extending downwind at the edges of the outward-directed secondary circulation in the wake of the obstacle. Because of the smoothly-rounded bedrock topography in the study area such dune forms are rare except at the scale of

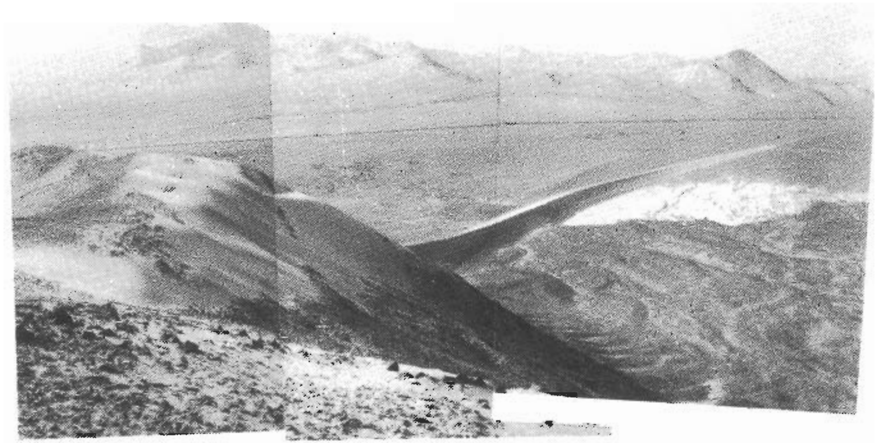


Figure 8. Narrow lee dune of micro scale similar to situation shown in Figure 6c. Note sharp crest and gradual diminishment of dune size downwind (to right) and sand accumulation downwind of hill crest in foreground. Dune located east of "F" in Figure 3.

individual boulders.

4b. Meso-scale eolian landforms

When the relief of the topography approches or exceeds the thickness of the maritime layer, the wind becomes increasingly constrained to flow around rather than over the topographic obstacle, much like water flowing through a rapids. And, like water in rapids, standing waves and hydraulic jumps are sometimes produced by interactions of the flow with the topography. However, due to the smaller density contrast between the maritime and the overlying layer than between water and air, the boundary layer circulation is more likely to be lifted through mountain passes and up valleys oriented parallel to the main flow. Penetrative convection into the overlying layer and dynamic mixing between the two layers is also more common.

The most spectacular example of meso-scale sand deposits are those associated with Cerro Las Lomas and associated peaks which rise to 1100 m on the coastline between Culebras and Casma ("A" in Fig. 2). Blocking of coastal southerlies by the mountain has created a streamlined, barchanoid hill of sand upstream from the mountain with a maximum relief of 400 m and a sand thickness almost as great ("B" in Fig. 2). The efficiency of the blocking effect is apparent from the absence of sand deposits at elevations above 500 m on the mountain itself. The crest of the sand hill is separated from the mountain front by about 1.5 km. The reason for this is uncertain, but probably involves one or more of the following mechanisms: 1) blocking effects, which cause slowing of wind approaching the mountain and a lateral diversion of sand around the mountain, possibly including the development of a horseshoe vortex; 2) erosion of sand by infrequent runoff from the mountain; 3) trapping of sand by Lomas vegetation, which is abundant on the sand hill, and 4) lateral diversion of sand around the mountain by the steep leading slopes of the sand hill. The large separation of the crest from the mountain and the smooth crest profile suggest that runoff erosion may be a contributing, but not dominant factor. The sand hill may now be at an equilibrium size, with additions of sand from upwind balanced by loss of sand at the wings where it flows around the two sides of the mountain. Forward growth on the southwest is restricted by the shoreline, and growth in height is probably limited by the thickness of the maritime layer. Similar, but less spectacular streamlined accumulations of sand in front of low mountains occur throughout the mapped areas, including at locations "D"-"K" in Figure 2 and locations "A"-"G" in Figure 3. Similar unlabeled deposits occur in both regions.

Downwind from <u>Cerro Las Lomas</u> extends the approximately 12 km-long narrow, wandering hill of sand called the <u>Cerro Manchan</u> ("C" in Fig. 2). The longitudinal profile of the hill is irregular, but sand thickness exceeds 100 m locally (Fig. 9). This is clearly a relatively fixed lee dune formed at the zone of reconvergence of the boundary layer flow diverted around the two sides of <u>Cerro Las Lomas</u>. However, this lee dune is distinguished from most other meso scale lee dunes by its narrowness relative to the size of the blocking mountain and by its wandering, nearly sinuous course.

The reasons for accumulation of sand into a narrow, steep, sharp-crested lee dune behind a blocking obstacle are not self-

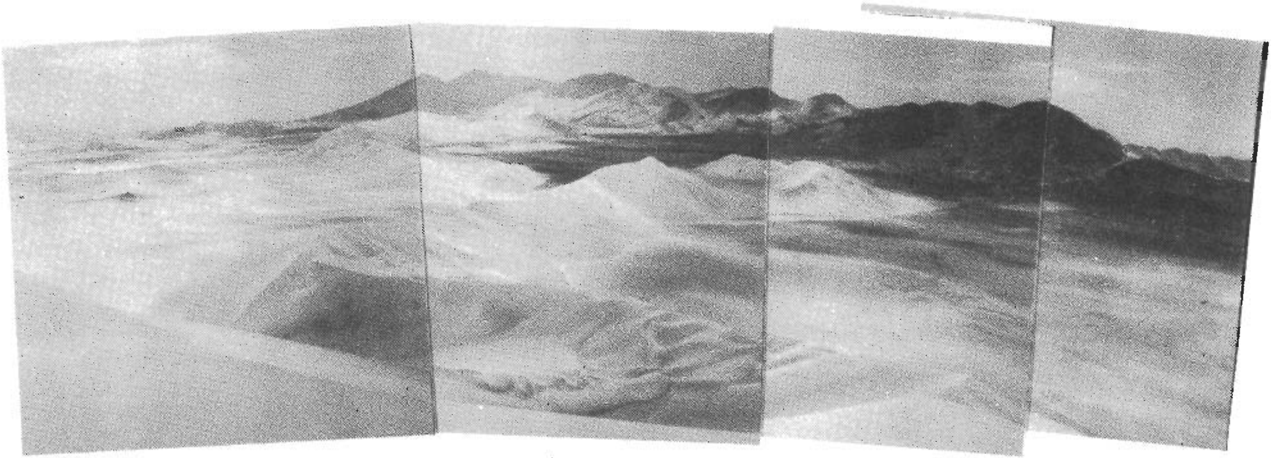


Figure 9. The wandering lee dune Cerro Manchan looking upwind towards Cerro Las Lomas. Note sharp crest and superimposed transverse dunes locally. View from near "C" in Figure 2.

evident. The wake generated by an isolated mountain penetrating the boundary layer is generally characterized by formation of horizontally-oriented vortices (as opposed to the more steady roll-type vortices discussed earlier) which migrate downwind. The zone of reconvergence is therefore one of unsteady flow, with wind coming from one side of the mountain alternating with wind from the other side. Without other factors intervening, a diffuse sand accumulation should result. Several mechanisms might be acting to create a well-defined dune:

1). Once a sufficient sand accumulation has occured, the crestline becomes a locus of separation for flow arriving at an angle to the crestline. As with a barchan, this separation encourages sand deposition, adding to the height of the dune (Tsoar, 1982).

2). The steepness of the dune deflects the sandflow downslope as discussed by Howard (1977). Since the wind direction has an upslope component, this results in a downwind sand drift, lengthening the dune in the downwind direction. The separated vortex on the reverse side of the dune has a upslope component subject to the same mechanism (Tsoar, 1982).

3). The crest of the dune may serve as a focus for convective eddies which therefore increase and localize the convergence effect. Several reinforcing factors may contribute to this, including the slope of the dune which encourages upslope air motion, the high afternoon sand temperatures due to high thermal inertia (Lettau, 1978b, notes temperatures reaching 69 C on dune slipfaces at the Pampa de La Joya), and possible enhancement of heat transfer between the radiationally-heated dune surface and the lower atmosphere as a result of particle motion in saltation. In addition, meso-scale lee dunes generally have superimposed transverse dunes (Figs. 4, 5b). Turbulence generated by the dunes enhances thermal transfer and thus convective tendencies.

4). The dune chain may serve as a locus for a hydraulic jump of a supercritical boundary layer flow descending from mountain passes. The resulting abrupt decelleration of flow would encourage deposition. Such a mechanism could only operate locally, but might occur at Cerro Manchan due to funneling of wind through cols of the hills extending NNW from Cerro Las Lomas.

Micrometeorological observations are insufficient to evaluate the likelihood of these several mechanisms.

The lee dunes of meso scale are generally distinguished from those of micro scale by the presence of superimposed transverse dunes, a larger size relative to the topographic obstacle (with the obvious exception of Cerro Manchan), and a linear extent that may exceed 20 times the height of the obstacle (Examples include "L" and "Q" in Figure 2, "H"-"L" in Figure 3, and Figures 4, 5b, and 9). This scale-dependent difference in morphology suggests that one or more of the convective mechanisms in 3), above, may be operating, because these would be favored by the larger spatial scale. Enhancement of convective motions may contribute to the formation of most dunes of draas size. Hanna (1969) and many authors since have suggested that the large longitudinal dunes of the Saharan deserts are created by convective eddies (roll vortices) of a size determined by the thickness of the boundary layer. This suggestion seems inconsistent with observations that the wind regime of areas of longitudinal dunes is generally bimodal, without necessarily having a strong component directly along the dune (Fryberger, 1979). However, micrometeorological observations in areas of draas dunes that might resolve this question are lacking.

Although some longitudinal dunes have a natural sinuosity (Tsoar, 1982), lee dunes in the Peruvian desert are generally nearly straight-crested. The unusual wandering or sinuous course of Cerro Manchan seems best explained as the result of spatial variability of intensity of the wind arriving from opposite sides of Cerro Las Lomas. Inspection of aerial photographs suggests that eastward deflections of the sand crest occur downwind of cols in the hills NNW of Cerro Las Lomas. Hills on the eastern side of Cerro Las Lomas are low and unlikely to create appreciable spatial variability of wind strength.

4c Macro-scale Dunes

The vast linear sand deposits extending generally eastward or northeastward from the coastal zone are aptly called "mountains of dunes" (Cerros Los Medanos) by local residents. Such macro dunes occur at "M"-"Q" in Figure 2 and at "M" and possibly at "E" and "N" in Figure 3, and at "A" and "B" in Figure 1. These dunes are the focus of strongly convergent sandflow (Fig. 1) and they tend along and towards broad salient ridges extending southwestward from the Andes. Even where massive macro dunes are not found, the broadscale sandflow pattern of the coastal desert is convergent flow towards salient ridges (for example, at "D"-"G" in Figure 1). This pattern is best explained as resulting from anabatic flow, with sand deposition occuring both because of flow convergence and because of low velocity wind along ridgecrests as compared to the concentrated flow along the main valleys (convective upwelling diminishes longitudinal velocities). As noted by Wilson (1971) and Fryberger and Ahlbrandt (1979) major accumulations of sand are generally in areas of relatively little net sand drift. The accentuation of the topography by sand deposition reinforces the upslope winds and contributes to enlargement ofthe sand massifs.

The valleys trending perpendicular to the coastline are generally kept clear of sand both by the high wind velocities along the valley axes and by the divergent flow (along major valleys fluvial removal is also important). As an indication of the strong flow along valleys, the valley at "R" in Figure 2 is fairly clear of sand although positioned between two sand massifs. Also the ends of the two macro dunes at "S" and "T" in Figure 2 are diverted counterclockwise where they intersect the northeastward-trending valley.

In summary, areas of positive relief can have contradictory influence on sand movement and accumulation depending upon their size, orientation, and steepness. Abrupt hills or mountains tend to block wind and thus divert sandflow, so that sand deposits occur before, behind or around the obstacle, but not on it. By contrast, large, broad ridges elongated in the direction of net sand transport tend to accumulate sand on their ridgecrests due to anabatic wind flow.

5. Conclusions

The windflow patterns revealed by sand deposits are mostly consistent with the descriptions and models of coastal climate advanced by Prohaska (1973), Lettau and Costa (1978), and Lettau (1978a,c). The major discrepancy compared to the area of southern Peru studied by Lettau and Costa are the rapid but smooth transition from the along-shore maritime jet to mountain-valley winds within a few kilometers inland. Both Prohaska and Lettau and Costa stress that the stability of the maritime layer caused by upper-air subsidence limits penetrative convection and contributes to the the regional aridity. However, the sandflow patterns indicate a considerable degree of convergence and, presumably, upwelling of (summer) afternoon winds along salient ridges of the Andes.

The Peruvian coastal desert offers a unique opportunity to study the interaction of sand transport with topography and meteorology. Particularly attractive are the accessability of the region, the consistency and unidirectionality of winds, the moderate temperatures, the abundant sandflow, and the rich variety of topographic forms and meteorologic phenomena. Field measurement of sand transport and associated meteorological phenomena in the spirit of the Lettau studies could resolve important questions not only about sand transport but also about subtle controls of atmospheric circulation.

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7. References

Costa, J. R. (1978): Paleoclimatology, oceanography, el Nino, and climate-related resources of the arid zones of Peru and Northern Chile. in Lettau, H. H. and Lettau, K., Exploring the World's Driest Climate, Center for Climatic Research, University of Wisconsin, Madison, 156-162.

Craig, A. K. and Psuty, N. P. (1968): The Paracus papers, Studies in marine desert ecology. Occasional Publ. No. 1, Dept. of Geography, Florida Atlantic University, Boca Raton, Florida, 1, 196 p.

Finkel, H. J. (1959): The barchans of southern Puru. Jour. Geology, 67, 614-647.

Fryberger, S. G. (1979): Dune forms and wind regime. in McKee, E. D., editor, A Study of Global Sand Seas, U. S. Geological Survey Prof. Paper 1052, 137-169.

Fryberger, S. G., and Ahlbrandt, T. S. (1979): Mechanisms for the formation of eolian sand seas. Zeitschrift fur Geomorphologie, 23, 440-460.

Gay, Parker, Jr. (1962): Origin, distribucion, y movimiento de las arenal eolicas en el area de Yauca a Palpa. Anales Segundo Congreso Nacional de Geologia, Parte II, Sociedad Geologica del Peru, 37, 37-58.

Grolier, M. J., Ericksen, G. E., McCauley, J. F., and Morris, E. C. (1974): The Desert Land Forms of Peru: a Preliminary Photographic Atlas. U. S. Geological Survey, Open File

Report, Atrogeology 57, Flagstaff, Arizona, 146 p.

Hanna, S. R. (1969): The formation of longitudinal sand dunes by large helical eddies in the atmosphere. Jour. of Applied Meterorology, 8, 874-883.

Hastenrath, S. L. (1967): The barchans of the Arequipa region, southern Peru. Zeitschrift for Geomorphologie, 11, 300-331.

Howard, A. D. (1977): Effect of slope on the threshold of motion and its application to orientation of wind ripples. Bull. Geol. Society of America, 88, 853-856.

Howard, A. D., Morton, J. B., Gad-el-Hak, M., and Pierce, D. B. (1977): Simulation model of erosion and deposition on a barchan dune. Sedimentology, 25, 307-338.

Howell, W. E. (1954): Local weather of the Chicama Valley (Peru). Archiv fur Meteorologie, Geophysik, und Bioclimatologie, Ser B., 5, 41-51.

Lettau, H. H., (1978a): Introduction. in Lettau, H. H. and Lettau, K., Exploring the World's Driest Climate, Center for Climatic Research, University of Wisconsin, Madison, 12-29.

Lettau, H. H., (1978b): Extremes of dirunal surface temperature ranges. in Lettau, H. H. and Lettau, K., Exploring the World's Driest Climate, Center for Climatic Research, University of Wisconsin, Madison, 67-70.

Lettau, H. H. (1978c): Explaining the world's driest climate. in Lettau, H. H. and Lettau, K., Exploring the World's Driest Climate, Center for Climatic Research, University of Wisconsin, Madison, 182-248.

Lettau, H. H., and Lettau, K. (1969): Bulk transport of sand by barchans of the Pampa de La Joya in southern Peru.

Zeitschrift fur Geomorphologie, 13, 182-195.

Lettau, H. H. and Lettau, K. (1978): Experimental and micrometeorological field studies of dune migration. in Lettau, H. H. and Lettau, K., Exploring the World's Driest Climate, Center for Climatic Research, University of Wisconsin, Madison, 110-147.

Lettau, H. H. and Costa, J. R. (1978): Characteristic winds and boundary layer meteorology of the arid zones in Peru and Chile. in Lettau, H. H. and Lettau, K., Exploring the World's Driest Climate, Center for Climatic Research, University of Wisconsin, Madison, 163-181.

McKee, E. G. and Tibbitts, G. C. (1964): Primary structures of a seif dune and associated deposits in Libya. Jour. Sed. Petrology, 52, 823-831.

Prohaska, F. J. (1973): New evidence on the climatic controls along the Peruvian coast. in Amiran, D. H. K. and Wilson, A. W., editors, Coastal Deserts: Their Natural and Human Environments, Univ. of Arizona Press, Tuscon, 91-107.

Tsoar, H. (1982): Internal structure and surface geometry of longitudinal (seif) dunes. Jour. Sed. Petrology, 52, 823-831.

Wilson, I. G. (1971): Desert sandflow basins and a model for the development of ergs. Geographical Jour., 137, 180-197.