Katabatic Winds in the Equatorial Andes[1]

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ABSTRACT

Diurnal air tides that carry Pacific maritime air eastward through passes and over lower portions of the continental divide are remarkably similar over long stretches of the Andes from northern Chile to central Colombia, embracing both desert and jungle climates on the Pacific Coast. Even where the maritime air is at its warmest, it overflows the divide as a cold current producing katabatic flows down the east slope of the range and often producing hydraulic jump phenomena in the valleys immediately to the east. The significance of this pattern for the rainfall climate of western Colombia is discussed.

1. Introduction

This study originated in connection with a rainfall stimulation program in the Pacific slopes of the Andes in western Colombia (López and Howell, 1965). Since the target area was mostly uninhabited jungle of very difficult access, an attempt was made at the beginning of the program to supplement the seeding by releasing additional silver iodide smoke from ground generators located in the more accessible east side of the range in the expectation that thermally-driven afternoon upslope winds would carry the smoke to the target west of the divide by way of the prevailing easterlies aloft. The upslope winds indeed occurred as expected after mid-morning, but a sudden reversal of wind direction often took place early in the afternoon, and relatively cool winds blew downslope, at times with remarkable force, making it impractical to carry on the afternoon missions originally planned. Similar katabatic winds had been observed by one of the authors in northern Peru (Howell, 1953), but their presence so far north was a distinct surprise. In addition, as more observations became available, it was realized that these winds play an important role in the microclimate of the eastern slopes of the western Andes and in the triggering of convection in the Cauca Valley east of the range. The present study is an attempt to analyze the dynamics of the observed flows, and to suggest the origin of these winds and related questions.

2. Physiography and climate

The area in which the winds were studied is shown schematically in Fig. 1. A notable topographic feature is the Cauca Valley, which is 230 km long, about 35 km wide, and slopes gently northward with an average elevation of 1000 m narrowing abruptly to almost a gorge 180 km north of Cali. The valley is surrounded on both sides by the central and western Cordilleras, which branch out from the main Andean chain 250 km south of Cali. The central Cordillera is the higher, with a mean ridge elevation of about 3500 m and peaks which exceed 5000 m. The western Cordillera is much lower, the average ridge elevation being about 2300 m, its main peaks reach just over 4000 m, and it has many passes under 2000 m.

A general easterly flow prevails aloft, and two main rainy and dry seasons alternate with the seasonal migrations of the Intertropical Convergence Zone over the latitude of the valley. Westerlies can occur occasionally above 4 km, but without any set rules, although the presence of a deep southwesterly flow in the rainy season usually indicates an advance of the Intertropical Convergence Zone and is generally accompanied by widespread rains. The Andean chains present a very effective barrier in the lower levels to the prevailing easterly flow, with the result that the Pacific side is dominated in the lower levels by the coastal influence. The meteorological aspects of this influence have been described in detail elsewhere (López, 1966).

The Cauca Valley, sheltered between two main ranges, develops weather characteristics of its own in which local convection is unusually responsive to the initial impulses. These impulses may come from either large-scale traveling disturbances such as easterly waves or may be local in nature. One interesting feature of the local climate is the frequent occurrence of vigorous late afternoon or early evening convective developments in localized areas of the valley. Similar developments have been observed elsewhere in Colombia (López and Howell, 1965) whenever a relatively narrow valley occurs between two main ridges. Since the intensification in the convection often takes place near sunset, explanations based on the development of valley winds

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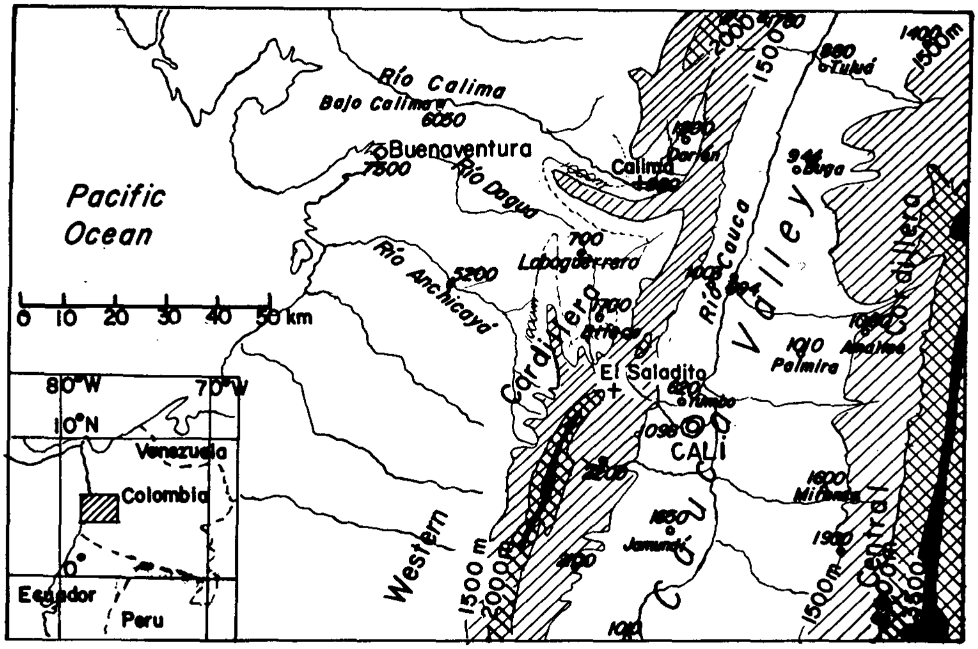

FIG. 1. Map of the area showing approximate topography (m) and average annual precipitation (mm).

or strong radiational cooling of the cloud cover over the valley are ruled out. Cool gusty winds also frequently blow downslope in the late afternoon from the western Cordillera and it seemed logical to investigate if there was a connection between the two phenomena.

3. Description of the katabatic winds and related phenomena

The cold waters of the Humboldt Current maintain an inversion over the coast of South America throughout the year. The Intertropical Convergence Zone never wanders south of the equator here and quite frequently cool Southern Hemisphere recurved trades (the Southwest Monsoon) invade the offshore waters of western Colombia where they arrive with a large moisture content after their long overwater trajectory. Moist air in the Pacific marine layer under the inversion is able to penetrate deeply inland because of the weakness of the Coriolis acceleration in these low latitudes, aided by a sea-breeze effect. The already relatively cool air is further chilled as it advances inland by partial evaporation of the copious orographic-induced precipitation. The temperature at Buenaventura, on the Pacific coast, for instance, averages 4C less than similarly wind-exposed areas at sea level on the Atlantic.

The Pacific air flows inland in a sort of diurnal tide. Measurements made with pilot balloons at Buenaventura have shown a sharp wind direction reversal connected with a sea-breeze effect at a height of about 1500 m. The advancing air tide is well known locally, and long-time inland residents speak of the arrival of "la marea," literally meaning tide, when referring to the morning wind. The advance of the air tide is easily traceable in the cloud motions along the road from Buenaventura to Cali, the depth of the oncoming air diminishing steadily as one advances inland. The stability of the air mass is such that the airflow tends to follow the ground contours closely in spite of the rugged terrain, so that most of the cloudiness at lower elevations is stratiform. The equivalent potential temperature increases an average of 10K between Buenaventura and Cali, showing that water vapor is constantly added to the air during its passage over the jungle. The air is dammed against the western Cordillera and eventually overflows the mountain passes and lower ridges, descending as a katabatic wind east of the range where it interacts with the easterly upslope mountain wind. The strength of these katabatic winds varies from day to day and between the rainy and dry seasons. Fig. 2 shows the schematics of the usual late afternoon flow as determined from pilot balloon and cloud drift observations. Fig. 3 shows the afternoon aspect of the ridges at the Calima River hydroelectric plant, about 50 km north-northwest of Cali, where katabatic flows are especially marked because the Calima River actually cuts here across the western Cordillera, forming a low pass.

The formation of reversed convective cells in the late afternoon over the western Cordillera is favored by the change of circulation induced by the downslope katabatic flows. A marked increase in the vigor, or a rejuvenation of afternoon convection when the katabatic flow is well established, takes place in the higher parts of the Pacific slopes of the western Cordillera and in certain areas in the eastern part of the Cauca Valley.

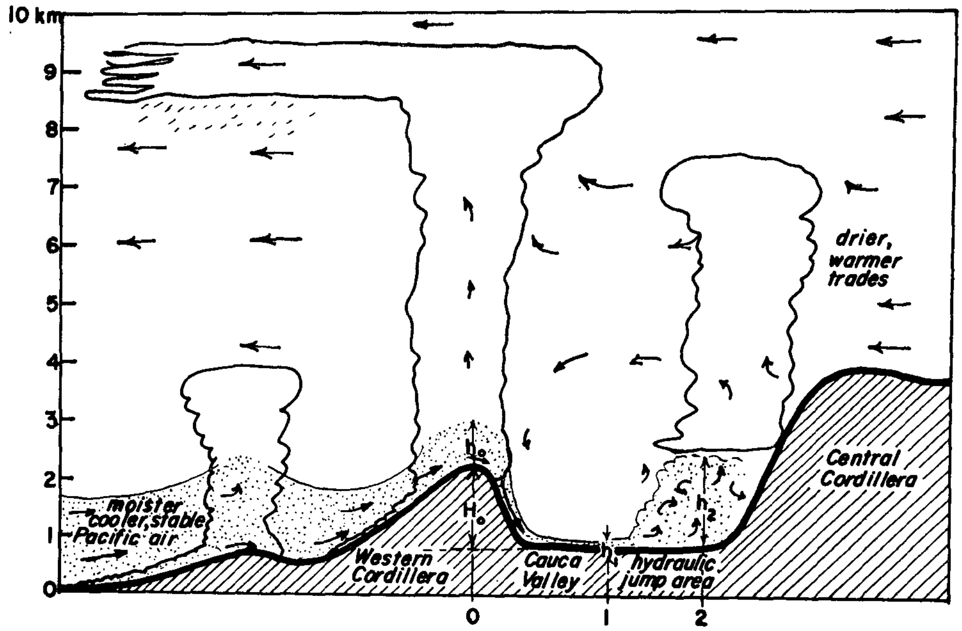

FIG. 2. Schematics of the late afternoon flow.

4. The heated hydraulic jump

In the following exposition we will follow, with small variations, the development given by Kuettner (1959).

When water flows through a weir out of a large reservoir of depth H into a sloping channel, a critical height, $h_0=2/3H$, for which the outflow reaches a maximum, is automatically established over the weir, and this maximum discharge takes place at a definite critical velocity

$$v_0=(gh_0)^{\frac{1}{2}}. \tag{1}$$

This critical velocity is also the maximum speed of propagation of long surface waves. If the channel floor slopes downward, gravity accelerates the shooting flow to a supercritical velocity, and disturbances created by the eventual return to subcritical velocity further downstream cannot penetrate upstream.

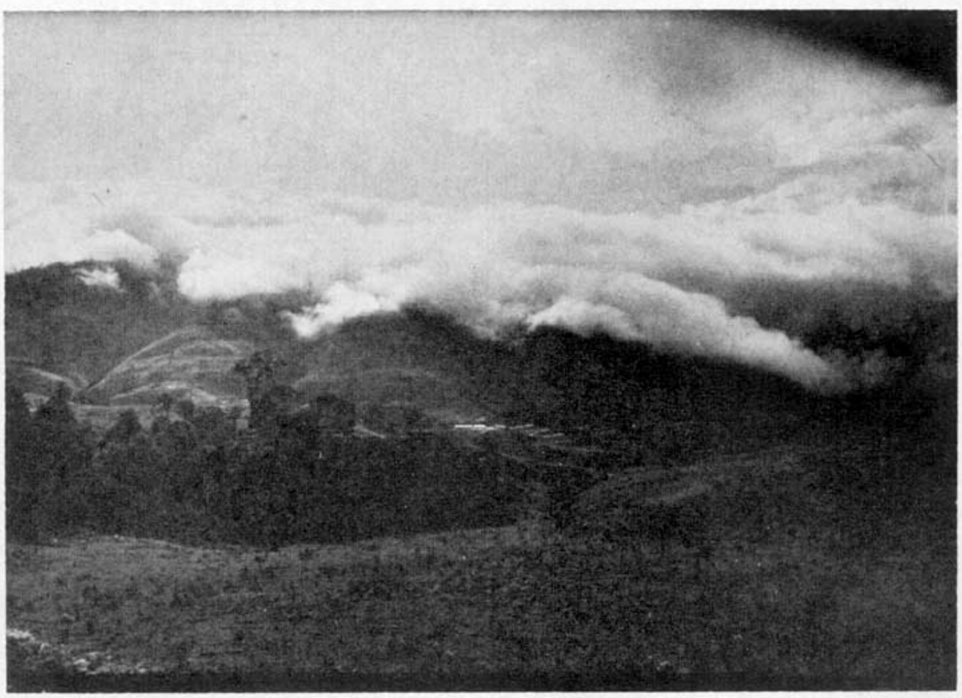

FIG. 3. Afternoon cloud cap over Rio Bravo divide accompanying katabatic wind flow at Calima River hydroelectric plant.

When cool air is considered instead of water, gravity acts on the density difference between the air strata in accelerating the flow, and g in (1) is replaced, at least for relatively small vertical displacements, by

$$\gamma=(\Delta\theta/\theta)g, \tag{2}$$

no longer constant, but dependent on the proportion of potential temperature increase. The critical spill velocity (1) then becomes

$$v_0=(\gamma h_0)^{\frac{1}{2}}. \tag{3}$$

Let us compare two cross sections 0 and 1 (Fig. 2), assuming a frictionless streamline flow. If H_0 is the height of the pass over the valley floor, we apply Bernoulli's equation to cross sections 0 and 1 to obtain

$$\gamma(H_0+h_0-h_1)=1/2(v_1^2-v_0^2). \tag{4}$$

The equation of continuity requires that

$$\text{discharge } Q=v_1h_1=v_0h_0=\text{constant}. \tag{5}$$

Combining (4) and (5) with the critical velocity (3) we arrive at

$$(h_0/h_1)^3-[2(H_0/h_0)+3](h_0/h_1)+2=0. \tag{6}$$

The Froude number, $F_1^2=v_1^2/\gamma h_1$, expresses the degree to which the flow is supercritical in section 1. The ratio (h_1/h_0) is always less than 1 and equals the shrink ratio between the depth of the supercritical shooting flow at the base of the pass and the depth of the overflowing air.

As the valley floor flattens, the supercritical velocity of the flow is rapidly diminished by friction and a "tranquil" subcritical flow region is approached. In order to satisfy continuity requirements, the depth of the flow must increase from h_1 to a certain h_2 at section 2. If this happens, waves can travel upstream in the subcritical area near h_2, but cannot do so in the supercritical area above h_1. The wave front will therefore steepen until it breaks down somewhere between cross sections 1 and 2. In this way, a hydraulic jump is formed which tends to remain stationary. This jump is often referred to as the "pressure jump" when it occurs in the air.

So far, addition of heat to the descending air has been neglected. This is unrealistic, since insolation on the slopes and the valley floor is usually strong. Following is the potential temperature distribution between El Saladito, at the pass, and Cali at the bottom of the slope, on a typical dry-season afternoon.

h(m)	t(°C)	θ(°K)
1100	30.6	314
1200	28.6	313
1400	25.8	312
1600	22.8	311
1800	20.0	311

There was an increase of about 3K in potential temperature during descent of the air. It is not unrealistic to estimate that at least half of the inversion is destroyed beyond Cali in the late afternoon by vigorous heating and mixing in the valley floor. The resulting reduction of the difference in densities between air masses reduces the gravitational constraint and allows the jump to bounce to greater heights. This, of course, presupposes that the destruction of the inversion takes place only in the heated area.

Comparing sections 1 and 2, the modified gravity is changed from

$$\gamma_1=(\Delta\theta_1/\theta_1)g \quad \text{to} \quad \gamma_2=(\Delta\theta_2/\theta_2)g. \tag{7}$$

The "heating rate" β, or amount of destruction of the inversion is given by

$$\beta=\frac{(\gamma_1-\gamma_2)}{\gamma_1}. \tag{8}$$

The streamline flow is destroyed in the jump area but we can apply the momentum equation which states that the net force acting on the vertical cross section to the right and left of the jump equals the rate of change of momentum, i.e.,

$$Q(v_1-v_2)=1/2(\gamma_2h_2^2-\gamma_1h_1^2)=h_1v_1^2-h_2v_2^2, \tag{9}$$

which combined with (8) and (5) yields

$$(h_2/h_1)^3-[1+2(h_0/h_1)^3](1-\beta)^{-1}(h_2/h_1)+2(1-\beta)^{-1}(h_0/h_1)^3=0. \tag{10}$$

Solutions of this equation can be found more easily in parametric form by making the substitutions

$$(h_2/h_1)=C>1 \tag{11a}$$

$$\begin{aligned}(h_0/h_1)^3&=(v_1/v_0)^3=A^3\\&=(1/2)C(C-1)^{-1}[C^2(1-\beta)-1]>1\end{aligned} \tag{11b}$$

and using as dependent variables

$$\left.\begin{aligned}(h_0/H_0)&=2A/(A^3-3A+2)\\h_2/(h_0+H_0)&=2C/(A^3-A+2)\end{aligned}\right\}. \tag{12}$$

There is another limitation on the parameter C, besides the obvious physical one of $C>1$, since by (11b), A must also be greater than one. The limiting values of C for various heating rates β are

$\beta=0$	0.2	0.4	0.6	0.8	0.9	0.95	$0.99\to1$
$C>1$	1.40	1.76	2.31	3.48	5.11	7.39	$17.00\to\infty$.

These limiting values of C correspond physically to an infinitely deep overflow. The nomogram (Fig. 4) has been prepared to solve Eq. (10) graphically for different heating rates. Additional scales for $1/A=(h_1/h_0)=(v_0/v_1)$ and for the ratio $C=(h_2/h_1)$ have been incorporated to

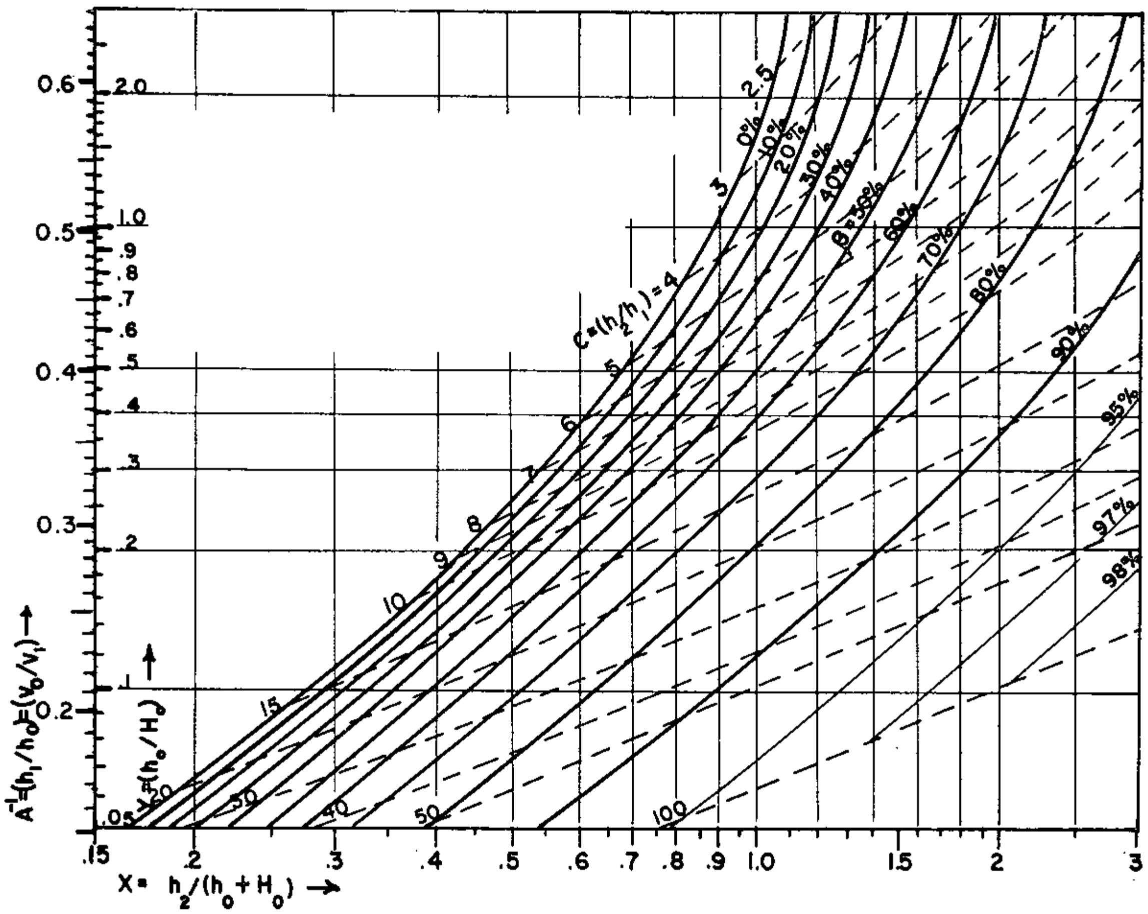

FIG. 4. Universal nomogram for the determination of the heights of the descending shooting flow h_1 and of the hydraulic jump h_2 in terms of the depth h_0 of the overflowing colder air, the height H_0 of the base of the overflow over the valley floor and of β, the amount of destruction of the overlying inversion. A^{-1} is the contraction ratio of the descending flow and C is the expansion ratio at the jump.

facilitate estimates of the increase in wind speeds, the contraction ratio of the descending flow and of the expansion ratio of the latter at the jump. In general, the spilling air usually has an initial velocity and this will result in practice in slightly higher values of the critical overflow depth h_0 and of the critical velocity v_0.

Fig. 4 shows that the height of the jump is quite sensitive to the amount of destruction of the overlying inversion. For instance, a nearly 50 per cent greater jump would be realized if half of the inversion is destroyed by local heating.

5. Application of the theory of the heated hydraulic jump to the katabatic flows observed over the western Andes

Measurements were made at various elevations, both during the dry and wet seasons, of the speeds of the spilling air, depths of the overflows and potential temperatures between El Saladito and Cali. No upper air temperature measurements are available locally, but radiosonde data were obtained from Bogota 200 km to the east, affording an estimate of conditions aloft. Sometimes it was possible to measure directly the depth of the overflow by using low-lift pilot balloons especially in the cloud-free air near Cali. The overflow in the pass could be estimated from the height of the accompanying cloud cap. The pass at El Saladito is about 700 m above the valley floor. In a typical dry-season day there is a 4C difference in potential temperature between the two air masses, and a typical overflow depth is 200 m. These values yield speeds of about 5 m sec^{-1} for the critical velocity at the top of the pass and about 16 m sec^{-1} near Cali. Gusty winds approaching these speeds are not uncommon during the late afternoon at Cali, indicating reasonable agreement with theory. The local buzzards are evidently familiar with these strong low-level afternoon winds, for they use them to advantage during their foraging cruises by rising on the westwardbound legs and flying lower on the return trip.

Katabatic wind speeds in this area are more sensitive to changes in the strength of the inversion than to changes in the depth of the overflow. For instance, if the potential temperature difference between the two air masses increases from 2 to 4 degrees, the valley wind speeds increase by 41 per cent, but if an overflow depth of 200 m were doubled the speed would increase only by 10 per cent. This helps to explain why, in general, katabatic wind speeds are greater during the dry season when temperature differences are greatest. This circumstance is unfortunate for the local foresters, for the resulting higher winds greatly increase the danger of forest fires over the slopes during the dry season. It is not unusual here to witness the unfamiliar sight of an afternoon grass fire working its way *down* a slope. It is mostly the great increase in wind speeds over the slopes during the afternoon that makes forest fires so prevalent at this time of the year.

The katabatic winds provide an added impulse for local convection which depends on the height of the jump that is generated when turbulent flow slows the winds below the critical velocity. This impulse is quite dependent on the depth of the overflow, and on the heating and mixing that may take place locally over the valley floor. Heating rates exceeding 50 per cent have been estimated from potential temperature measurements near the valley floor, but the real amount remains speculative, since no detailed local temperature measurements are available. It is interesting to compare reasonable estimates of the heights of the jump between the dry and the wet seasons. If a heating rate of 50 per cent is assumed for the dry-season example examined before, a jump of 1400 m above the valley floor is indicated. This is above the local condensation level, even for the dry season, and increased convection and cloud formation along the jump can be expected. During the wet season, the potential temperature differences between the air masses are smaller but the depths of the overflows at the passes are larger. Overflow depths of 400 m have often been measured during the wet season at El Saladito. Even if we assume that the inverson is reduced to half the dry season value, and that the heating rate is only 30 per cent in view of the greater amount of cloudy skies over the valley, a wind speed of 12 m sec^{-1} at the bottom of the El Saladito pass and a jump of nearly 2000 m somewhere east of Cali would be realized. Thus, although winds speeds during the rainy season should be smaller, the increased depth of the overflow (which is helped by increased cooling caused by heavier precipitation on the Pacific side), greatly increases the impulse provided by the jump.

6. Influence of the katabatic winds on the local climate

That mountain ridges facing a prevailing wind induce a rain shadow on the lee is a well-known fact. The Pacific coast of Colombia, for instance, averages well over 5000 mm of annual precipitation while the Cauca Valley, on the lee of the western Andes, averages only one-fourth as much. Rainfall within such sheltered valleys, however, is far from uniform.

Annual rainfall averages for the area studied are shown in Fig. 1. It is very apparent that precipitation is much less in the areas immediately east of low passes in the western Cordillera than on the opposite sides of the valleys, reversing on a local scale the general rain-shadow rule expressed above. Outstanding examples are Yumbo (620 mm) compared with downwind Miranda (1600 mm), Calima hydro (980 mm) and Darién (1800 mm) and Loboguerrero (700 mm) against Bitaco (1700 mm). Conversely, a station like Jamundí, sheltered by the high Farallones to the west, averages 50 per cent more rain than Cali 20 km to the north although there is practically no difference in elevation between the two. Stations at the east end of the Cauca Valley downwind from a low pass (at El Saladito), like Miranda, have much higher rainfall, for instance, than Amaime (1000 mm) 35 km to the north even though the latter is similarly situated against the central Cordillera. These rainfall differences could be explained in terms of the prevalence of downslope afternoon katabatic winds near the low passes which tend to reduce the local convective rains there and to concomitant hydraulic jump impulses downstream to the east of the valleys, which would tend to produce the opposite effect.

The ecological map of the Valle Department, now under compilation, also underscores a striking fact first noted by Espinal (personal communication, 1966): areas of low and high rainfall on the western and central Cordilleras, as deduced from primeval vegetation, are aligned in a herringbone fashion so that dry spots on the east slopes of the western Cordillera correspond to wet spots on the opposite side of the central Cordillera facing them. The dry areas on the east slopes of the western Cordillera are found in general along the low passes. Another factor noted in these ecological studies is the absence, in these low passes, of the so-called cloud-zone climate present elsewhere above 1500 m in which a high humidity prevails most of the year. It can be postulated that the cloud zone is missing there because the prevailing katabatic descending winds play the greatest role in reducing the relative humidity locally through adiabatic heating.

This drying effect of the descending winds, which would also result in greatly increased evapotranspiration, has a large influence on the local climate and hydrology. Where these winds are persistent, as happens at Loboguerrero, even an isolated steppe-like climate may result. Precipitation figures alone do not tell the whole story unless the low relative humidity and increased evapotranspiration are also taken into consideration. The increase in evapotranspiration, as shown by Crawford (1965), can also result in an important reduction of stream runoff. In some of the models used by Crawford, a 10 per cent increase in evapotranspiration may result in as much as 20 per cent decrease in runoff. The combination of decreased rainfall and increased evapotranspiration caused by prevailing strong afternoon katabatic winds could also help to explain the great variability observed in the yield of the various streams in this area.

7. Summary and conclusions

The regular occurrence of a tidal flow of relatively cool air from the Pacific produces local winds on the east slopes of the western Andes and nearby valleys that can be explained as a heated katabatic flow and related hydraulic jump. These phenomena have a marked influence on the local climate and ecology of the area.

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