10.1 MODEL SENSITIVITY TO SPACE - AND IN SITU - BASED SUB-GRID FLUX PARAMETERIZATIONS

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1. INTRODUCTION

Large-scale models do not adequately resolve the spatial variability that contributes to surface fluxes in the atmosphere and thus can underestimate surface fluxes considerably, especially in areas of low winds where the grid-scale directional variability may be largely averaged out. In oceanic regions where data that can be assimilated are sparse or non-existent, and the availability of latent energy abundant, considerable underestimation of surface fluxes may result. One way of improving flux estimates in large scale models is to parameterize the contribution to the surface flux by the unresolved directional variability in the near surface wind field. Based on a combination of Scatterometer and buoy observations collocated at different locations spanning a range of climatic regimes, Levy and Vickers, 1999 and Levy, 2000 have formulated such a parameterization. In a series of sensitivity tests, we examine the impact of this subgrid parameterization on the NCEP operational model at different resolutions. The parameterization is formulated as a resolution dependent velocity scale term that is adaptable to varying model resolution.

2. MODELED FLUX, COEFFICIENT AND VELOCITY-SCALE FORMULATIONS

The most common way of estimating surface fluxes is by using a bulk formulation, developed based on correlations with direct measurements:

$$F_\phi = \rho \cdot C_\phi \cdot V(\Phi_{sfc} - \Phi) \qquad (1)$$

where Φ is the quantity whose surface flux is estimated (i.e., wind, temperature, and humidity herein) at the interface and a reference level, C is the transfer coefficient, and V is the magnitude of the wind vector relative to the surface.

The bulk formula for momentum (u component) then takes the form:

$$\overline{w'u'} = -C_d(z/L, z/z_o)V\overline{u} \qquad (2)$$

where C_d is the drag coefficient, which is a function of stability, the aerodynamic roughness length and over the ocean, also the age of the wind driven waves, V is the wind speed of the vector-averaged (over a grid-box) flow, $\overline{u}$ is the average of the u-component. A similar equation can be derived for the v-component of the stress.

Physically over water, the Charnock relationship is used to relate the surface stress (LHS of 2) to the wind speed and other physical (atmospheric and oceanic) parameters:

$$Z_o = Z_* \frac{u_*^2}{g} \qquad (3)$$

and the roughness parameter (Zo) is defined through the logarithmic wind profile extending to the surface from a (Z=) 10 meters reference height:

$$u(z) = \frac{u_*}{\kappa} \ln \frac{z}{z_o} \qquad (4)$$

u* is the friction velocity, g acceleration of gravity, κ the von Karman constant, and z* the dimensionless Charnock parameter.

The surface stress or sea drag is defined as

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$$\frac{\tau}{\rho} = u_*^2 \quad (5)$$

Details in constants vary between parameterizations, but most large scale models, including the GFS model tested here use a bulk relationship (1), (2) above and a Charnock relationship to determine the drag coefficient (C_d) over the ocean.

Unresolved variability present in the model grid box spatial averages can result in systematic errors in the flux estimation. To account for such errors, subgrid effects can be incorporated as a modification to the exchange coefficient (e.g., C_d in equation 2) used. Using the momentum (equation 2) as an example

$$\overline{w'u'} = -C_{deff} V \bar{u} \quad (6)$$

one can define an effective drag coefficient, C_{deff}, and relate it to information on subgrid scale variability. For example:

$$C_{deff} = C_d(z/z_o, z/L) g\left(\frac{\Delta x}{h}, \frac{\vartheta_{Tsfc}}{[T_{sfc}] - [T(z)]}\right)$$

where C_d is the prediction of the drag coefficient using Monin-Obukhov similarity theory and the Obukhov length based on grid-averaged heat and momentum fluxes.

The effective coefficient C_{deff} above can include all subgrid effects that are known and can be parameterized. Some of the effects in the effective coefficient C_{deff}, are already included in the existing model Boundary Layer parameterization (e.g., stratification effects). Other effects (e.g., thermal subgrid variability) are more complicated and are yet to be parameterized. Of all the subgrid effects, those that are due to vector averaging (subgrid directional variability) are most consistent as they always act to enhance fluxes. Indeed although unresolved directional variability is only one source of inaccuracy in flux estimates, it was shown to have the largest effect on flux underestimation [*Esbensen and McPhaden*, 1996; *Vickers and Esbensen*, 1998], and it is one source for which inaccuracy can be estimated and uncertainty can be reduced. Such subgrid variability had been estimated from different data sets by several authors (e.g., Sun et al., 1996; Mahrt and Sun, 1995; *Vickers and Esbensen*, 1998; Levy and Vickers, 1999; Levy 2000). In this paper we study the impact of a parameterization constructed from such estimates on model forecasts.

Since the subgrid parameterization we are testing in this study is entirely due to directional variability, accounting for it in V in equations (1) and (2) above as a velocity scale as in Levy and Vickers, 1999 is entirely analogous to including it as a correction term in an effective coefficient C_{deff}, but is easier to implement as it does not require reprogramming the model C_d parameterization.

One can define a velocity scale, V_{sg}, by writing the magnitude of the mean wind vector, U in the form:

$$U = (V^2 + V_{sg}^2)^{1/2} \quad (7)$$

where V_{sg}, the velocity scale is a correction due to unresolved spatial (sub-grid), or convective variability (Godfrey and Beljaars, 1991; Beljaars, 1995). The velocity scale formulation of Levy and Vickers (1999) and Levy (2000) is a function of scale and is based of a best fit (in a least square sense) of estimates to scatterometer and buoy data at different scales and different locations. It takes the form:

$$V_{sg} = \alpha(\Delta X/D-1)^{\beta} \quad (8)$$

where ΔX is the grid-scale and D is the smallest scale resolvable in the data.

An example of the sub-grid velocity scale (V_{sg}) from QuikScat data collocated with NDBC buoy in the Gulf of Mexico is shown in Figure 1 (from Levy, 2000).

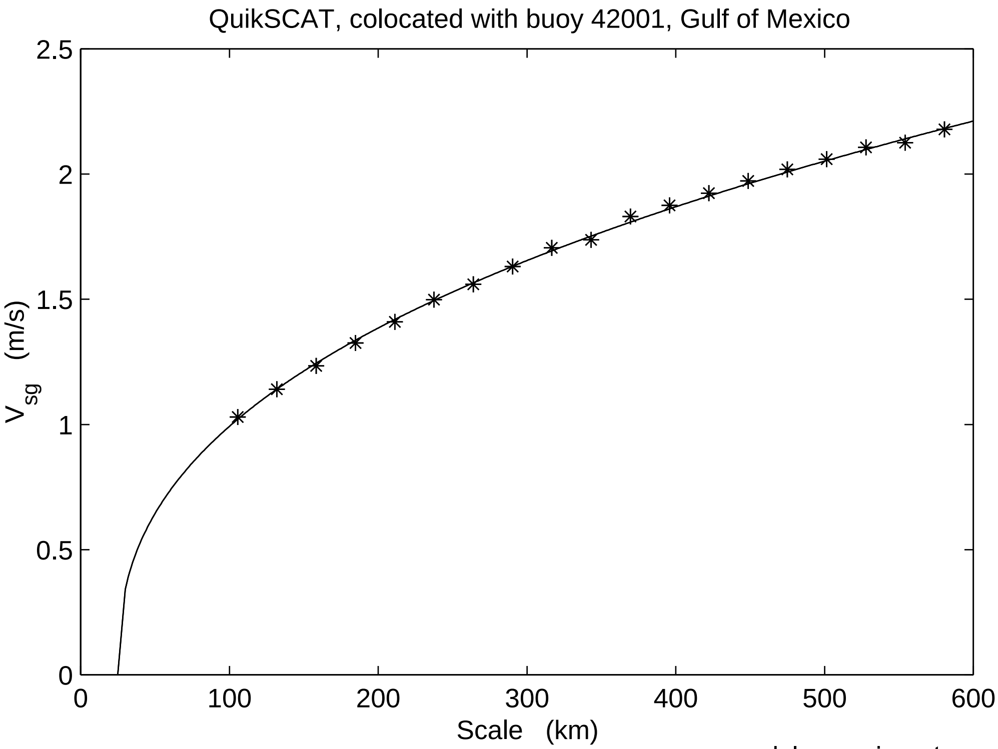


Figure 1: Estimates of the sub-grid velocity scale (V_{sg}, m/s) from QuikScat data collocated with NDBC buoy 42001 versus grid scale and the associated model fit (solid line) using equation (8) (after Levy, 2000).

3. MODEL

The data used in this study are from experiments using the NCEP global forecast system (GFS) model. The GFS currently is a global spectral model using a 64 sigma vertical coordinate system for weather service operations incorporating a complete set of physical parameterizations. The model is used for data assimilation from which operational initial conditions are determined, and aviation, medium range and climate forecasts are made. The GFS is one of the most essential components of weather service forecasts facilities. Our experiments were configured to run at triangular truncation of T254 with a physical Gaussian grid of 768x384 which is further post processed to pressure levels and a physical grid of 1x1 degree (Kanamitsu et al 1991, Sela, 1982). Recent changes to the model can be found at the NWS/NCEP web site.

4. EXPERIMENTAL DESIGN

To show the impact of the flux enhancement from sub-grid scale wind variability, two model experiments were run and compared with control runs. Suitable values for α=.30, β=.43 in equation 8 were taken from Table 2 in Levy and Vickers, (1999). The resolution of the GFS and the smallest resolvable scale in the data require to set the term in the parenthesis in equation 8 to 9.0. The sub grid velocity scale correction becomes a constant and is simply added to the wind magnitude when the boundary layer bulk formulations are calculated. We begin with initial conditions from 8 February 2004 and make 5-day forecasts for two experiments:

1) The wind enhancement is applied only over the oceans (ASF1); and
2) The wind enhancement is applied over all model grid points (ASF2).

The enhancement to the wind from this parameterization is on the order of less than 1 m/s (.77m/s) although they are applied at many grid points and all time steps. To see the average impact would require many model runs for sufficient statistics. However, as a first step to analyzing the impact a small number of experiments are run and special synoptic areas are examined where the surface fluxes show response to the enhancement. These locations are used as case studies to see if the parameterization makes sense synoptically before a large amount of resources is committed. The experiments with the wind enhancement are compared to the control runs and the differences are displayed and discussed in the next section.

5. RESULTS

As already mentioned, the effect of the subgrid variability in the wind on the absolute magnitude of the wind is a function of the grid scale and for the current runs is < 1 m/s. Indeed, when the global differences in the wind at the surface (not shown) are inspected, it is evident that the differences in wind speeds are small. It follows that in a relative sense, the sub-grid effects would be maximized in low wind regimes. In this weak wind regime the effects can be important as noted in some previous studies. For example, in light wind convective conditions, Godfrey and Beljaars (1991) and Beljaars (1995) argue that the wind used as input for flux calculation should be enhanced at small scales to account for wind direction variability due to free convection. The current (control) model parameterization does not allow the wind to vanish and therefore already partially compensates for the directional variability at the lowest limit.

The changes to the prediction of the momentum flux (equation 2) can be significant especially in coupled models. However the assessment of the synoptic impact of such changes in the stress field is somewhat illusive in atmospheric only model runs. We therefore focus on the dynamic changes and impact through analysis of the height field, and the thermodynamic changes through the analysis of the heat fluxes and precipitation predictions. We present changes at the surface but note that changes to the height field extend to higher levels (500 mb and 200 mb).

Figure 2 shows the differences in the 1000 mb geopotential height between the run with the sub-grid parameterization (ASF1) and the control run at 120 hrs. (5 days) simulation time. Several areas of significant dynamical impact are apparent. The changes in the Northeast Pacific (Off the Canadian West Coast and over Alaska) as well as over the Atlantic display a dipole structure indicative of an overall amplification of the mid and high latitude

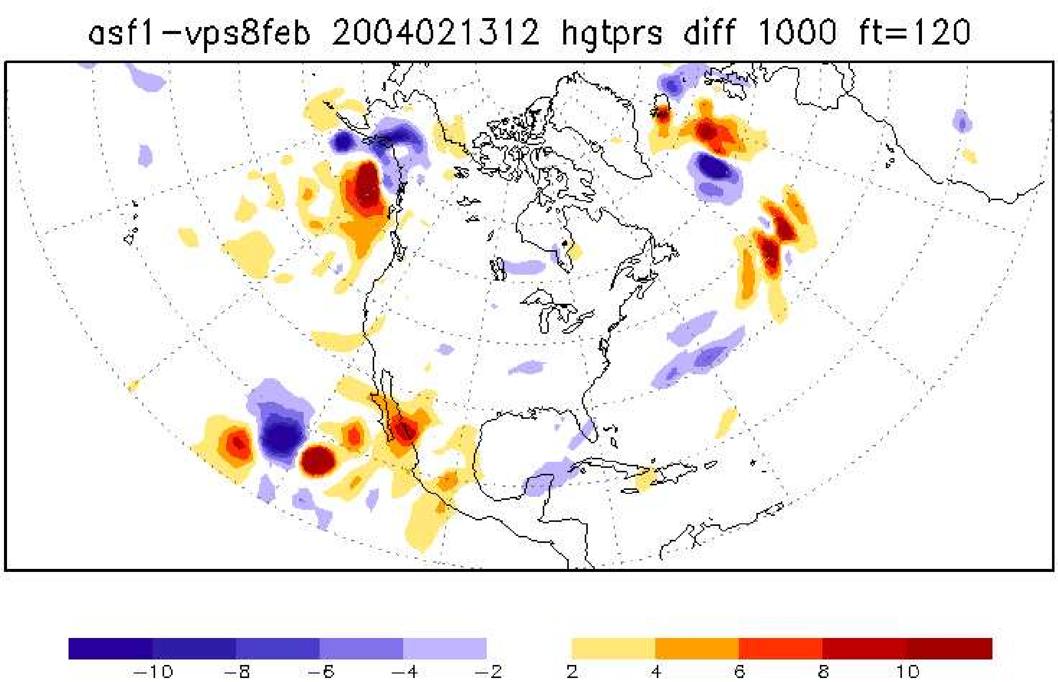


Figure 2: Differences in geopotential height field (m) at 5 days forecast time between ASF1 and control forecasts (Feb. 13, 2004, 12 GMT)

synoptic waves that appear in the control runs when the sub-grid variability parameterization is added (i.e., ASF1 and ASF2 runs). Of these waves, the system over Alaska and the Northeast Pacific is shallower and the impact, while strong at the 500 mb level, diminishes considerably at the 200 mb level. The southern flank of the Atlantic disturbance is associated with a frontal system.

In the tropics, the dynamical impact of the parameterization is associated with a deepening of a system in the Caribbean and Yucatan and significant dynamical changes in convective systems over the tropical Eastern Pacific and Central America.

Figures 3 and 4 show the impact of the sub-grid variability parameterization (ASF1) on the sensible and latent heat fluxes at the surface, respectively. The impact of applying the parameterization over the oceans on the sensible heat flux (Figure 3) is limited to just a few locations. This is expected, since over the oceans the sensible component of the heat flux is small. Nonetheless, a dipole structure evident along the Atlantic frontal system suggests that adding the subgrid parameterization impacted the propagation and structure of this front. Significant changes in the sensible heat flux are evident also at low latitudes areas over land (West Africa and the Mexican Pacific coast).

The changes in the latent heat fluxes (Figure 4) are much more pronounced but are also limited to just several areas. The changes along the Atlantic frontal system are generally in agreement with those seen in the geopotential height field and the sensible heat flux. Significant changes are also evident at low latitudes and equatorial regions in the central and eastern Pacific and in the Caribbean. Those changes indicate a significant impact of the subgrid parameterization on the simulated convective activity. Latent heat flux differences and enhancements in equatorial regions range from 20% in the Caribbean and equatorial Atlantic to 20%-40% in the central Pacific and 60%-80% in the equatorial eastern Pacific.

Figure 3: As in Figure 2 except for Sensible heat flux (Wm^{-2}) at the surface.

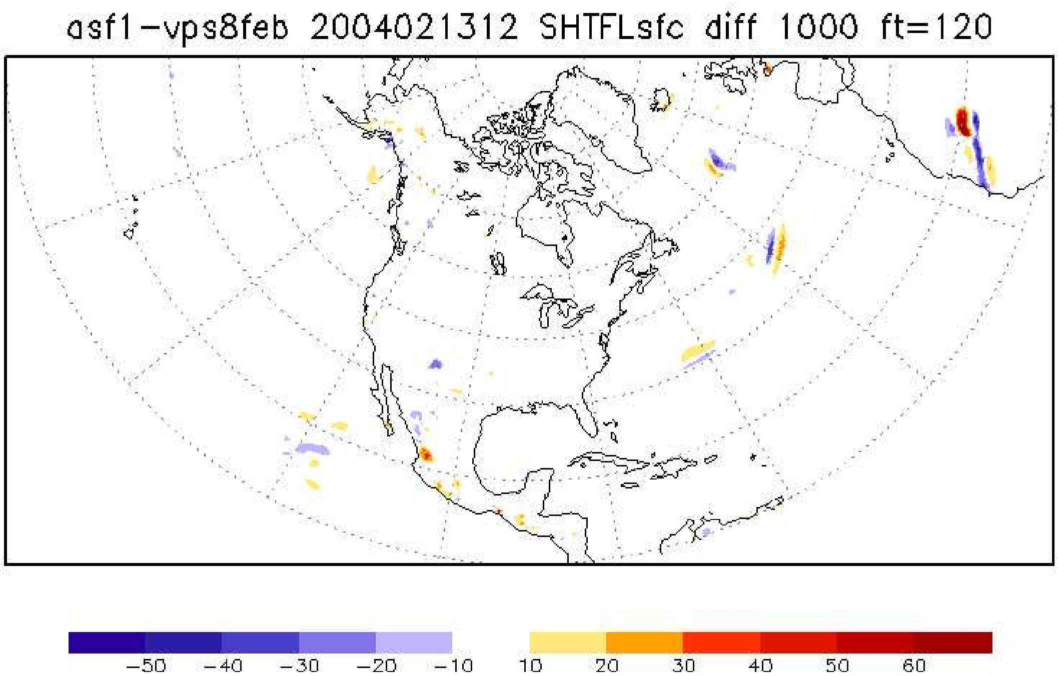
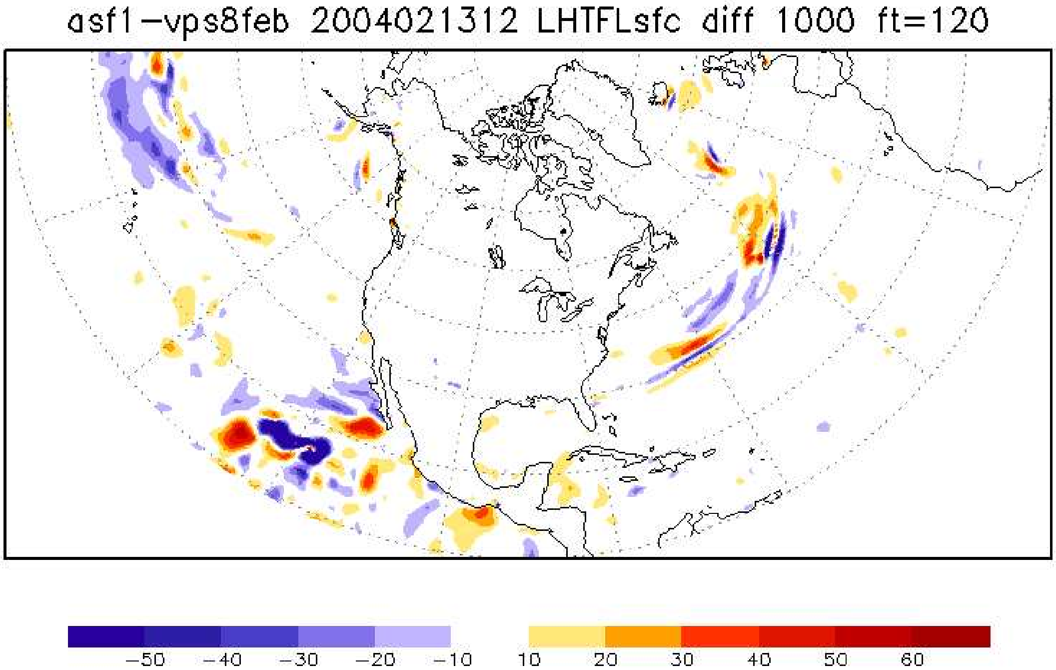


Figure 4: As in Figure 2 except for latent heat flux (Wm^{-2}) at the surface.



The changes in the latent heat flux indicate significant changes in the availability of convective energy and moisture and are therefore also reflected as changes in the precipitation that are shown in figures 5 and 6.

Figure 5: As in Figure 2 except for convective precipitation (inches over 6 hrs.) at the surface.

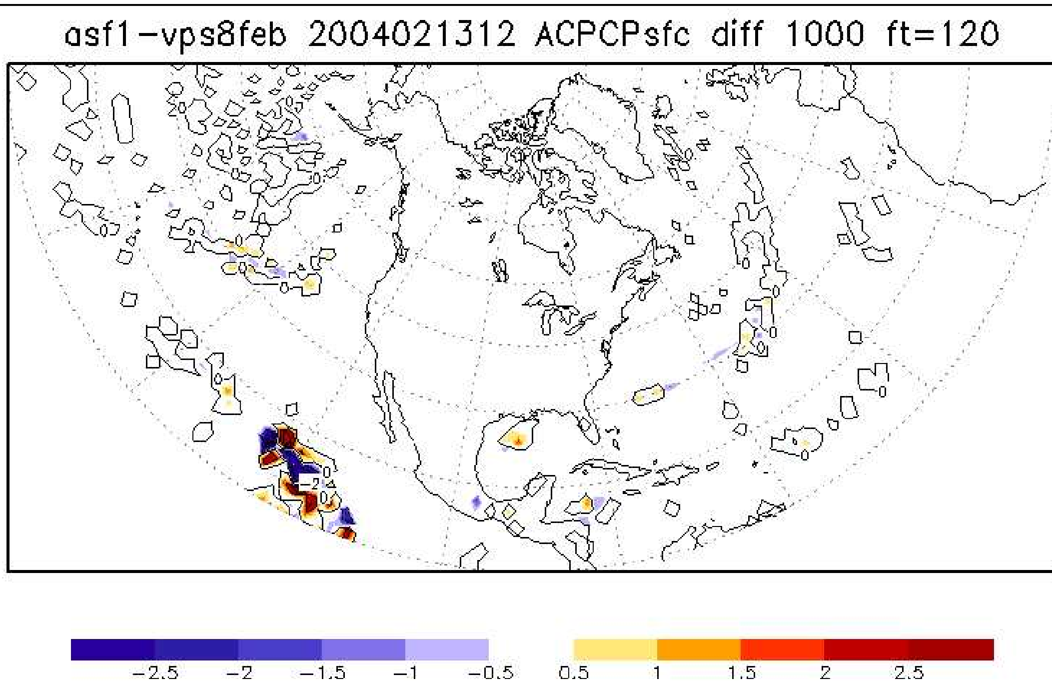
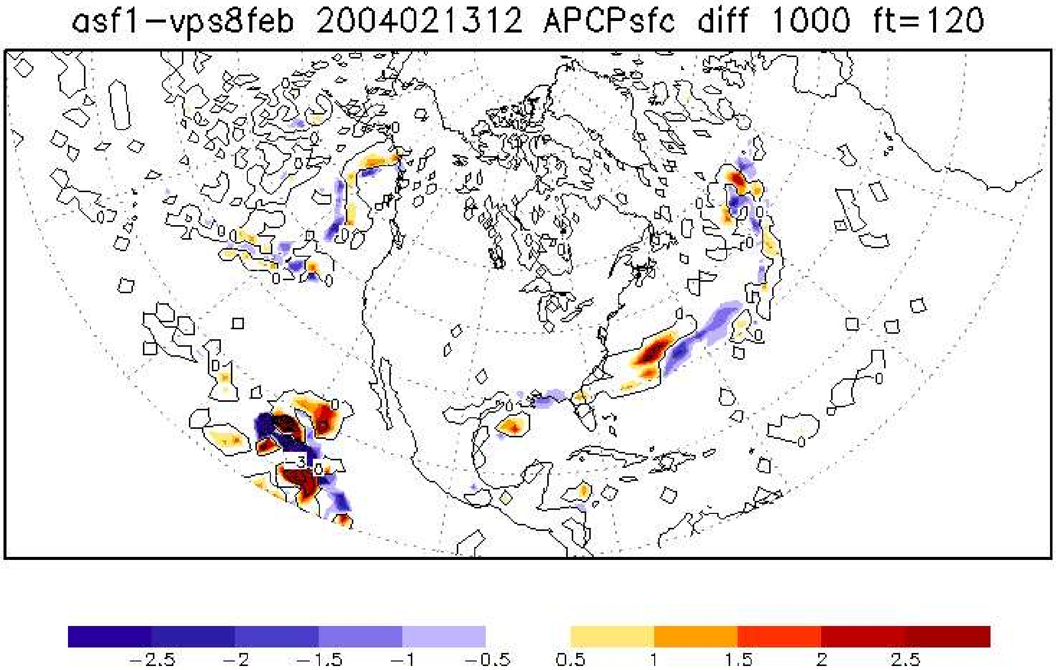


Figure 6: As in Figure 2 except for total precipitation (inches over 6 hrs.) at the surface.



The impact of the parameterization on the convective precipitation (figure 5) in the tropics is significant in the eastern tropical Pacific. Additionally, convective precipitation is enhanced (by 2 inches) in the Caribbean and the Gulf of Mexico) and somewhat in midlatitudes convective cells in the central Pacific. While the changes in precipitation in the tropics are entirely due to convection, changes in total precipitation occur also at higher latitudes in the synoptic systems in the north Atlantic and north Pacific (around the front and storm in the Atlantic and in the Gulf of Alaska). These changes reflect changes in stratiform precipitation.

6. SUMMARY

Numerical experiments using the NCEP global forecast system (GFS) model are conducted. A series of model forecasts are run with and without a subgrid parameterization and the impact on the model dynamics and thermodynamics is tested by comparing the resulting fields of geopotential height, heat fluxes, and convective and total precipitation at five days forecast time for select synoptic cases. The major impact of the parameterization is to amplify the synoptic waves at mid and high latitudes, enhance latent heat fluxes and tropical convective precipitation, and redistribute stratiform and total precipitation in middle and higher latitude systems.

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