Blocking in areas of complex topography, and its influence on rainfall distribution

Mimi Hughes *, Alex Hall, and Robert G. Fovell

UCLA, Los Angeles, California

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*Corresponding author address: Mimi Hughes, University of California, Los Angeles, Los Angeles, CA 90095.

E-mail: mhughes@atmos.ucla.edu

Abstract

Using a 6-km resolution regional climate simulation of Southern California, the effect of orographic blocking on the precipitation climatology is examined. To diagnose whether blocking occurs, precipitating hours are categorized by a bulk Froude number. The precipitation distribution becomes much more spatially homogeneous as Froude number decreases, and an inspection of winds confirms that this is due to increasing prevalence of orographic blocking. Simulated precipitation distributions are compared to those predicted by a simple linear model that includes only rainfall arising from direct forced topographic ascent. The agreement is nearly perfect for high Froude cases but degrades dramatically as the index decreases; as blocking becomes more prevalent, the precipitation/slope relationship becomes continuously weaker than that predicted by the linear model. We therefore surmise the linear model would be significantly improved during low Froude hours by the addition of a term to reduce the effective slope of the topography. Low Froude, blocked cases account for a large fraction of climatological precipitation, particularly at the coastline where more than half is attributable to blocked cases. Thus the climatological precipitation/slope relationship seen in observations and in the simulation is a hybrid of blocked and unblocked cases. These results suggest orographic blocking may substantially affect climatological precipitation distributions in similarly configured coastal areas.

1. Introduction

Southern California receives almost all precipitation from synoptic-scale disturbances between October and April; during summer months precipitation is almost entirely suppressed by the North Pacific high pressure. These systems move inland from the Pacific Ocean, so that nearly all precipitation falls when the wind direction is generally onshore (i.e. from the west/southwest, *Conil and Hall* 2006). Thus when it rains in Southern California, winds blow roughly parallel to the elevation gradient on the ocean side of the coastal mountains, resulting in a precipitation increase with elevation on the ocean side, and a well-defined rain shadow in the interior (viz., the Mojave Desert). These orographic influences are clearly visible in the observed climatological precipitation distribution (Fig. 1a).

The simplest way topography modifies the precipitation distribution is through forced ascent: as air moves over topography it is forced to rise parallel to the slope, causing moisture to condense and fall out. This effect is captured by the upslope model discussed in *Smith* (1979). The upslope model predicts precipitation increasing linearly with slope, with a constant of proportionality given by the product of the wind component parallel to the elevation gradient and the column-integrated moisture. Evidence for this simple relationship can be seen in the observed precipitation in Southern California as a function of elevation gradient magnitude in the coastal zone (Fig. 1c). There is a large increase in rainfall amount as slope increases. *Smith* (2003) and *Smith and Barstad* (2004) extend the upslope model by including the effects of more realistic microphysics and linear wave dynamics: Condensation and fallout of cloud droplets occur over a finite time interval during which the droplets are advected, shifting precipitation downstream with respect to that predicted by the upslope model; flow over a mountain also causes gravity waves that tilt upstream with height, creating lift ahead of the topograph-

ical barrier and shifting precipitation upstream. *Smith et al.* (2005) showed that this refined upslope model, hereafter called the linear model (LM), can roughly reproduce the precipitation of coastal Oregon. Because the LM is a state-of-the-art conceptual model for understanding orographic effects on precipitation, we use it as a tool to examine the mechanisms controlling Southern California's precipitation distribution. To the extent the LM accurately captures the precipitation distribution, the mechanisms controlling the distribution are probably limited to those contained in the LM. On the other hand, inaccuracies in the LM's predictions imply other mechanisms are important.

A potential shortcoming of the LM is that it does not capture nonlinear dynamical mechanisms such as upstream blocking (*Pierrehumbert and Wyman* 1985). When the kinetic energy of the flow approaching a mountain is less than the energy necessary to surmount it in a stably stratified environment, the flow decelerates well ahead of the topographic barrier (*Smolarkiewicz and Rotunno* 1990; *Pierrehumbert and Wyman* 1985; *Baines* 1987; *Sheppard* 1956), causing a region of 'blocked' flow that enhances precipitation upwind of the mountain range (e.g. *Jiang* 2003; *Neiman et al.* 2004; *Medina and Houze Jr* 2003). Using a numerical model with idealized topography, *Jiang* (2003) characterized the difference in circulation and orographically forced precipitation during blocked and unblocked cases. In these simulations, unblocked cases have precipitation increasing with elevation, with a maximum close to the apex of the topography. In blocked cases, precipitation falls in a broad band upstream of the mountain peak, with some still falling over the mountain, similar to what we see in Fig. 3b. The difference in precipitation distributions is due to a change in the response of the approaching flow to the topography: In the unblocked case, the approaching air has enough kinetic energy to surmount the approaching mountains; thus the air flows directly over the topography and is mechanically lifted by the

orography itself. In the blocked case, the flow is deflected and stagnates ahead of the mountain, causing convergence well ahead of the elevation increase. The pile-up of air acts as a barrier, forcing the air to flow up and over it, creating a region of ascent upwind of the topography.

Although most studies of orographic precipitation acknowledge potential impacts of blocking—indeed, even *Smith and Barstad* (2004) point it out as a possible limitation of the LM—very few attempts have been made to quantify the climatological frequency of blocking or its influence on the climatological distribution of precipitation in a mountainous region. In Southern California, precipitation over the open ocean is roughly half that at the nearby coastal stations with near zero slope (Fig. 1c). Although part of the increase over land could be explained by frictional convergence, this near-doubling suggests orographic blocking may enhance precipitation falling at small-elevation-gradient, coastal locations, a fact also pointed out in the case studies of *Neiman et al.* (2004); *Rotunno and Ferretti* (2001) and *Grossman and Durran* (1984).

Blocking therefore has the potential to substantially alter the simple precipitation/slope relationship encapsulated by the upslope model. In this study, we find precipitation distributions are substantially different when blocking events occur. Moreover, these events occur frequently enough that they account for a large fraction of precipitation, particularly in low-slope coastal areas, and their signature can be clearly seen in the climatological precipitation distribution. We find that the LM can reproduce the spatial pattern of precipitation when blocking is weak, but that the model's exclusion of blocking effects is the primary factor preventing it from producing an accurate description of precipitation distributions at other times. As blocking becomes more prevalent, the precipitation/slope relationship becomes continuously weaker than that predicted by the LM, and the observed precipitation/slope relationship of Fig. 1c is a hybrid of blocked and unblocked cases.

2. Tools

It is difficult to investigate the impact of topography on precipitation in Southern California using observations alone because the observations are so sparse, particularly at high elevations, and we cannot determine how representative they are of the region's unsampled areas. In addition, the observational data contain only a few surface variables, preventing us from rigorously investigating mechanisms regulating precipitation distribution. Reanalysis products offer a full suite of 3-D variables, but are likely too coarse in resolution; for example, the North American Regional Reanalysis (*Mesinger et al.* 2006), with 32-km grid spacing, resolves only the grossest terrain features in Southern California, and would barely resolve differences between the coast and the coastal mountains. Here we circumvent these problems by relying on a multi-year 6-km resolution simulation of the southern third of California, described below. In this section we also describe the LM introduced in section one in more detail.

a. MM5 simulation

1) MODEL CONFIGURATION

This simulation was created with the Penn State/NCAR mesoscale model version 5, release 3.6.0 (MM5, *Grell et al.* 1994). The 6-km domain was nested within an 18-km domain covering southern California and parts of Arizona, Nevada, and Mexico, which was likewise nested within a 54-km domain encompassing most of the western US. At 6-km resolution, all major mountain complexes in Southern California were represented, as were the Channel Islands just off the coast (e.g., Fig. 1b, thick black lines show coastline). Each domain had 23 vertical levels, with the vertical grid stretched to place the highest resolution in the lower troposphere.

In the outer two domains, the Kain-Fritsch 2 (*Kain* 2002) cumulus parameterization scheme was used. In the 6-km domain, only explicitly resolved convection could occur. In all domains, we used the MRF boundary layer scheme (*Hong and Pan* 1996), Dudhia simple ice microphysics (*Dudhia* 1989), and a radiation scheme simulating longwave and shortwave interactions with clear-air and cloud (*Dudhia* 1989).

The boundary conditions came from the National Center for Environmental Prediction's 40-km resolution Eta model analysis data from the National Center for Atmospheric Research's GCIP archive. The time period covered was from May 1995 to April 2006. Throughout this period, MM5 was initialized every 3 days at 18 UTC (10 a.m. local time) and run for 78 hours, with the first six hours being discarded as model spin-up. The interior boundary conditions and sea-surface temperatures were updated at each initialization, with the lateral boundary conditions updated continuously throughout the run. Thus the simulation in the 6-km nest acts as a reconstruction of the local atmospheric conditions based on known large scale atmospheric conditions. More specific details about the land surface and topography used in the simulation can be found in *Hughes et al.* (2007).

2) MODEL VALIDATION

Figure 1b shows the climatological precipitation distribution for the MM5 simulation, for comparison with observations (Fig. 1a). Visual inspection of the two panels shows qualitative agreement: In both cases, most low-lying coastal locations have precipitation rates between 2 and 4 cm month⁻¹, with values generally increasing with elevation on the coastal side of the mountains. Both datasets show close to zero precipitation inland of the coastal mountain ranges. This tight correspondence between observed and simulated climatological precipitation is even

more clear in a point-by-point comparison (Fig. 2, correlation coefficient = 0.87). If the observed data are linearly regressed onto MM5 values, the slope is 1.13 and the y-intercept is 0.39 (gray line, Fig. 2). This indicates MM5 is somewhat underpredicting precipitation throughout the region, and that the error is slightly larger for stations with larger climatological precipitation (i.e., locations at higher elevation). No corrections were applied to MM5 precipitation to account for differences in elevation between model gridpoints and station locations, nor did we attempt to account for features unresolved by the 6-km grid spacing. However, the high degree of correspondence between the two datasets indicates these sub-grid-scale features are unimportant, and so we are confident in the simulation’s ability to represent the features of the region’s precipitation with a reasonable degree of accuracy.

Because precipitation is so closely tied to atmospheric circulation, the accurate precipitation climatology in the simulation suggests it also correctly represents local circulation, a fact verified by two previous studies (*Conil and Hall 2006; Hughes et al. 2007*). For example, correlating this simulation’s winds with available observations, *Conil and Hall (2006)* confirmed the simulation captures synoptic time-scale variability in daily-mean wind. Thus we have confidence that the simulation accurately captures not only precipitation, but also the dynamics underlying its spatial distribution.

b. Linear Model for orographic precipitation

Here we briefly describe the linear model (LM) used for comparison with simulation output in section four. The LM was developed by *Smith and Barstad (2004)* to describe the pattern of precipitation arising from forced ascent of moist air over topography. It assumes forced ascent converts moist air to cloud water, which then converts to hydrometeors with timescale τ_c and

fallout with timescale τ_f (Smith 2003). The hydrometeors and cloud water are advected by a mean wind $\mathbf{U} = U\mathbf{i} + V\mathbf{j}$. The LM also includes a gravity wave term allowing the vertical velocity to change with height; this term is dependent on the vertical wavenumber, m . Thus in the LM, precipitation broadly scales with the gradient of the terrain modified by advection processes. For convenience the LM operates in Fourier space, where representations of physical processes can be combined into a single transfer function. When this transfer function is multiplied with the Fourier transform of the topography, $h(k, l)$, it produces a Fourier transform of precipitation, $P(k, l)$. The equation for $P(k, l)$ is:

$$P(k, l) = \frac{C_w i \sigma h(k, l)}{(1 - imH_w)(1 + i\sigma\tau_c)(1 + i\sigma\tau_f)}, \quad (1)$$

where H_w is a scale height for water vapor:

$$H_w = \frac{R_v T_{ref}^2}{L\gamma}, \quad (2)$$

C_w relates moisture and vertical motion:

$$C_w = \frac{\rho_{s,ref} \Gamma_m}{\gamma}, \quad (3)$$

σ is the intrinsic frequency: $\sigma = Uk + Vl$, and m is the vertical wavenumber:

$$m(k, l) = [(\frac{N_m^2 - \sigma^2}{\sigma^2 - f^2})(k^2 + l^2)]^{1/2}. \quad (4)$$

In the above equations, $R_v = 461 \text{ J kg}^{-1} \text{ K}^{-1}$ is the gas constant for vapor, $L = 2.5 * 10^6 \text{ J kg}^{-1}$ is the latent heat of vaporization, f is the Coriolis parameter ($f = 8 * 10^{-5}$ at 34N), γ is the environmental lapse rate, Γ_m is the moist adiabatic lapse rate, N_m^2 is the moist Brunt-Väisälä frequency (Durrant and Klemp 1982):

$$N_m^2 = \frac{g}{T} (\frac{dT}{dz} + \Gamma_m) (1 + \frac{Lq}{R_d T}) - \frac{g}{1 + q} \frac{dq}{dz}, \quad (5)$$

$R_d = 287 \text{ J kg}^{-1} \text{ K}^{-1}$ is the ideal gas constant for dry air, q is the water vapor mixing ratio and $\rho_{s,ref}$ and T_{ref} are saturation vapor pressure density and temperature at a reference height. A background precipitation rate, P_{back} , representing the precipitation falling at zero elevation far from topographical influence, is then added to the inverse Fourier transform, $P(x, y)$. The solution is then truncated to eliminate negative values.

3. Categorization of events by Froude number

For analysis and categorization of precipitation events, we devised a simple index corresponding to the degree to which the flow is blocked. We categorize raining hours (i.e., hours when domain average precipitation is larger than 0.1 cm hr^{-1}) of the MM5 simulation by the value of a bulk Froude number (Fr^2), the ratio of the kinetic energy of the ambient flow to the energy required to surmount the coastal mountains:

$$Fr^2 = U^2 / (N^2 H^2), \quad (6)$$

where U is the cross-mountain wind speed, N is the Brunt-Väisälä frequency, and H is the mountain height. Most generally, Fr^2 is defined using the dry Brunt-Väisälä frequency: $N_d^2 = \frac{g}{\theta_v} \frac{d\theta_v}{dz}$, where θ_v is virtual potential temperature; however, when the air impinging on the mountains is saturated, many studies (e.g., *Durran and Klemp* 1982) have shown that a more appropriate choice of stability parameter is the moist Brunt-Väisälä frequency, N_m^2 (equation 5). Because N_m is only applicable in saturated conditions, we use N_m when the coastal-average near-surface relative humidity is greater than 90%, and otherwise we use N_d . The results are insensitive to moderate variations of the relative humidity threshold (i.e. between 85% and 95%). We then compute the mean Brunt-Väisälä frequency in the coastal zone for our calculation of

Fr^2 . For the wind speed in equation 6 we use the mean open-ocean wind speed (i.e., averaged over the area from 32.75N to 33.3N and 119.5W to 120.75W).

To complete equation 6, we need to know H . If we were calculating Fr^2 for a single mountain, as is common practice, this would be straightforward; but no unique mountain height exists for the entire coastal zone. However, within this region the heights of the main mountain complexes are comparable, and each may have a similar impact on the flow. Therefore using one value for H for the entire region is reasonable. We use barrier height $H = 1$ km, slightly larger than the mean elevation in the coastal zone. Furthermore, spatial variations of both the Brunt-Väisälä frequency and the unperturbed winds are temporally correlated throughout our small domain. In this way, our bulk Fr^2 provides an estimate of the relative likelihood of blocking given various large-scale flow configurations; for example, if the bulk Fr^2 is small, the likelihood of blocking is larger for every mountain complex, even if the absolute likelihood of blocking is not the same. As we show in section five, blocked flows are systematically more likely throughout the region as bulk Fr^2 decreases. In the following section, we will examine low Fr^2 hours ($0 < Fr^2 < 0.4$) and high Fr^2 hours ($Fr^2 > 10$ or $Fr^2 < 0$). Note that we categorized hours in which the coastal zone is on average conditionally unstable ($N_m^2 < 0$ and thus $Fr^2 < 0$) as high Fr^2 , because during those hours blocking cannot occur. Each group accounts for approximately 20% of the total raining hours.

4. Effect of blocking on climatology

Fig 3 shows the normalized average precipitation for the high Fr^2 (panel a) and low Fr^2 (panel b) cases. Here we have normalized by the domain-maximum precipitation to focus on the

distribution differences between the two composites. During high Fr^2 hours, precipitation over the ocean and in the desert interior is relatively small, and rainfall increases with elevation, with the largest values on the coastal side of the mountains. Compared with the high Fr^2 average, the low Fr^2 average shows slightly more precipitation over the open ocean and in the desert interior. However, the largest difference between the two composites is within the coastal zone; the low Fr^2 distribution shows a far more complex relationship between precipitation and topography, and rainfall is much more uniformly distributed with elevation. If we examine the relationship between precipitation rate and slope in the raw data for each of the two cases (Fig. 3c), we see that the low Fr^2 precipitation is nearly independent of slope, while its high Fr^2 counterpart is strongly constrained by it: The ratio of precipitation at the steepest slopes (above 80 m km^{-1}) versus the gentlest slopes (below 10 m km^{-1}) is nearly four to one for the high Fr^2 cases, and close to one to one for the low Fr^2 cases.

Orographic blocking very likely explains the difference in spatial distribution of precipitation between low and high Fr^2 averages. The average near surface winds for the high Fr^2 (Fig. 4a) and low (Fig. 4b) Fr^2 cases reveal that the characteristics of blocking found by *Jiang* (2003); *Smolarkiewicz and Rotunno* (1990); *Pierrehumbert and Wyman* (1985), and others, are more prevalent in the simulation's low Fr^2 cases. As with precipitation distribution, we have normalized the wind values to emphasize the distribution difference. Both low and high Fr^2 cases have strong, roughly onshore flow over the ocean that slightly decelerates when it reaches land. However between the mountains and the coast, the contrast between the high and low Fr^2 cases is quite apparent: The high Fr^2 winds show slight deceleration through this area and little change in direction, whereas the low Fr^2 winds turn sharply to the left as they approach the coast. This indicates the surface flow is on average blocked from the base of the mountains

to the coastline during low Fr^2 hours. Fig 4 also reveals localized areas with no clear cut-off between low and high Fr^2 ; a few areas upwind of the sharpest elevation gradients exhibit characteristics of blocking (e.g. nearly zero wind speeds south of the San Gabriel Mountains) in the high Fr^2 average (Fig. 4a). This is most likely due to our defining Fr^2 in a bulk sense for the entire region, despite the complexity of the topography.

We have established that blocking modifies precipitation distribution during composites of raining hours: How is this manifested in the climatological distribution of precipitation? High elevations receive copious precipitation during unblocked cases, (Fig. 3a), so that unblocked cases, which are only about 20% of the total hours, account for over 30% of the total precipitation at these locations (Fig. 5a). However, over the coastline and nearby ocean, more precipitation falls during the low Fr^2 cases, resulting in blocked cases substantially contributing to the total precipitation at these locations (20 to 30%, Fig. 5b). If the percentage of total precipitation is calculated for the remaining intermediate quintiles of hours binned by bulk Fr^2 , the percentages at high (low) elevations steadily increase (decrease) as Fr^2 increases (not shown). This explains the large discrepancy between climatological precipitation in low elevation areas and the precipitation over the adjacent open ocean seen in the observations (Fig. 1c): Blocking causes precipitation to extend well upstream of the topography, and this occurs during a large fraction of precipitating hours in Southern California, substantially enhancing precipitation at low coastal elevations.

5. Mechanisms underlying precipitation distribution

Up to this point we have shown that orographic blocking occurs frequently in Southern California, and that it substantially influences the geographic distribution of rainfall by enhancing rainfall at low elevations. Here we show that classifying hours by bulk Fr^2 not only allows us to distinguish between blocked and unblocked hours, but also gives us a quantitative estimate of the extent to which blocking is affecting the geographic distribution of rainfall. To do this, we utilize the LM described in section 2b for composites of the rainy hours binned by Fr^2 . Fig. 6a shows the LM result for the conditionally unstable hours, for comparison with the MM5 composite (Fig. 6b). During these hours, the LM accurately reproduces the spatial pattern and magnitude of precipitation throughout the coastal zone, a fact emphasized by a high spatial correlation (0.83). In contrast, the LM result for the hours with lowest Fr^2 (Fig. 6c) bears little resemblance to the MM5 result (Fig. 6d), because it drastically overestimates the precipitation at high elevations and underestimates it at the coastline, and this results in a low spatial correlation in the coastal zone (0.36). To quantify the effect of blocking on precipitation distribution for all raining hours, we calculate the spatial correlation between the LM solution and the MM5 composite for all the hours binned by the value of the bulk Fr^2 (Fig. 6e). We see that as Fr^2 increases, the spatial distribution of precipitation becomes more and more consistent with that of the LM, indicating blocking is less and less likely to occur and influence precipitation distribution. Thus the magnitude of the bulk Fr^2 is an excellent predictor of the extent to which blocking reduces the accuracy of the LM.

These results imply that exclusion of blocking effects is the main shortcoming of the LM. An examination of the relationship between precipitation rate and elevation-gradient magnitude for composites of MM5 precipitation binned by Fr^2 (Fig. 7) elucidates the effect blocking has

on the precipitation/slope relationship, clarifying the reason for the LM's poor performance at low Fr^2 . When Fr^2 is very large or negative, the precipitation distribution increases strongly with terrain slope, and the LM can reproduce the precipitation distribution because it is tightly constrained by the elevation gradient. In contrast, when Fr^2 is low, blocking effects redistribute precipitation, causing relatively less to fall at large slopes and more at small slopes, effectively changing the 'topography' that the approaching air encounters. Our results further imply that the LM would be quite accurate for all cases if a term based on bulk Fr^2 were added to account for the effects of orographic blocking, and Fig. 7 suggests that this term should act to reduce the windward 'slope' of the topography in the LM as Fr^2 decreases.

One reason coarse resolution atmospheric models cannot accurately reproduce precipitation distributions is that they do not adequately resolve topography, and so cannot capture its effect on precipitation. Our results imply that the large-scale Fr^2 is a primary modulator of the relationship between slope and rainfall amount in areas of complex topography. Statistical downscaling techniques could employ this information to get a better representation of rainfall distribution for differing ambient conditions. Further, this relationship could also improve statistical interpolation techniques, such as Parameter-Elevation Regressions on Independent Slopes Model (PRISM) (Daly *et al.* 1994).

Southern California's Mediterranean climate and the fact that all its precipitation falls when the ambient flow is in the onshore direction allowed us to use a very simple definition of Fr^2 that does not include any information about the direction of the flow. Furthermore, the close proximity of stable marine air to the coastal mountains may increase the frequency of low bulk Fr^2 cases in this region, increasing the frequency of orographic blocking and amplifying its effect on precipitation distribution. Nevertheless, our conclusions would probably be quite

similar in many other regions where precipitation is mainly non-convective, particularly those that, like California, have complex topography next to a large region of moist but stable air such as the subtropical oceans. For instance, we expect blocking to play a large role in determining the distribution of precipitation along much of the west coast of North America, as well as the middle portion of the South American west coast.

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FIG. 2. Scatterplot of climatological precipitation, observed versus MM5 at nearest gridpoint.

Gray line shows linear regression line.

FIG. 3. Average precipitation rates, normalized by the domain maximum precipitation rate ($Max P$), for (a) hours with $Fr^2 > 10$ or $Fr^2 < 0$ (b) hours with $0 < Fr^2 < 0.4$. Gray dotted line in (a) shows grid points used in (c). Terrain is shown as in Fig. 1. (c) Average precipitation rates for coastal land points versus elevation gradient for hours with $0 < Fr^2 < 0.4$ (light blue dots) and hours with $Fr^2 > 10$ or $Fr^2 < 0$ (darker blue circles). Light and dark blue lines show corresponding regression lines.

FIG. 4. Average 10-m wind values for (a) hours with $Fr^2 > 10$ or $Fr^2 < 0$ (b) hours with $0 < Fr^2 < 0.4$. Winds are normalized by the average open ocean wind speed, which is 6.17 m s^{-1} and 2.09 m s^{-1} for panel a and b, respectively. Terrain is shown as in figure 1.

FIG. 5. Percentage of total precipitation that falls during hours with (a) $Fr^2 > 10$ or $Fr^2 < 0$ and (b) hours with $0 < Fr^2 < 0.4$. Terrain is shown as in figure 1.

FIG. 6. (a, c) LM and (b, d) MM5 precipitation for hours with (a, b) $Fr^2 < 0$ and (c, d) $0 < Fr^2 < 0.2$. (e) Spatial correlation coefficient of LM and MM5 precipitation pattern within the coastal land areas (gray dotted line, Fig. 1a) for cases binned by Fr^2 . Bin centers are shown on the abscissa, with bin ranges shown in Fig. 7. All parameters necessary for the LM were calculated from the average surface temperature, lapse rate, and wind in the coastal zone (Fig. 1a, gray dotted line) for each group of hours. τ_c and τ_f are fixed at 600 s, and P_{back} is the average open-ocean value of precipitation.

FIG. 7. Regression lines of average precipitation rates for coastal land points versus elevation gradient for cases binned by Fr^2 . Bin ranges are labeled, and are the same as in Fig. 6; bin ranges were chosen to have approximately the same number of cases per bin.