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OROGRAPHIC PRECIPITATION AND OREGON'S CLIMATE TRANSITION

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1. INTRODUCTION

We examined Oregon's sharp east-west climate transition (Bastach, 1998) using a linear model of orographic precipitation and four data sets: a) interpolated annual rain-gauge data; b) satellite-derived precipitation proxies (vegetation and brightness temperature); c) streamflow data for a small catchment, d) stable isotope analysis of water samples from streams. [Note: Data sets b and c are not discussed in this abbreviated version.] The success of the linear model against these data sets suggests that the main elements in the model (i.e. airflow dynamics, cloud time delays, condensed water advection and lee side evaporation) are behaving reasonably, although the high Oregon terrain may push the linear theory beyond its range of applicability.

A key parameter in the linear model is the cloud delay time (τ), encapsulating the action of orographic cloud processes. Each data set is examined to see if it can constrain the tau values. The state-wide precipitation patterns from raingauge and satellite constrain the taus only within a broad range from about 500 to 5000 seconds. The study of the small Alsea watershed also constrains tau little, as it receives a mixture of upslope and spill-over precipitation. Oxygen-18 isotope ratios in stream water indicate an atmospheric drying ratio of about 43%; requiring an average cloud physics delay time greater than τ = 600 seconds. (See details in Smith et al. 2004)

2. LINEAR MODEL

The primary model we use in this paper is the linear theory (LT) of orographic precipitation proposed by Smith and Barstad (2004). This model solves for the airflow patterns using linear mountain wave theory and solves for the resulting precipitation field using a linear cloud physics representation. In this representation, ascent creates cloud water that advects downstream while converting to hydrometeors on a time scale (τ_c). Hydrometeors also advect downstream while falling to earth on a time scale (τ_f). Descent evaporates cloud water and, if the air becomes subsaturated, hydrometeors evaporate as well. The model is built on ideas developed earlier by Hobbs et al., (1973), Fraser et al., (1973), Smith (1989) and Jiang and Smith (2003).

The linear model is well suited for orographic situations where air passes over complex terrain, so that only a fraction of condensed water precipitates before evaporating on lee slopes. Inputs are given in Table 1. Disadvantages of the LT model include the simplification of vertical structure by vertical integration, linearization of the fluid and cloud dynamics and the lack of a full water budget. Far downstream, all perturbation quantities return to zero, implying a return to a saturated state with a background precipitation rate. Further description of LT is given by Smith (2003), Barstad and Smith (2004) and Smith (2005).

The LT model can be succinctly represented by the equation

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

in which the double Fourier transform of the terrain function

$$\hat{h}(k,l) = (2\pi)^{-2}\iint h(x,y)e^{-i(kx+ly)}dxdy$$

is multiplied by a transfer function to obtain the double Fourier transform of the precipitation field ($\hat{P}(k,l)$). The spatial pattern of precipitation is recovered from an inverse Fourier Transform followed by truncation of negative values.

$$P(x,y) = Max[(\iint \hat{P}(k,l)e^{i(kx+ly)}dkdl + P_\infty),0]$$

C_w is a condensation coefficient depending on the surface humidity and the lapse rate. H_w is the depth of the ambient moist layer. The symbols τ_c and τ_f represent the time scales for conversion and fallout. In the case of τ_c, it described both the rate of conversion of cloud water to hydrometeors and the rate of evaporation of hydrometeors when the air is subsaturated. The background precipitation caused by synoptic scale uplift is P_∞. The intrinsic frequency is $\sigma = Uk + Vl$ and the vertical wave number is the proper root of

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

The three brackets in the denominator of the transfer function describe respectively: airflow dynamics, advection during conversion of cloud droplets to hydrometeors and advection of hydrometeors during gravitational fallout. The first bracket shifts the pattern upstream while the second and third brackets shift the pattern downstream. All three brackets generally decrease the precipitation amount. The Max function sets negative values equal to zero in regions of strong descent, capturing the effect of descent and drying. The transfer function is sensitive to scale. Vertical motions from small horizontal scales (D<1km) do not penetrate substantially into the moist layer and have little influence on precipitation. Intermediate scales (D~5km) generate lee wave clouds but little precipitation. Longer scales (D>15km) are increasingly efficient at producing precipitation. The LT model reduces to the simple upslope model (1) when $C_w = 1$, the vertical motion penetrates the moist layer ($mH_w << 1$) and the cloud conversion and fallout are instantaneous ($\tau_c = \tau_f = 0$).

To illustrate the application of the LT to Oregon terrain, we first consider a smooth 1-dimensional idealization of the coastal and Cascade ranges (Fig 1). Referring to the actual terrain in Figures 2, we represented the two ranges with 750m and 1500m Gaussian ridges, 200 km apart. Each ridge has a width scale of 30 km. The coordinate system is centered on the crest of the coastal range. East of the Cascades, the terrain forms a slowly descending plateau. As a reference run, we select $\rho = 1.2kg/m3$, $q_v = 0.0044$, $U = 15m/s$, $N_m = 0.003s^{-1}$, $H_w = 3000m$, $f = 10^{-4}s^{-1}$ $\tau_c = \tau_f = 2400s$. The resulting precipitation pattern from is shown in Fig 1, prior to adding a background value (P_∞) or truncating the negative values. Maximum values of 1.4 and 3.4 mm/hr nearly coincide with the highest terrain. Unless the background precipitation exceeded 1 mm/hr, dry regions would be found downstream of the coastal range and over most of the plateau east of the Cascades. The dry plateau is the evaporative effect of dynamic subsidence. Several modifications to the reference run are described below.

When the Coriolis and non-hydrostatic terms are neglected in m(k), the changes are too small to perceive. The orographic scales in our "idealized Oregon" are too large for vertical acceleration effects and too small for Coriolis effects to be important. When the coastal range is removed, the predicted precipitation over the Cascades is nearly unchanged (Fig 1). This is so because the dynamic effect of the coastal range decays on the scale of the ridge (30km) and the cloud effect decays on the advection scale $U\tau = 15m/s * 2400s = 36km$ while the ridges are 200 km apart.

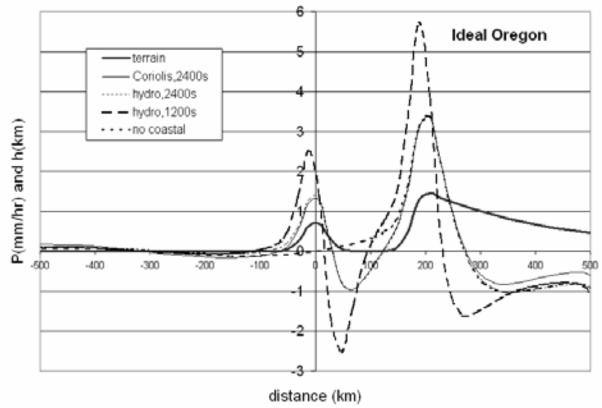

Figure 1. Precipitation rate (mm/hr) according to Linear Theory (3, 5) for idealized Oregon terrain. The terrain height is given in kilometers. Four runs are shown: full dynamics with $\tau = 2400s$; hydrostatic non-rotating dynamics with $\tau = 2400s$, hydrostatic non-rotating dynamics with $\tau = 1200s$, and hydrostatic non-rotating dynamics with $\tau = 2400s$ and no coastal range

When the cloud delay times are reduced from 2400s to 1200s, the change is much more significant (Fig 1). The two precipitation maxima nearly double in magnitude and they move upstream of the ridge crest slightly. The choice of the shorter $\tau = 1200s$ amplifies the precipitation so much for this smooth high terrain that it violates the assumption of linear theory. The total precipitation in the 2400s and 1200s runs are 140 and 240kg/ms respectively, while the estimated influx of vapor is

$$F = \rho q_v U H_m = (1.2)(0.0044)(15)(3000) = 216kg/ms$$

. Thus, in the 1200s run, the total precipitation exceeds the incoming flux!

3. "CLIMATE RUNS" WITH REAL TERRAIN

To simulate the climate of Oregon, we ran the model with three SW wind directions and speeds and weighted the results with the observed frequency distribution. A value of $\tau = 1200s$ is used. The model run used the 1km terrain shown in Fig 2a.

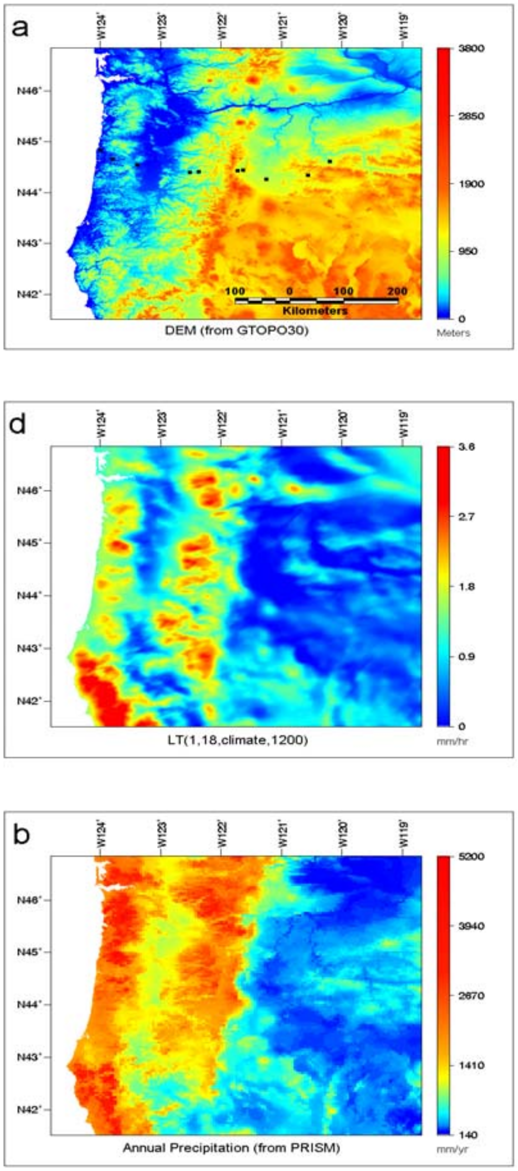

Fig 2. *Three maps of Oregon. a) 1-km terrain from GTopo30, b) Linear model prediction of precipitation rate (mm/hr) for a typical event, c) Annual rainfall (mm) interpolated using PRISM.*

The result of the climate run (Fig 2b) agrees reasonably well with the interpolated annual precipitation map by Daly et al., (2002, Fig 2c). Other trials of the model suggests that as long as the tau values are kept in the range 500 to 5000 seconds, the main features of the field are captured. Taus less than 500s give dry mountain crests. Taus longer than 5000s give wet conditions east of the Cascades. Focusing only on the sharp gradient on the lee slopes of the Cascades suggest that tau values in the range of 1800 to 2400s would give the best fit.

4. ISOTOPE DETERMINATION OF DRYING RATIO

To improve our estimation of cloud delay time, we sampled stream water along an east-west cross section in Oregon (see Fig 2a). Water samples were analysed in a mass spectrometer for their deuterium and oxygen-18 concentrations. A strong isotope gradient was found (Fig 3) indicating progressive fractionation as moist air masses pass over the coastal and Cascade ranges.

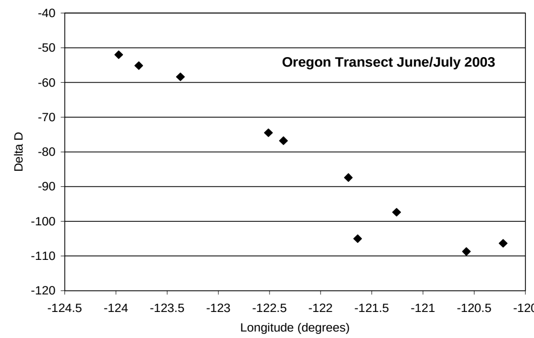

Figure 3. Normalized deuterium concentration in streamwater as a function of longitude across the Oregon mountains.

The drying ratio

DR = Total precipitation/ Water vapor influx

is helpful for evaluating models and setting adjustable coefficients. In this section we use isotopic data to estimate DR. Assuming that the fractionation coefficient is constant, we obtain a relationship between isotopic concentration in precipitation and the fraction of vapor remaining in the atmosphere is

$$\Theta = (R_P / R_{P0})^{1/(\alpha-1)}$$

where the R values are the isotope ratios upstream and downstream of the range and theta is the percent vapor remaining (Friedman, 1955, 1964; Dansgaard, 1964). From the definition of the drying ratio

$$DR = 1 - \Theta = 1 - (R_P / R_{P0})^{1/(\alpha-1)}$$

Using values from Figure 3 and using a fractionation factor of $\alpha_D = 1.106$ (for T=0C) gives

$$DR = 1 - (0.894 / 0.948)^{9.43} = 1 - 0.58 = 0.42$$

5. ESTIMATING THE CLOUD DELAY TIME

Using the isotope-derived drying ratio, we can estimate the cloud delay time. The greater the value of tau, the less precipitation and drying will occur. To establish the relationship between tau and DR, several runs of the linear model are carried out with an ensemble of realistic environmental conditions, and a variety of tau values. The resulting relationship is shown in Fig 4. Using the isotope-derived DR, a value of $\tau = 600s$ can be determined.

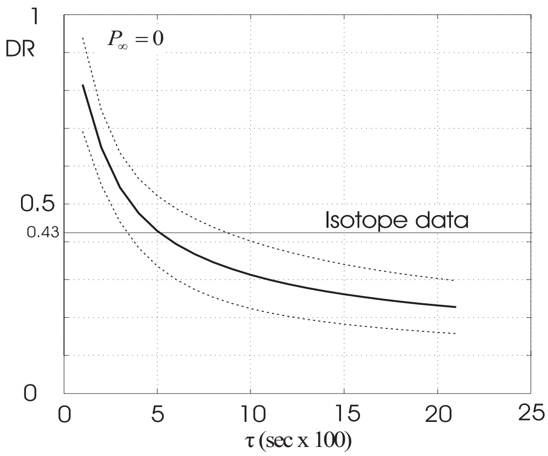

Figure 4. The relationship between cloud delay time (τ) and drying ratio(DR) for Oregon terrain, according to the linear model. The isotope-derived drying ratio is also shown.

The value of background precipitation must also be accounted for. As orographic precipitation events in Oregon usually occur within precipitating cyclonic systems, the background or "non-orographic" component of the precipitation is significant The presence of background precipitation will lift the DR curve in Fig. 4 everywhere, so for a given value DR the inferred τ is larger. Using the isotope-derived value of DR = 0.43, our estimate of τ is increased from 600s to 1200s or even greater. Values ranging from 800 to 1500 seconds appear reasonable in this context, and agree with earlier estimates (Jiang and Smith, 2003, Smith 2003).

6. CONCLUSIONS

Using a linear theory framework, we attempted to determine the cloud physics delay time (τ) from macroscopic measurements. Our estimates varied from 600s to more than 2000s. The strong sensitivity to τ indicates that terrain scales of 20 to 30km cause repeated up and downdrafts with condensed water advection and evaporation.

7. ACKNOWLEDGEMENTS

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