OROGRAPHIC PRECIPITATION

Precipitation that has been generated or modified by topography, typically through the forcing of vertical atmospheric motions.

Introduction

The influence of mountains upon rain and snowfall is often profound, creating some of the Earth's wettest places (e.g. Cherrapunji in India, where monsoon flow encounters the southern Himalayas, has received 26.5 m in one year) and driest places (e.g. The central valleys of the Atacama desert, shielded by surrounding mountains, can go for decades without rainfall). Orographic effects on precipitation are also responsible for some of the planet's sharpest climatic transitions. The classic example is the so-called 'rain shadow'; for a mountain range oriented perpendicular to the prevailing winds, precipitation is greatly enhanced on the windward side and suppressed in the lee. However, the full gamut of orographic influences is much broader than this: precipitation can be enhanced in the lee, over the crest, or well upwind of a mountain.

The mass balance of Earth's snow and ice is strongly affected by orographic precipitation. Accumulation of mountain snowpack and on alpine glaciers is typically dominated by orographic snowfall. The Greenland and Antarctic Ice sheets are themselves substantial topographic features responsible for orographic effects on precipitation. Avalanches are sensitive to the detailed stratigraphy of the snowpack, which is partly determined by the sequence of orographic graupel, rain, ice, and various snow crystals that fall during storms.

This article will focus on the important physical processes controlling orographic precipitation, and the observational and modeling techniques that have been used to characterize and understand it. More complete reviews may be found in: Smith (1979); Roe (2005); and Smith (2006).

Fundamentals

Orographic precipitation is shaped by myriad non-linear processes operating on scales ranging from the 1000 km size of storms and major mountains to the sub-micron size of cloud droplets. Still, the most fundamental of these processes are thermodynamic in nature and are well understood. Almost all orographic influences on precipitation occur due to rising and descending atmospheric motions forced by topography. These motions can be forced mechanically, as air impinging on a mountain is lifted over it, or thermally, as heated mountain slopes trigger buoyancy-driven circulations. Rising motion causes the air to expand and cool, which is important since the amount of water that may exist as vapor in air is an approximately exponential function of temperature (described by the Clausius Clayperon equation). Thus if cooling is sufficient, air saturates and the water vapor condenses into cloud droplets or forms cloud ice crystals. These droplets and crystals grow by various processes until they become large enough to fall as rain and snow. It is important to emphasize that moist ascent over topography alone is typically insufficient to generate precipitation: these orographic effects mainly modify precipitation during preexisting storms (e.g. Browning et al., 1974; Smith, 2006). Conversely, when air descends it warms and dries, and both cloud and precipitation evaporate.

A useful tool for understanding some of the basic controls on orographic precipitation is the "upslope" model (e.g. Smith, 1979; Smith, 2006). This idealized and physically-based model predicts the water condensed when flow with given surface specific humidity (q_v, expressed as a mixing ratio), density (ρ), and uniform wind velocity ($\overline{U}$), impinges upon topography (with height: $h(x,y)$). The model assumes saturated air, an idealized temperature profile, and flow that parallels the topography at all heights. Under these assumptions the vertically-integrated source of condensed water per unit time is:

$$S(x,y) = \rho q_v \overline{U} \cdot \nabla h(x,y) \qquad \text{(Eq. 1).}$$

This is also the precipitation rate at the surface if it is further assumed that conversion of cloud condensate to precipitation and fallout of precipitation are instantaneous. This model reveals some key controlling parameters: the moisture flux ($\rho\, q_v\, \overline{U}$), which determines the vapor available for condensation, and the topographic slope (∇h) in the direction of the airflow, which determines the rate of the forced vertical motion.

Airflow Dynamics

Actual flow over topography during precipitation is seldom as simple as that assumed in the upslope model. Atmospheric density and temperature stratification strongly control the flow, since the typically stable stratification of the atmosphere means that a parcel of air displaced upwards becomes negatively buoyant (since it is cooler and denser than it surroundings) and is pulled back downwards. The strength of this effect may be quantified by the Brunt-Vaisala buoyancy frequency:

$$N^2 = \frac{g}{T}(\gamma - \Gamma) \qquad \text{(Eq. 2),}$$

with γ representing the observed atmospheric lapse rate (i.e. the rate of decrease of temperature with height), and Γ the theoretical dry adiabatic lapse rate for a rising air parcel (-9.8 K km^{-1}). When stratification is stable (N^2>0) the buoyancy restoring force causes airflow over mountains to take the form of waves, which oscillate with frequency N. These wave motions cause orographic uplift to be displaced upstream by varying degrees (enhancing precipitation ahead of the mountain) or decay with height (limiting precipitation enhancement) depending on the strength of the incoming flow and stratification relative to the mountain width (Smith, 1979; Smith, 2004).

Under conditions where N^2 is large and positive, the incoming flow is weak, and/or the mountain is high, the effects of stratification can be overwhelming and the low-level flow may be unable to surmount the mountain. In such cases the flow is said to be blocked, and air is forced to deflect around the mountain, stagnate, or even reverse (e.g. Smith 1979; Marwitz, 1981). This can result in orographic enhancement that is limited or forced to occur further upstream (e.g. Houze and Rotunno, 2008). A parameter useful for predicting the onset of blocking is the non-dimensional mountain height:

$$M = \frac{hN}{U} \qquad \text{(Eq. 3),}$$

where U is the incoming flow speed and h is the mountain height. When M exceeds unity blocking is favored.

When the atmosphere is unstably stratified (N^2< 0) convective overturning motions may be triggered. Convective cells embedded within a storm are a common feature in orographic precipitation (e.g. Browning; 1974). When the atmosphere is strongly unstable orographic thunderstorms may be triggered.

As air rises and moisture condenses latent heat is released. This heating effectively reduces the stratification. As a result, many flows that would be blocked are able to flow over mountains when condensation occurs, leading to important impacts precipitation distributions (e.g. Jiang; 2003).

Microphysics

Conversion of water from vapor to cloud to precipitation is a substantial task. Typical cloud particles must grow about 1 billion-fold in volume before they are large enough to fall as precipitation. The evolution of cloud and precipitation particles occurs on scales from millimeter to micron, earning it the term cloud microphysics. Clouds droplets initiate and grow on fine particulates known as cloud condensation nuclei or ice nuclei, the concentration of which determines the size and number of cloud drops. The growth of cloud particles to the size of rain and snow occurs by the diffusion of vapor onto cloud particles and by the collision, coalescence, and aggregation of droplets and crystals. These growth processes depend on temperature, humidity, and the character of the airflow.

Hobbs et al. (1975) demonstrated the importance of microphysical processes by simulating a winter orographic storm with varying initial droplet concentrations. For small concentrations of cloud ice, snow grew quickly and fell to the ground over the windward slopes, whereas for large concentrations growth was slow and snow was blown nearly 100 km into the lee. Precipitation phase is also crucial, as snow falls much more slowly than rain (roughly 0.5-2 m s^{-1} vs. 7-10 m s^{-1}), and snow rimed with super-cooled water (graupel) falls at intermediate speeds. These fall speeds determine how far downwind precipitation drifts as it falls.

The Melting-Level

Knowing the melting level --the elevation at which snow turns to rain as it falls-- is important for studies of mountain snow and glaciers. Characterizing the melting level over topography is not as simple as just measuring it upstream since orographic effects may modify melting levels during storms. The lifting of air on the upwind side of the mountain leads to expansion and cooling, while the phase change of orographic snow falling through the melting level causes latent cooling, both of which are responsible for observed depression of the melting level over windward slopes that can amount to 0.5 km or more (e.g. Marwitz, 1981). The thermal inertia of the land, cold air trapped in valleys, and the degree of turbulent mixing near the surface may all influence the melting level as well.

Observations

The most direct observations characterizing orographic precipitation come from rain and snow gauges that measure accumulation at mountain sites. For example, the Cascade and Olympic Mountains (Figure 1(a)), located in the northwestern United States, receive plentiful orographic rain and snow from the mid-latitude cyclones of the Pacific storm track. Rain and snowfall varies greatly over the Cascades, but fully characterizing these variations with gauges alone is quite challenging due to the paucity of observations located within the mountains away from the populated valleys and foothills (Figure 1(a)). The high concentration of gauges needed to characterize orographic precipitation is highlighted by observations from a dense gauge network in the southwestern Olympics (Figure 1(a), Minder et al., 2008). These observations show large differences in annual mean precipitation over scales of a few kilometers, maximizing on ridge-tops (Figure 1(e)). This pattern of precipitation is distinct from the rain shadow predicted by the upslope model. It arises because precipitation from aloft falls through low-level orographic clouds and grows by colliding with and collecting cloud droplets, in what is termed the “seeder-feeder” mechanism (Bergeron, 1969).

Statistical techniques can fill in gaps in observational networks. For example, the Parameter Regression on Independent Slopes Method (PRISM; Daly et. al., 1994) uses localized regressions of elevation and precipitation to interpolate between observations. PRISM output is shown for the Cascades and Olympics in Figure 1(c). Other gridded gauge analyses from the well-instrumented European Alps (Frei and Schär, 1998) reveal more complex large-scale patterns than shown in Figure 1(c). The Alps receive storms arriving from a much wider range of directions, erasing any simple rain shadow and producing precipitation maxima on both sides of the range.

Remote sensing offers an alternative method for studying orographic precipitation. Satellite methods are particularly useful for remote, poorly instrumented regions. For example, the Tropical Rainfall Measuring Mission (TRMM) satellite operates by emitting pulses of microwave radiation, which are reflected by precipitation. Data from TRMM have been used to characterize the pattern of precipitation over the Himalayas at 10 km scales (Anders et al., 2006), revealing a broad double-band of maximum precipitation along the southern slopes and local enhancements within windward valleys relative to the 4 km-high flanking ridges where the moisture content is quite low (Anders et al., 2006). Additional remotely-sensed data come from ground-based radars, a great number of which are deployed for weather forecasting. These can be used to make detailed observations of precipitation, including precipitation phase, with high spatial and temporal resolution. In a classic study, Browning et al. (1974) used radar over the coastal hills of Wales to show that intense periods of mountain precipitation occur when rainfall cells from upwind of the mountains are advected over the mountains and enhanced as instability is released and the seeder-feeder mechanism acts. Unfortunately, radar can be challenging to use in mountainous terrain where the beam is often blocked by topography.

Both *in situ* and remote observations from aircraft have been a central component of several field projects devoted to better understanding orographic precipitation. The most expansive of these efforts to date, the Mesoscale Alpine Programe (MAP), focused on the southern slopes of the European Alps. Results from MAP revealed "that detailed knowledge of the orographically-modified flow is crucial for predicting the intensity, location, and duration of orographic precipitation" (Houze and Rotunno, 2007, p.811), and that this flow is a strong function of the low level stability. Furthermore, under different flow regimes contrasting microphysical growth mechanisms become important, influencing the enhancement and distribution of precipitation (Houze and Rotunno, 2007).

Models

A vast array of models, each with their own advantages and drawbacks, have been used to characterize and understand orographic precipitation. The most basic of these are statistical in nature, relying upon empirical relations to estimate precipitation as is done for PRISM. Such models can be quite quantitatively successful, but need adequate data for calibration and can fail dramatically when observations are sparse or when anomalous atmospheric conditions occur.

The upslope model, described above, is an example of a class of simple physically-based models that rely upon a series of idealizing assumptions to estimate precipitation with only minimal information about the incoming flow. Such models can illuminate fundamental processes and make ballpark estimates of precipitation, but neglecting airflow dynamics and cloud microphysics severely limits their physical realism.

Another class models are intermediate in complexity, maintaining simplicity while incorporating more governing physics than the upslope model. An example of this is the linear theory model put forth by Smith and Barstad (2004), which builds on the upslope model to include linearized mountain wave airflow dynamics, microphysical conversion and fallout timescales, and lee side evaporation of precipitation. Such models are useful for the same reasons as the upslope model, but offer a much more complete physical representation and better performance. Still, these models neglect important non-linear processes such as airflow blocking and microphysical collection and must be calibrated to perform well. An application of Smith and Barstad (2004)'s model to the Cascades and Olympics is shown in Figure 1(d).

Mesoscale numerical weather prediction models are the most sophisticated modeling tool used in the study of orographic precipitation. They solve the full time-dependent equations of atmospheric motion and thermodynamics numerically on a three dimensional grid and use schemes that simulate the interactions occurring on the microphysical scale between vapor, clouds, and precipitation. These models are capable of realistically representing transient interactions between large-scale storms and mountains, and non-linear effects of blocking and microphysics. Yet, this physical realism comes at a computational cost, and these models can take substantial time to run even on fast computers with parallelization. Precipitation from the MM5 mesoscale model, used for operational weather forecasting, is shown in Figure 1(c) over the Olympics and Cascades. For some regions the model performance is excellent even on small scales, as shown in Figure 1 (e). However, these models cannot be taken for truth as they can be configured in a multitude of ways that give differing results. Even the best models can still have major errors, for individual storms and climatological averages, due to the challenges of simulating microphysical processes as well as inherent limits that exist on atmospheric predictability.

Climate change and variability

The sensitivity of orographic precipitation to large-scale climate variability and climate change is an active area of research. It is well known that year-to-year variations in mountain rain and snowfall for ranges such as the Cascades are largely due to variations in the intensity and location of the mid-latitude storminess, with some of those variations related to large-scale patterns of climate variability such as the El Nino Southern Oscillation. Understanding how orographic precipitation will be altered due to anthropogenic climate change requires understanding the temperature sensitivity of orographic precipitation processes, as well as knowledge of how storm tracks and large-scale circulation will change.

The temperature dependence of orographic precipitation was investigated in depth by Kirshbaum and Smith (2008) using a mesoscale model. They found that while precipitation increases with the temperature and humidity of the atmosphere, these increases are buffered since orographic precipitation becomes less efficient at extracting moisture from the flow, due both to thermodynamic and microphysical effects. Salathé et al. (2008) used a mesoscale model to downscale global climate model projections over the Cascade and Olympic mountains and showed that possible changes in the direction of airflow during storms may alter the intensity and distribution of precipitation over the region. Generally, orographic snowfall is very likely to decrease with climate warming as melting levels during storms rise and a larger fraction of precipitation falls as rain. Some loss may be offset by orographic precipitation rate increases, but for mountains like the Cascades and Olympics, where temperatures are not typically far below freezing during storms, this compensation is can be only modest due to the substantial loss of snow accumulation area.

Summary

Orographic precipitation processes strongly shape the climate in and around mountainous regions. Orographic influences can be pronounced on spatial scales ranging from the size of individual hills to the scale of major mountain ranges, and on temporal scales from the duration of a brief snow squall to the long-term climatology. Almost all orographic influences are fundamentally caused by topographically driven ascending and descending atmospheric motions that force condensation and evaporation. However, these basic forcings combine with a wide range of dynamical and microphysical processes to shape the precipitation distribution. Since different physical processes can be important for different storms and for different mountain ranges, orographic precipitation influences may take many forms.

Characterizing and understanding the effects of topography on precipitation remains an active field of research. Current research questions include: How does orographic precipitation change with climate? How do turbulent atmospheric motions affect orographic enhancement? What are the limits on predictability of orographic precipitation? Synthesis of new theories, models, and observational techniques continues to aid us in trying to answer these and other important questions.

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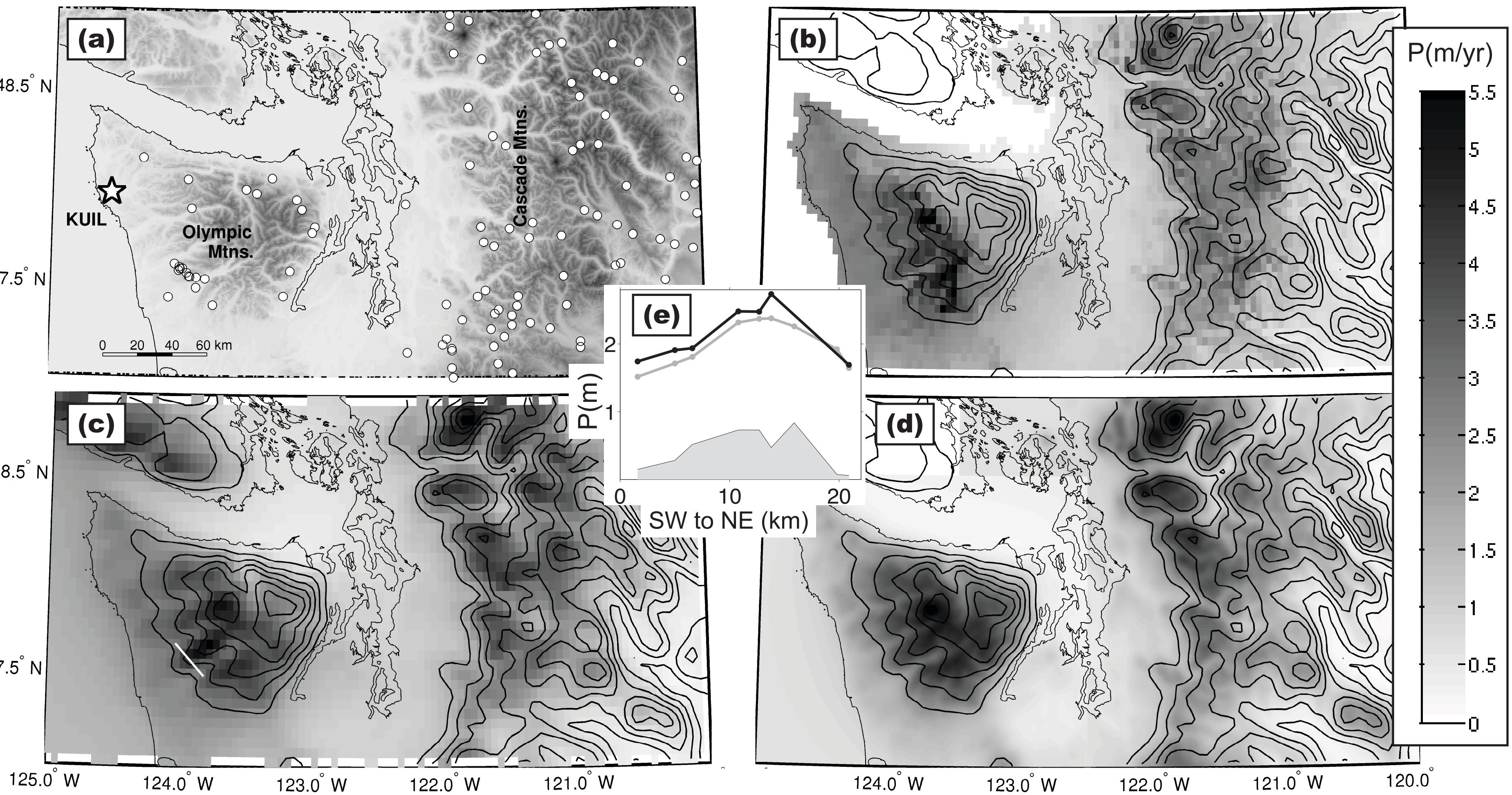
Cross References

Accumulation
Atmospheric processes and snow / ice formation
Cascade Mountains, USA
Global warming and its affect on snow / ice / glaciers
Latent heat of condensation
Latent heat of fusion
Precipitation
Snow Line
Snow Storms
Snow Fall
Solid precipitation
Spatial / temporal variation in snow cover/snow melt

Figure Caption

Figure 1. Topography and precipitation for the Olympic and Cascade Mountains of northwestern Washington, USA. (a) shows elevation in grayscale (black corresponds to 3.5 km) and the location of regularly reporting precipitation gauges located above 150 m elevation (white dots).(b)-(d) shows precipitation from October 2000 – September 2007 in gray shading, and smoothed contours of elevation every 250 m. (b) is the PRISM analysis of gauge observations (Daly et al. 1994; data from PRISM Group, Oregon State University, http://www.prismclimate.org). (c) is from the model of Smith and Barstad (2004) forced with data taken from atmospheric soundings at KUIL (shown in (a)). (d) is from the operational MM5 numerical weather predictions (e.g. Minder et al. 2008; http://www.atmos.washington.edu/mm5rt/). (e) compares the observed (gray) and MM5 (black) precipitation for a gauge transect in the southwestern Olympics (location shown with white line in (c)) for the winter of 2004-2005. The topographic profile (peak elevation of 800 m) is shaded (modified from Minder et al. (2008) and reproduced with permission from Wiley-Blackwell).


(a)
(b)
(c)
(d)
(e)
48.5° N
47.5° N
KUIL
Olympic
Mtns.
Cascade Mtns.
0 20 40 60 km
P(m/yr)
5.5
5
4.5
4
3.5
3
2.5
2
1.5
1
0.5
0
P(m)
SW to NE (km)
125.0° W
124.0° W
123.0° W
122.0° W
121.0° W
120.0°