DUST CYCLES AND STORMS IN A MARS GCM.

C. E. Newman, P. L. Read, S. R. Lewis, *Atmospheric, Oceanic and Planetary Physics, Department of Physics, University of Oxford, England (newmanc@atm.ox.ac.uk)*, **F. Forget,** *Laboratoire de Météorologie Dynamique du Centre National de la Recherche Scientifique, Université Paris 6, 4 place Jussieu, 75252 Paris Cedex 05, France.***.**

Introduction

A number of different dust lifting parameterizations have been used to model the injection of dust from the Martian surface into the atmosphere, and the form of the resulting dust cycles and dust storms produced are found to be highly dependent on the precise form of the parameterization used, provided that it includes some threshold dependence, and particularly where radiatively active dust transport is employed. This talk will review the most interesting results from previous work. We have recently altered a key factor which particularly affects the dust lifting due to near-surface wind stress, however, so we will also present results using the new dust lifting formulation, and make some comparisons.

In the boundary layer, the drag velocity u_* may be given by (see e.g. [1])

$$u_{drag} = \frac{ku(z)}{\ln\left(\frac{z}{z_0}\right)}. \tag{1}$$

where k=von Karman's constant=0.4, z=height above the surface, z_0=roughness height=0.01m and $u(z)$ is the magnitude of the near-surface wind velocity (at height z). In the near-surface wind stress lifting parameterization, surface particles are assumed to be lifted by near-surface wind stress whenever $u_* = \sqrt{(stress/density)}$ exceeds a threshold value u_*^t, which may be calculated using a set of semi-empirical equations [2]. For expected values of interparticle cohesion, the threshold is predicted to be lowest, thus allowing the greatest movement, for large $\sim$100μm particles. We therefore calculate u_*^t for these particles, and assume that the smaller dust particles are lifted via saltation. We then assume that the vertical dust flux F is proportional to the horizontal flux of saltating particles [3], giving

$$F = \alpha_N \times 2.61 \frac{\rho}{g} (u_*)^3 \left(1 - \frac{u_*^t}{u_*}\right)\left(1 + \frac{u_*^t}{u_*}\right)^2. \tag{2}$$

A simpler (though less accurate) alternative method is to find u_*^t using a constant threshold stress, and then use equation 2 as before. This has the advantage of having only one free parameter (the constant threshold stress used) which enables parameter space to be explored more rapidly.

Lifting by the convective vortices known as dust devils is accounted for in two ways, both of which involve modelling the vortices as convective heat engines [4]. In the first method, dust lifting is set proportional to the 'dust devil activity', which is related mostly to the thickness of the convective boundary layer and the surface sensible heat flux, and no thresholds are used. This results in a relatively smooth variation of dust devil lifting over the planet, and represents a background lifting process. Computationally, it may also prevent the model failing due to unrealistically strong gradients in temperature, etc. – these can be produced by overly localised dust distributions, which sometimes arise when near-surface wind stress lifting is confined to one or two gridpoints. The second method uses the heat engine model to predict the tangential velocity V around a vortex, again depending on the thickness of the convective boundary layer and the surface sensible heat flux, but now dust is only lifted if this exceeds a threshold value, V^t, calculated using a semi-empirical equation derived from laboratory dust devil experiments [2].

Results using the previous formulation

In the previous formulation, u_* values were calculated from values of $u(z)$ found *prior* to vertical diffusion being carried out on the model wind field. The main results using this formulation were as follows:

If dust is lifted only by near-surface wind stress, or with the first dust devil lifting parameterization used in addition to produce reasonable background dust loading, certain 'typical' Martian dust storms are produced in some experiments. The most interesting behaviour is found when a high threshold u_*^t is used (this is the case if the interparticle cohesion between dust particles is assumed to be large) as this allows for dust lifting to occur suddenly and in localised regions when these thresholds are exceeded. Regional storms are often simulated to begin in the Hellas region, where the rapid growth produced is due largely to the positive feedbacks on surface wind speeds at the edge of the growing dust cloud, where there are large temperature contrasts. Figures 1 and 2 show respectively visible dust opacities and zonally averaged mixing ratios for one such simulation.

Figure 3 shows visible dust opacities for the growth phase of another type of storm, beginning near the Chryse region, in which the dust cloud grows as it moves southwards in a strong western boundary current to the east of Tharsis. Again there is a positive feedback between the winds at the cloud's edge and dust lifting, although the latter finishes once the storm reaches the southern hemisphere, and it decays over the south pole.

These two regional storms are similar to those ob-

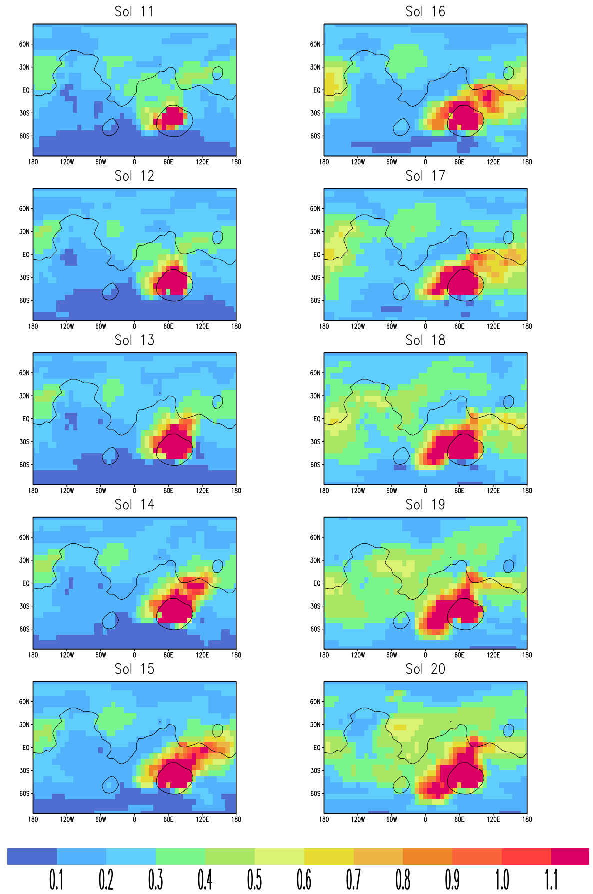

Figure 1: Visible dust opacities during a simulated storm beginning in the Hellas region at Ls∼185 degrees. Also shown is the 500Pa surface pressure contour, to indicate roughly the position of topography.

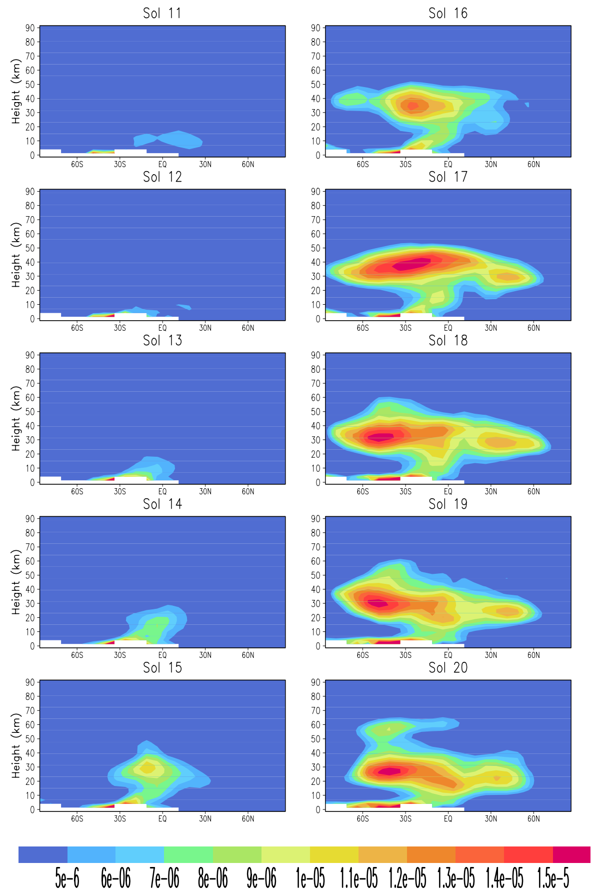

Figure 2: Zonally averaged dust mixing ratios corresponding to the dust opacities shown in figure 1.

served on Mars [5], [6], yet other types of storms, those beginning in other locations, or those which go on to become global in extent, do not appear to be easily produced using the near-surface wind stress lifting parameterization alone, at least with this formulation. The Chryse storm type also represents the greatest source of interannual variability produced in these simulations, as it does not occur in every year, yet on real Mars there is far more variability observed.

If dust is lifted by a combination of near-surface wind stress and the second dust devil lifting parameterization, other interesting behaviour is seen. Near-surface wind stress lifting tends to produce a positive feedback between increased dustiness and further lifting (due to local wind speed increases, or increases in the strength of the global circulation), but dust devil lifting tends to have a negative feedback (as increased dustiness reduces the surface heating, hence reduces the drive to the dust devils). The interplay between the two processes therefore produces increased variability on timescales of a few sols, and increases the potential for interannual variability. Dust devil lifting also peaks in different regions to near-surface wind stress lifting in many cases, so increases the number of main storm onset regions. The peak in dust lifting also shifts from region to region, as is often observed on Mars, and the areas involved are those commonly observed to be dusty during large storms [7].

With near-surface wind stress lifting only, storms were produced which grew realistically rapidly and became global, but in general they then went on to attain unrealistically high opacities without decaying. The advantage of combining this with the second (threshold-sensitive) form of dust devil lifting is that the negative feedbacks involved for the dust devil mechanism moderate this growth, and thus global storms which decay spontaneously are easily produced. Figure 4 shows zonally averaged visible dust opacities averaged over lati-

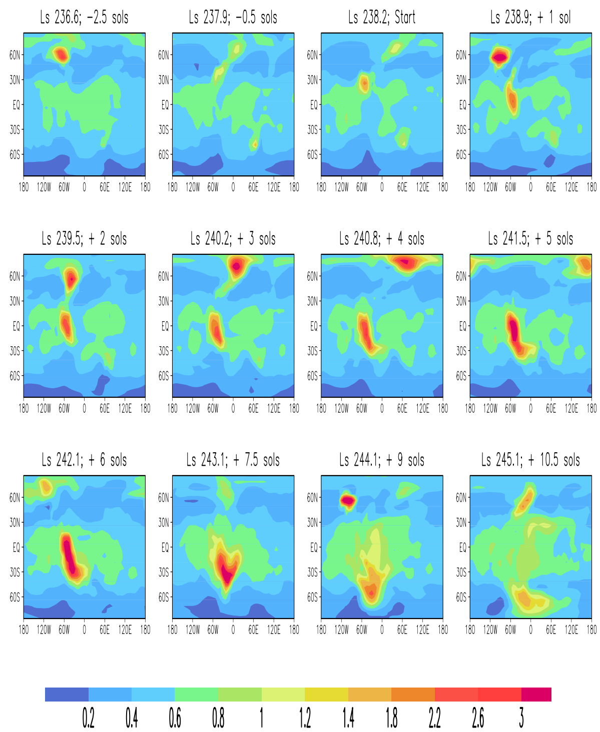

Figure 3: Visible dust opacities during a simulated storm beginning in the northern Chryse region at Ls∼240 degrees.

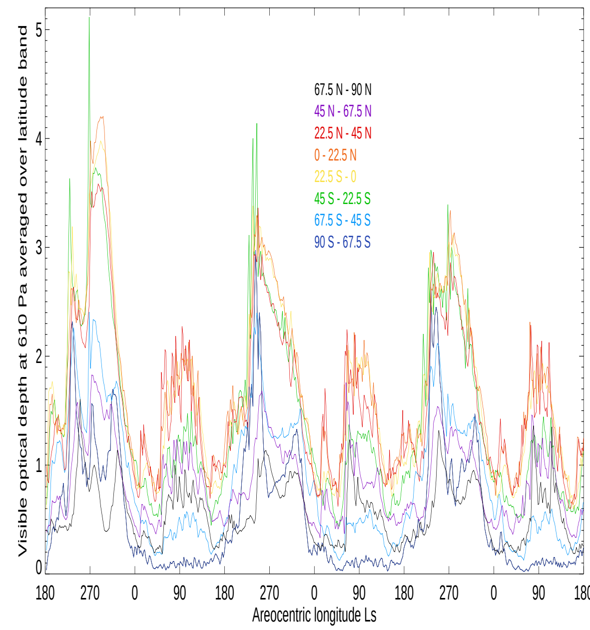

Figure 4: Zonally averaged visible dust opacities in latitude bands for three years of a simulation with threshold-sensitive near-surface wind stress and dust devil lifting. Two global storms are produced during the first year which compare well with the 1977a and b storms observed by Viking.

tude bands for three years of the experiment described above. This simulation is still deficient, however, in that it does not produce any ‘clear’ years, only years with one or more global storms, and northern spring and summer opacities in particular are far higher than observed. The latter is due mostly to dust devil lifting, and seems to suggest that dust devils cannot be responsible for as high a proportion of the dust lifting as this simulation allows, unless some other mechanisms (e.g. scavenging by ice clouds) removes much of the dust lifted at this time.

Results using the newer formulation

In the newer formulation, u_* values are calculated from values of $u(z)$ found *after* vertical diffusion is carried out on the model wind field. Initial results indicate that using u_* after vertical diffusion has been applied reduces wind stresses considerably, thus requiring lower thresholds than before to produce comparable opacities. Using post-vertical diffusion u_* also, however, removes suspiciously high northern hemisphere lifting peaks (particularly in regions of large topographic gradients) which were primarily responsible for global storms growing uncontrollably in the case of near-surface wind stress lifting experiments. Also, the Hellas and Chryse dust storms shown above are still found to occur, but additional regions (such as Noachis, where a regional storm was observed in 1997, see [8]) are found to be peak dust sources at different times of year. This work is still in the early stages, but new results will be presented at the workshop.

Discussion

The previous formulation produced some reasonably realistic dust storms and dust opacity cycles, but these were still deficient in some respects, and problems were encountered with excessive near-surface dust lifting via positive feedbacks, particularly in the northern hemisphere and at higher model resolutions. The newer formulation may produce a greater variety of stable dust storms via the near-surface wind stress parameterization (thought to be the major contributor to dust lifting on Mars), including global storms which do not grow uncontrollably.

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